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**1 A New Paradigm for Large Earthquakes in Stable**  
**2 Continental Plate Interiors**

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3 Large earthquakes within stable continen-  
4 tal regions (SCR) show that significant amounts  
5 of elastic strain can be released on geological  
6 structures far from plate boundary faults, where  
7 the vast majority of the Earth's seismic ac-  
8 tivity takes place. SCR earthquakes show spa-  
9 tial and temporal patterns that differ from those  
10 at plate boundaries and occur in regions where  
11 tectonic loading rates are negligible. However,  
12 in the absence of a more appropriate model,  
13 they are traditionally viewed as analogous to  
14 their plate boundary counterparts, occurring  
15 when the accrual of tectonic stress localized  
16 at long-lived active faults reaches failure thresh-  
17 old. Here we argue that SCR earthquakes are  
18 better explained by transient perturbations  
19 of local stress or fault strength that release elas-  
20 tic energy from a pre-stressed lithosphere. As  
21 a result, SCR earthquakes can occur in regions  
22 with no previous seismicity and no surface ev-  
23 idence for strain accumulation. They need not  
24 repeat, since the tectonic loading rate is close

25 to zero. Therefore, concepts of recurrence time  
26 or fault slip rate do not apply. As a consequence,  
27 seismic hazard in SCRs is likely more spatially  
28 distributed than indicated by paleoearthquakes,  
29 current seismicity, or geodetic strain rates.

## 1. Introduction

30 Shortly after the discovery of plate tectonics, it was recognized that significant amounts  
31 of elastic strain can be released by large earthquakes on geological structures far from plate  
32 boundary faults, where the vast majority of the Earth's seismic activity takes place [*Sykes*  
33 *and Sbar, 1973; Sbar and Sykes, 1977; Sykes, 1978*]. *Johnston* [1989] discussed issues posed  
34 by large events in stable continental regions (SCRs), which he defined as “*areas where the*  
35 *continental crust is largely unaffected by currently active plate-boundary processes*”. The  
36 diffuse and weak imprint on geology and topography of the active tectonic processes  
37 causing these earthquakes suggested that they involve very low strain rates [*Johnston*  
38 *et al., 1994; Johnston, 1996*], as now confirmed by space geodetic measurements [e.g.,  
39 *Calais et al., 2006; Sella et al., 2007; Nocquet, 2012; Tregoning et al., 2013*]. Parts of  
40 SCRs appear devoid of seismic activity, while others show scattered low to moderate  
41 magnitude earthquakes that are rarely localized on well-defined crustal structures, as  
42 opposed to plate boundaries.

43 Large earthquakes in SCRs are rare: only two dozen events with magnitude 6 or higher  
44 are reported in the historical record worldwide (Figure 1). They are however widespread,  
45 and affect every continent. The 1811-1812 New Madrid events in the Mississippi valley  
46 of the central U.S., the 1988 Tennant Creek earthquakes in Australia, the Basel (1356),  
47 Verviers (1692), Lisbon (1755), or Nice (1887) earthquakes in western Europe, the 1819  
48 and 2001 earthquakes in the ancient Kachchh rift basin in Western India, or the 1690  
49 Manaus and 1955 Parecis basin earthquakes in Brazil are examples of such events in his-  
50 torical times. Some occur at passive margins, glaciated or not in the Late Pleistocene

51 [*Stein et al.*, 1989; *Wolin et al.*, 2012], while others occur well inside continents, previ-  
52 ously glaciated or not [*Johnston*, 1996]. About half of all SCR events occur within rifted  
53 crust at passive margins or within continental interiors, while the other half occur in  
54 other geological settings [*Schulte and Mooney*, 2005]. *Tesauro et al.* [2015] argue that  
55 SCR earthquakes in North America tend to follow craton edges and that tectonic stress  
56 accumulates there, but that correlation is not clear elsewhere [*Schulte and Mooney*, 2005;  
57 *Wolin et al.*, 2012].

58 Although rare, SCR earthquakes can cause widespread damage because attenuation  
59 of seismic energy with distance is typically low in plate interiors [*Hanks and Johnston*,  
60 1992] so that even moderate-size events can be devastating. Examples are the  $M_w$ 6.2  
61 1993 Latur earthquake (India) that caused over 8,000 fatalities and 300 million dollars in  
62 property damage [*Greene et al.*, 2000] or the  $M_w$ 7.7 2001 Bhuj earthquake (India) that  
63 caused more than 20,000 fatalities and more than 4 billion dollars in total damage [*Maurer*  
64 *and Oblitas*, 2001]. A repeat of the 1811-1812  $M$ 7-7.5 New Madrid earthquakes in the  
65 central U.S. today is estimated to cost up to 300 billion dollars in damage [*Spencer et al.*,  
66 2008] and similar figures would likely result from repeats of the 1756 Düren (Germany) or  
67 1356 Basel (Switzerland) earthquakes in these highly populated regions of western Europe  
68 [*Allman and Smolka*, 2001]. Even though SCR earthquakes release only a few percent of  
69 the total seismic energy of the planet, they strike regions where the population and the  
70 infrastructures are often ill-prepared, even in developed countries.

71 Because SRC earthquakes are infrequent, occur in regions where present-day strain rates  
72 are very low, and rupture faults that are difficult to identify geologically, quantifying the

73 associated hazard is a challenge [Ellsworth *et al.*, 2015; Petersen *et al.*, 2015]. Yet, accurate  
74 hazard estimates in SCRs are important for engineering design, in particular in the “post-  
75 Fukushima” era [Joskow and Parsons, 2012] for nuclear infrastructure that are designed  
76 for safety on a 10,000 yr timescale. Because our understanding of the earthquake process  
77 in SCRs leaves much to be desired, hazard calculations often implicitly use concepts and  
78 methods developed for plate boundaries.

79 In this paper, we briefly review the current paradigm [Kuhn, 1962] – i.e., the conceptual  
80 framework shared and applied by the seismology community to explain the earthquake  
81 process. We then describe common characteristics of large SCR earthquakes and review  
82 the state of knowledge on strain build-up in SCRs. We separate these two issues because,  
83 contrary to plate boundaries where the earthquake energy budget is dominated by stress  
84 loading of well-defined active faults where major events occur, the balance between energy  
85 release and strain build up in SCRs is less well understood. We finally discuss the state of  
86 stress and its variations in SCRs and how they may trigger earthquakes. We argue that the  
87 geological and geophysical data currently available require a paradigm shift and propose  
88 that SCR earthquakes are better explained by transient perturbations of local stress or  
89 fault strength than by the slow and localized accrual of tectonic stress on long-lived active  
90 faults.

## 2. The paradigm

91 The current paradigm for the earthquake process is well established, at least in general  
92 [e.g., Kanamori and Brodsky, 2004] (Figure 2, top). Stress builds up on faults over time  
93 as a result of steady plate motions, until their frictional strength is exceeded – at which

94 point they rupture in an earthquake with a certain stress drop. Once unloaded by the  
95 earthquake, the fault takes a variable amount of time to be reloaded to the point of rupture  
96 depending on its strength, the magnitude of the stress drop, and the reloading stressing  
97 rate. Hence strain - the associated measurable quantity - continues building up and  
98 the fault will eventually rupture again so that the cycle repeats, regularly or irregularly.  
99 Over several such cycles, a balance results between the rates at which stress (or strain)  
100 accrues and is released. As a result, “geological slip rates” (actually strain release rates)  
101 should agree with “geodetic slip rates” (actually strain accumulation rates) so that a fault  
102 system conserves energy. As a corollary, past earthquakes, strain accrual rates, and fault  
103 segmentation contain some predictive information for long to medium-range forecasting  
104 of future earthquakes [e.g., *Field et al.*, 2014].

105 This view is supported by geodetic studies at plate boundaries showing that steady plate  
106 motions are accommodated by localized elastic deformation of the crust that accumulates  
107 at steady rates close to active faults. Fault slip during large earthquakes episodically re-  
108 leases this elastic strain so that, over a few hundred years, the rates of strain accumulation  
109 and release balance. For instance, *Tong et al.* [2014] show that geodetic and geologic slip  
110 rates agree within uncertainties along the San Andreas fault, one of the best known active  
111 plate boundary fault systems. This reasoning also holds for slower systems within broad  
112 regions of continental deformation such as the Wasatch fault separating the 3600 m high  
113 Wasatch range from the Great Salt Lake basin. Paleoseismological data showing that the  
114 major normal fault strand has slipped at an average rate of  $1.7\pm 0.5$  mm/yr over the past  
115 10 ka [*Friedrich et al.*, 2003] are consistent with the  $1.6\pm 0.4$  mm/yr strain loading rate



116 determined from GPS measurements [*Chang et al.*, 2006]. In addition, the Gutenberg-  
117 Richter relation derived from the instrumental earthquake catalog, consistent with the  
118 rate of paleoearthquakes identified on the Wasatch fault [*Schwartz and Coppersmith*,  
119 1984; *McCalpin and Nishenko*, 1996] with one M7 event per  $\sim 1,000$  years [*Pechmann*  
120 *and Arabasz*, 1995; *Stein et al.*, 2005], which requires a strain accumulation rate on the  
121 order of 1.5 mm/yr, consistent with the geodetic observations.

122 Whether this steady state model applies to SCRs remains an open question with far-  
123 reaching consequences since it is the underlying principle for probabilistic earthquake  
124 hazard calculations. Traditionally, faults in SCRs have been viewed as analogous to their  
125 plate boundary counterparts, although accumulating strain at very slow rates. If so, large  
126 earthquakes should repeat over time on individual faults as they do at plate boundaries  
127 but with very long recurrence intervals, and faulting should reflect a consistent and ob-  
128 servable strain-rate field. This view is consistent with the interpretation of present-day  
129 intraplate seismic clusters as indicative of long-lived deformation [*Page and Hough*, 2014].  
130 However, unambiguous measurements of strain accumulation on seismically active geo-  
131 logic structures far from plate boundaries remain elusive. In addition, there is increasing  
132 evidence from the paleoearthquake record that SCR faults experience long periods of seis-  
133 mic quiescence separated by short periods of clustered activity [*Clark et al.*, 2012], and  
134 that the loci of large earthquakes varies over time among fault systems [*Liu et al.*, 2011].

135 Alternatively, intraplate faults may be releasing strain stored in the elastic crust over  
136 long intervals but not necessarily localizing observable interseismic strain at their time of  
137 failure [*Calais and Stein*, 2009]. Transient variations in crustal stress or fault strength,

138 if large enough compared to the background tectonic stressing rates, may trigger rupture  
139 [*Long, 1988; Calais et al., 2010*], releasing elastic energy from a pre-stressed lithosphere  
140 [*Feldl and Bilham, 2006*]. Once the available stresses on a fault segment have been released  
141 in an earthquake, the low background tectonic stressing-rate is insufficient to reload that  
142 segment to failure threshold on an observable timescale. Faults may consequently appear  
143 to fail only once, as described below for a number of SCR ruptures [*Crone et al., 2003*].  
144 Thus some clusters of present-day intraplate seismicity are long aftershock sequences  
145 of large events [*Stein and Liu, 2009; Boyd et al., 2015*]. In this view, SCR seismicity  
146 is predominantly a transient feature triggered or inhibited by secondary, non-tectonic  
147 sources of stress change rather than a steady-state response of faults to constant tectonic  
148 loading.

### 3. Some characteristics of large SCR earthquakes

149 Scientific interest in large SCR earthquakes was enhanced after the 1968 Meckering, 1986  
150 Marryat Creek, and 1988 Tennant Creek earthquakes in Australia, and the 1989 Ungava  
151 earthquake in northern Canada. These events, which formed scarps up to 30 km long  
152 and 2 m high, reactivated pre-existing faults within Precambrian crust. They occurred in  
153 landscapes lacking geomorphological features indicative of surface-rupturing earthquakes  
154 during at least the past hundreds of thousands of years [*Adams et al., 1991; Crone et al.,*  
155 *1992; Machette et al., 1993; Crone et al., 1997a; Bent, 1994*].

156 Such sporadic occurrence of large earthquakes, sometimes in the form of a single event  
157 on an old fault lacking evidence of Quaternary or recent activity, is a characteristic shared  
158 by other SCRs (Figure 3). The M6.3 1969 Ceres (South Africa) earthquake, for instance,

159 which ruptured a 20 km-long strike-slip fault segment that did not reach the surface,  
160 occurred in a region with no evidence for previous earthquake activity [*Smit et al.*, 2015].  
161 The Hebron fault in western Namibia shows a well-preserved, 40 km-long scarp with up  
162 to 10 m-throw that formed during a single Holocene event with no evidence for other  
163 Quaternary rupture [*Viola et al.*, 2005; *White et al.*, 2009]. In central India, the 40-  
164 50 km-long Tarpi fault scarp formed in the Holocene in one (or more) thrust-faulting  
165 earthquakes with no evidence for previous events or additional rupture since then [*Copley*  
166 *et al.*, 2014]. In Australia, one of the most arid and slowly eroding SCRs where scarps  
167 may be preserved for tens of thousands to millions of years, *Clark et al.* [2012] identified  
168 300 small scarps and noted their poor spatial correlation with contemporary seismicity  
169 [*Clark and Leonard*, 2015].

170 In the Central and Eastern U.S., where environmental conditions are less favorable for  
171 the preservation of scarps, *Madole* [1988] and *Crone and Luza* [1990] report two surface-  
172 breaking events 1,200–1,300 years ago that ruptured a ~60 km-long segment of the Pa-  
173 leozoic Meers fault in Oklahoma, with no evidence for other events in at least the past  
174 120,000 years. *Crone et al.* [1997b] report evidence for three large earthquakes during  
175 the past 25 ka on the 44 km-long Cheraw fault in Colorado, with the most recent event  
176 in the early Holocene. Large historical earthquakes occurred in 1755 ( $M > 6.0$ , near Cape  
177 Ann, MA), 1811-1812 ( $M 7-7.5$ , near New Madrid, MO), 1843 ( $M 6.3$ , northeast Arkansas),  
178 1895 ( $M 6.6$ , near New Madrid, MO), and 1886 ( $\sim M 7$  near Charleston, SC) [e.g., *Johnston*  
179 *et al.*, 1994].

180 The New Madrid sequence consisted of four  $M > 7$  earthquakes between December 1811  
181 and February 1812 [Nuttli, 1973; Johnston, 1996; Hough *et al.*, 2000]. Seismicity continues  
182 today and likely outlines the 1811-1812 ruptures of the Cottonwood Grove and Reelfoot  
183 faults [Mueller *et al.*, 2004; Johnson *et al.*, 2014]. Liquefaction features in the upper  
184 Mississippi embayment show earthquakes similar in magnitude and location to the 1811–  
185 1812 events in  $1450 \pm 150$  C.E.,  $900 \pm 100$  C.E.,  $300 \pm 200$  C.E., and  $2350 \pm 200$  B.C.E. [Tuttle  
186 *et al.*, 2002]. Holbrook *et al.* [2006] use reconstructed Holocene Mississippi River channels  
187 to document possible additional events – or clusters of events – at  $2244 \pm 269$  B.C.E and  
188  $1620 \pm 220$  B.C.E likely related to activations of the Reelfoot fault. Put together, these  
189 observations suggest that, during the Holocene, the region experienced millennial-scale  
190 temporal clustering of earthquakes interrupted by very long – up to several thousand  
191 years – intervals of seismic quiescence. The lack of significant topography in the region –  
192 indicating a relatively short period of fault activity – together with seismic reflection and  
193 trenching studies that find an increase in slip rate on the Reelfoot fault by four orders of  
194 magnitude about 10 kyr ago [Van Arsdale, 2000], show that the NMSZ must have been  
195 recently activated.

196 Contrary to other SCRs, Western Europe underwent relatively recent, large-scale tec-  
197 tonic activity with the emplacement of the Cenozoic rift system of western and central  
198 Europe [Illies *et al.*, 1981; Dèzes *et al.*, 2004]. Some of the most seismically active re-  
199 gions today follow the overall trace of that structure, such as the Roer Valley Graben,  
200 bounded by Quaternary scarps related to earthquake activity [Ahorner, 1975; Camelbeeck  
201 *and Meghraoui*, 1998; Vanneste *et al.*, 2013]. Paleoseismological investigations along its

202 western border have identified a 12 km–long by 15–20 m–high scarp along the Bree fault  
203 that offsets the 350–700 ka main terrace of the Maas River by about 40 m [*Camelbeek and*  
204 *Meghraoui*, 1996, 1998]. One of the trenches across the Bree scarp shows evidence for five  
205 earthquakes in the past 100 ka, the most recent  $\sim 3$  ka B.P. associated with a 10 km–long  
206 by 0.5 m–high rupture scarp [*Vanneste et al.*, 2001]. On the opposite side of the Roer  
207 graben (Peel fault), similar investigations identified three large earthquakes within the  
208 past 25 ka [*van den Berg et al.*, 2002]. Detailed geomorphic analyses of the Bree and Peel  
209 scarps show that these border faults of the Roer Valley Graben were continuously active  
210 since the Middle Pleistocene, with earthquake magnitudes likely ranging from 6.3 to 7.0  
211 [*Camelbeek et al.*, 2007; *Vanneste et al.*, 2013].

212 The largest known earthquake in the Roer Valley Graben is the M5.7 18 February 1756  
213 earthquake near Düren, Germany [*Camelbeek et al.*, 2007]. However, the three largest  
214 historical earthquakes in this part of Europe with estimated magnitude around or greater  
215 than 6.0 occurred outside of the graben in 1382 (southern North Sea), 1580 (Dover Strait),  
216 and 1692 (Verviers, northern Belgian Ardenne). Therefore, most of the seismic energy  
217 release since the Middle Ages in this part of Western Europe occurred outside the Roer  
218 Valley Graben, despite the graben’s dominance in Western European seismic activity over  
219 the Quaternary. A recent offshore survey in the epicentral area of the 1580 Dover Strait  
220 earthquake showed no evidence for persistent faulting during the Quaternary [*Garcia-*  
221 *Moreno et al.*, 2015]. Similarly, the Hercynian–age Hockai fault activated during the 1692  
222 Verviers earthquake shows no evidence for previous events in the Quaternary [*Lecocq et al.*,  
223 2008].

224 The largest documented paleo-earthquakes in stable Europe likely occurred in the tec-  
225 tonically stable Fennoscandian craton in the late Pleistocene/early Holocene between 11  
226 and 9 ka [*Muir-Wood*, 1989; *Olesen et al.*, 2004; *Mörner*, 2005; *Lagerbäck and Sundh*,  
227 2008; *Jakobsson et al.*, 2014; *Olesen et al.*, 2013; *Smith et al.*, 2014]. They formed nu-  
228 merous scarps ranging from small-scale fractures to the 150 km-long Pärvie fault scarp,  
229 with offsets exceeding 15 m in places [*Muir-Wood*, 1989; *Lagerbäck and Sundh*, 2008].  
230 Some of these earthquakes may have been larger than M8 [*Muir-Wood*, 1989; *Lagerbäck*,  
231 1992; *Arvidsson*, 1996; *Lindblom et al.*, 2015] whereas the historic and instrumental seis-  
232 micity of Fennoscandia rarely exceeds  $M_w$ 5. There is little field evidence for on-going or  
233 repeated ruptures, but trenches across some faults indicate that faulting occurred in a  
234 single event [*Lagerbäck and Sundh*, 2008]. The clustering of these events 11-9 ka ago is a  
235 strong indication of a link with the last deglaciation [*Muir-Wood*, 1989; *Mörner*, 2005], as  
236 demonstrated by mechanical modelling studies [*Wu et al.*, 1999; *Wu and Johnston*, 2000;  
237 *Lambeck and Purcell*, 2003; *Turpeinen et al.*, 2008; *Steffen et al.*, 2014].

238 These examples illustrate the diversity of faults capable of generating large earthquakes  
239 in SCRs. Some occur in regions devoid of current seismicity or evidence for Quaternary  
240 ruptures. Some appear to have ruptured only once in recent times, while others show  
241 evidence for multiple events, sometimes clustered in time, separated by quiescent intervals  
242 of 10,000 to more than 100,000 years. Steady-state earthquake activity does not appear  
243 to persist in the long-term on any single fault. Hence seismic activity in SCRs appears to  
244 be episodic and sometimes clustered on faults that are active during relatively short time

245 intervals, and then migrates to other structures [*Crone et al.*, 1997a; *Stein et al.*, 2009;  
246 *Crone et al.*, 2003; *Clark et al.*, 2012].

247 A spectacular example of this “clustered and migrating” nature of large earthquakes in  
248 low strain rate regions is the North China plain, a flat-lying area bounded by the Shanxi  
249 rift and the coast of the Yellow Sea to the west and east, and extending north-south from  
250 Beijing to Shanghai. Geodetic strain rates in this region are very low, less than  $10^{-9}$  yr<sup>-1</sup>  
251 [*Calais et al.*, 2006; *Zhao et al.*, 2015]. *Liu et al.* [2011] use a historical earthquake  
252 catalog complete to  $M > 6$  since 1300 A.D that includes 49 events with  $M > 6.5$  and at  
253 least four earthquakes with  $M > 8$  to show that these large earthquakes migrate between  
254 fault systems across distances much larger than their rupture length, hence precluding  
255 static stress transfer as triggering mechanism. Over the time interval considered, none  
256 of the fault systems was activated more than once. The slow tectonic loading in such a  
257 system therefore appears to be shared by many faults of similar strength. Individual fault  
258 may remain stable for a long time and become active for a short period only. *Liu et al.*  
259 [2011] also document complementary transfer in moment release rate between some faults:  
260 increase on one correspond to decrease on the other, indicating that they are mechanically  
261 coupled over large distances.

262 These examples of SCR earthquakes and active – or capable [*Machette*, 2000] – faults  
263 show a variety of behaviors that is not seen at plate boundaries. Faults like Meers, Hebron,  
264 or Tennant Creek are isolated structures that show no evidence for more than one event  
265 in the paleoearthquake record. Faults like the Reelfoot fault in the NMSZ or the Bree  
266 fault in the Lower Rhine Graben show repeated earthquakes over 10,000 to 100,000 yr.

267 In North China and Western Europe, faults are organized in a system with indications of  
268 long-distance interactions between them [*Liu et al.*, 2011].

269 The lack of persistence of the seismic activity on the rarely activated faults in SCRs  
270 raises three additional issues. First, the behavior described above implies that a meaning-  
271 ful recurrence interval cannot be defined for many SCR faults, particularly those where  
272 only one earthquake or long intervals of seismic quiescence are documented. Hence, the  
273 notion of a “seismic cycle” or that of a “slip rate”, which fail to capture their highly  
274 non-steady state behavior, may not be applicable to SCR faults.

275 Second, if these inherited structures are only reactivated a few times – some perhaps  
276 only once – with long intervals of seismic quiescence, they are likely not loaded individually  
277 at a constant rate, in contrast to plate boundary faults. The most active SCR region in  
278 the late Holocene, the New Madrid seismic zone, shows strain accumulation at a rate that  
279 is indistinguishable from zero while the seismic energy release over the past 3,000 years  
280 would require at least 2 mm/yr of strain accrual at steady-state [*Calais and Stein*, 2009;  
281 *Craig and Calais*, 2014]. This argues against interseismic strain localization on individual  
282 SCR fault zones.

283 Third, the single or episodic activity of most SCR faults does not represent their long-  
284 term behavior, during which the faults are mostly inactive. Their short time intervals of  
285 seismicity require shorter term stress or fault strength variations, and thus argues against  
286 earthquake triggering being a direct manifestation of tectonic stresses, which change slowly  
287 on time scales of millions of years.



#### 4. Is there measurable strain within SCRs?

288 Earthquakes provide information on the rate at which elastic strain is released, which  
289 is related to – but distinct from – the rate at which it builds up in the crust. Space  
290 geodetic techniques such as the Global Positioning System (GPS), widely used to measure  
291 strain accumulation on plate boundary faults, have therefore been deployed to detect  
292 strain accrual on seismogenic SCR faults. The heavily populated New Madrid Seismic  
293 Zone (NMSZ) in the central U.S., locus of four earthquakes of magnitude 7 or greater in  
294 1811-1812 (see above) and where seismic activity continues today, became a prime target  
295 for both geodetic investigations and for research on paleoearthquakes and local crustal  
296 structures that may accommodate long term faulting.

297 Early geodetic measurements combining space and terrestrial data [*Liu et al.*, 1992]  
298 claimed 5-7 mm/yr of relative motion across the southern branch of the NMSZ (Fig-  
299 ure 4). It was argued that this rate was in agreement with a steady-state fault system  
300 releasing one  $M_w$ 8 earthquake every 500 to 1000 years, as expected then [*Johnston*, 1996].  
301 However, similar observations in the northern part of the NMSZ led to inconclusive re-  
302 sults, showing motions less than 3 mm/yr across the fault system [*Snay et al.*, 1994].  
303 Similarly, episodic GPS measurements over the entire NMSZ reported no motion within  
304 uncertainties, placing an upper bound on deformation of 2.5 mm/yr [*Newman et al.*, 1999].  
305 *Argus and Gordon* [1996] and *Dixon et al.* [1996] used continuously recording GPS sta-  
306 tions throughout the plate interior to establish an upper bound of 2 mm/yr for residual  
307 motions across the Central–Eastern U.S., that was later reduced to 0.5 mm/yr thanks to  
308 longer time series and a much larger number of measurement sites [*Calais et al.*, 2006].

309 Analyses of continuous GPS data within the NMSZ continued to show no motion within  
310 uncertainties with an upper bound that decreased as time series duration increased [*Calais*  
311 *et al.*, 2005; *Calais and Stein*, 2009]. Strain rates in the NMSZ “*comparable in magni-*  
312 *tude to those across active plate boundaries*” [*Smalley et al.*, 2005] were later shown to  
313 result from an unexplained instrumental offset in the data [*Calais et al.*, 2005]. A recent  
314 comprehensive reanalysis of continuous GPS data in the Central-Eastern U.S. confirms  
315 earlier results with motions that are consistently within the 95% confidence limit of zero  
316 deformation and places an upper bound on strain accrual of 0.2 mm/yr and 0.5 mm/yr in  
317 the New Madrid and Wabash Valley Seismic Zones, respectively [*Craig and Calais*, 2014;  
318 *Boyd et al.*, 2015].

319 Thus, the best geodetically studied SCR region, which experienced M7+ earthquakes in  
320 1811-1812 as part of a longer Holocene sequence of large events, shows no demonstrable  
321 deformation and a maximum rate of strain accrual  $\lesssim 0.2$  mm over 200 k, or  $\lesssim 10^{-9}$  yr $^{-1}$ .  
322 More importantly, this upper bound on strain accrual is too low to account for the moment  
323 released by known large earthquakes of the past  $\sim 5,000$  years in the NMSZ (Figure 5;  
324 *Craig and Calais* [2014]). Taken together, the geodetic and paleoseismological data there-  
325 fore exclude steady-state fault behavior over that time period. Thus the rate at which  
326 the NMSZ is loaded, its mechanical strength, or both, vary with time. The fact that  
327 strain is currently not accumulating fast enough to account for large Holocene earth-  
328 quakes also implies that the NMSZ seismic activity must be releasing elastic strain energy  
329 that accumulated over a longer time interval.

330 Similar results are available for other plate interiors. *Nocquet and Calais* [2003] used  
331 continuous GPS measurements to show that Central Europe, defined as the region east  
332 of the Rhine Graben and north of the Alps and the Carpathians, behaves rigidly at a  
333 0.4 mm/yr level. An updated Europe-wide solution [*Nocquet, 2012*] confirms these findings  
334 and lowers the upper bound to 0.2 mm/yr for stable Europe, i.e., the continental region  
335 south of 52° N where the effect of Glacial Isostatic Adjustment (GIA) is insignificant  
336 and north of the tectonically active Alpine belts. This upper bound of 0.2 mm/yr applies  
337 to the seismically active Pyrenees [*Rigo et al., 2015*] and the Rhine Graben [*Nocquet and*  
338 *Calais, 2004; Fuhrmann et al., 2013*]. In South Africa, *Saria et al.* [2013] and *Hackl et al.*  
339 [2011] analyze a country-wide continuous GPS network and show that relative motions  
340 are indistinguishable from zero with an upper bound of 0.6 mm/yr. *Beavan et al.* [2002]  
341 use continuous GPS stations on the Pacific and Australian Plates and show that they each  
342 fit a rigid plate model with an RMS residual of 0.4 mm/yr. *Tregoning et al.* [2013] recently  
343 updated this number for the Australian continent, showing that present-day deformation  
344 is indistinguishable from zero with an upper bound of 0.2 mm/yr.

345 The search for tectonic strain accumulation within plate interiors has so far failed to  
346 identify seismically active regions where strain currently accrues at a measurable rate.  
347 However, horizontal deformation caused by GIA in plate interiors is easily captured by  
348 space geodetic measurements, at least in the current uplift areas. The BIFROST perma-  
349 nent GPS network in Sweden and Finland provided the first three-dimensional map of  
350 GIA over Fennoscandia [*Johansson et al., 2002; Lidberg et al., 2010; Kierulf et al., 2014*].  
351 Larger-scale studies have also identified horizontal motions outside of the uplifting areas

352 in Western Europe [*Nocquet et al.*, 2005] and North America [*Calais et al.*, 2006; *Sella*  
353 *et al.*, 2007] indicating that large parts of those two continental interiors are experiencing  
354 long-wavelength strain caused by GIA.

355 Figure 6 shows a recently updated geodetic solution for stable North America following  
356 the methods described in *Calais et al.* [2006]. The velocities are residuals with respect  
357 to a subset of GPS stations located south of 42°N whose velocities fit a rigid rotation  
358 model with a reduced  $\chi^2$  close to unity. The regional pattern is consistent with that  
359 expected from GIA [*Peltier et al.*, 2015], with extension (up to  $10^{-8}$  yr $^{-1}$ ) coincident with  
360 the uplift areas and shortening ( $1-3 \times 10^{-9}$  yr $^{-1}$ ) associated with the subsiding forebulge.  
361 A comparison with instrumental seismicity shows an interesting anticorrelation with GIA  
362 strain rates, except perhaps in the Lower St Lawrence area [*Mazzotti et al.*, 2005]. In  
363 other words, intraplate areas that are being strained as a result of GIA are not the ones  
364 that experience seismicity today. In addition, if GIA strain accrual was responsible for  
365 NMSZ earthquakes, the  $\sim 1$  mm/yr N-S shortening observed between the Great Lakes  
366 and the Gulf of Mexico would cause left-lateral and normal slip on the NE-SW-oriented  
367 New Madrid faults [*Craig and Calais*, 2014]. This is opposite to observations that show  
368 right-lateral and reverse motion consistent with large-scale tectonic stresses [*Hurd and*  
369 *Zoback*, 2012a]. These observations indicate that GIA strain accrual did not trigger New  
370 Madrid earthquakes, as also inferred by *Wu and Johnston* [2000] on the basis of a modeling  
371 study.

## 5. The state of stress in SCRs

372 Stresses within the continental lithosphere result from the superposition of forces along  
373 plate boundaries that are transmitted into their interiors, forces at the base of the litho-  
374 sphere resulting from the relative motion between plates and mantle flow (the “shear  
375 tractions”), and buoyancy forces arising from lateral gradients of gravitational poten-  
376 tial energy caused by topography and intralithospheric density distributions [*Fleitout and*  
377 *Froidevaux*, 1983; *Bird et al.*, 2008]. These “tectonic stresses” remain regionally coherent  
378 over very long times – millions of years – because the underlying processes vary slowly.

379 Similarly, stresses within continents vary only slowly with distance. Stress indicators,  
380 notably earthquake focal mechanisms, show broad areas with consistent maximum com-  
381 pressive horizontal stress ( $Sh_{max}$ ) directions consistent with plate-driving forces, locally  
382 modified by lithospheric properties in some regions [*Zoback and Zoback*, 1989; *Müller*  
383 *et al.*, 1992; *Heidbach et al.*, 2007, 2010]. In North America,  $Sh_{max}$  shows a very consis-  
384 tent WSW-ENE direction across the central and eastern U.S., all the way to southeastern  
385 Canada [*Hurd and Zoback*, 2012a; *Herrmann et al.*, 2011]. This consistency is visible  
386 both in “natural” earthquakes and in the human-induced seismicity currently widespread  
387 throughout Oklahoma and part of Texas. *McNamara et al.* [2015] show that well-induced  
388 earthquakes in central Oklahoma, some with magnitudes reaching 5.7 [*Keranen et al.*,  
389 2013, 2014], occur on faults that are favorably oriented in a ENE-WSW compressive  
390 stress field, with focal mechanisms consistent with this background tectonic stress field.  
391 The same observation holds for the NMSZ, where *Hurd and Zoback* [2012b] show focal

392 mechanism P-axes consistent with ENE-WSW Shmax orientation over much of the central  
393 and eastern U.S.

394 That the crust breaks on pre-existing faults favorably oriented with respect to the re-  
395 gional tectonic stress field with source mechanisms consistent with that stress field does  
396 not necessarily mean that this background tectonic stress is responsible for bringing in-  
397 dividual faults to failure. For instance, the mechanism triggering recent seismicity in  
398 Oklahoma and Texas is wastewater injections following oil recovery, which increase pore  
399 pressure at depth, lowering effective normal stress on faults and bringing them closer to  
400 failure [Keranen *et al.*, 2014]. This mechanism was likely the cause of the  $M_w$ 5.7, Novem-  
401 ber 2011 earthquake in central Oklahoma, which was broadly felt and caused damage in  
402 the epicentral region [Keranen *et al.*, 2013]. Ample evidence shows that earthquakes are  
403 sometimes triggered by fluid injections during oil recovery or mining operations and by the  
404 filling of water reservoirs [Rothé, 1968; Raleigh *et al.*, 1972; Simpson, 1976; Gupta, 1985].  
405 Two mechanisms have been invoked to explain the latter, either the increase of elastic  
406 stresses due to the flexure of the crust under the load, or the lowering of effective normal  
407 stress on faults as water diffuses down to hypocentral depths [Simpson *et al.*, 1988].

408 Seismic swarms of natural origin are also attributed to fluid overpressure following the  
409 diffusion of mantle volatiles [Weise *et al.*, 2001; Špičák and Horálek, 2001; Cappa *et al.*,  
410 2009] or meteoric water [Hainzl *et al.*, 2006; Costain and Bollinger, 2010; Got *et al.*, 2011;  
411 Leclère *et al.*, 2013] to seismogenic depth. Many such studies argue that fluid overpressure  
412 at depth plays a key role in earthquake nucleation by lowering effective stress on fault

413 segments that are nearly critically stressed for shear failure [*Sibson, 1990; Cappa et al.,*  
414 *2009; Wang and Manga, 2009*].

415 Alternately, earthquakes can be triggered by changes in elastic stresses driven by the  
416 loading or unloading of the crust by surface or ground water. *González et al.* [2012] showed  
417 that stress changes caused by water extraction from a shallow aquifer likely triggered a  
418  $M_w$ 5.1 earthquake near Lorca, Spain, in 2011. Its source mechanism indicates reverse  
419 faulting on the SW-NE-oriented Murcia fault and reflects the regional stress field imparted  
420 by the oblique convergence between Nubia and Eurasia [*Nocquet and Calais, 2004*]. *Heki*  
421 [2003] explains seasonal cycles in earthquake occurrence in northern Japan as a result of  
422 the modulation of the regional stress field by stresses of a few kPa caused by snow loading.  
423 *Bollinger et al.* [2007] and *Bettinelli et al.* [2008] report seasonal strain and stress variations  
424 in the Nepal Himalaya that correlate with seasonal variations in seismicity, with summer  
425 seismicity suppressed by stress-loading accompanying monsoon rains.

426 Though most of the triggered earthquakes referred to above are small, some may be  
427 much larger. In northern Sweden and Finland, the series of M7–8 end-glacial earthquakes  
428 around 9,500 years ago has been interpreted as a result of decreased normal stresses on  
429 steeply dipping reverse faults as the Fennoscandian ice sheet was rapidly melting [*Wu*  
430 *et al., 1999; Turpeinen et al., 2008; Steffen et al., 2014*]. In the Basin and Range province  
431 of the Western U.S., *Hetzl and Hampel* [2005] show that the increased slip rate on the  
432 Wasatch fault since about 17 ka – more specifically the “strain release rate”, determined  
433 from the paleoearthquake record [e.g., *Friedrich et al., 2003*] – could be explained by the  
434 stress changes induced by a regression of Lake Bonneville and the melting of glaciers in

435 the Uinta and Wasatch mountains. In the upper Mississippi embayment of the Central  
436 U.S., *Calais et al.* [2010] showed that an intense erosional event between 16 and 10 ka  
437 caused upward flexure of the lithosphere and a reduction of normal stresses in the upper  
438 crust sufficient to unclamp pre-existing faults close to critical failure, possibly triggering  
439 the sequence of large Holocene earthquakes in the region. Once a large earthquake has  
440 occurred, stress changes may trigger additional regional events via elastic (coseismic) or  
441 viscoelastic (postseismic) stress transfer and a clustered sequence of events may develop  
442 [*Kenner and Segall*, 2000; *Mueller et al.*, 2004].

443 It therefore appears that the background tectonic stress field in the lithosphere can  
444 be effectively modulated by stress changes of external, non-tectonic, origin. Where the  
445 tectonic stressing rates are fast, as is typically the case at plate boundaries, external  
446 forcing may have only a minor modulating effect on the seismic cycle [*Luttrell et al.*,  
447 2007]. *Luttrell and Sandwell* [2010] show that eustatic sea level changes can modify stress  
448 on near shore faults at  $\sim 100$  Pa/yr, which is about 100 times slower than the stressing  
449 rate due to plate motions at major plate boundaries such as California or New Zealand.

450 In SCRs, however, tectonic stressing occurs at rates that are at least 100 times lower  
451 than at major plate boundaries, so external forcings may dominate and localize earthquake  
452 activity in space and time. Hence, the timing and location of SCR earthquakes may  
453 be largely independent of long-term tectonic loading under a regional, essentially time-  
454 invariant, tectonic stress field, but instead be determined by small transient stress changes  
455 in a crust close to failure equilibrium. Regardless of the specific transient stress change



456 that brings a fault to failure, the resulting rupture mechanism will however be consistent  
457 with the background static tectonic stress field, which defines the style of faulting.

458 The hypothesis that SCR faults are in a state of failure equilibrium is supported by (1)  
459 in situ stress measurements in deep wells, which agree well with predictions from Coulomb  
460 frictional-failure theory [Zoback *et al.*, 1993], (2) seismicity induced by fluid injection and  
461 reservoir impoundment, as described above, and (3) triggering of earthquakes by small  
462 static Coulomb stress changes caused by nearby earthquakes [Stein *et al.*, 1992, 1996].  
463 That SCR faults are critically stressed does not necessarily limit the strength of the crust  
464 as a whole. Townend and Zoback [2000] show that, for a high crustal permeability –  
465 hence near-hydrostatic pore pressures – critically stressed faults maintain a high crustal  
466 strength, allowing SCR crust to sustain large differential stresses.

467 Seismically active areas within SCRs are sometimes interpreted as the result of lo-  
468 cal concentrations of tectonic stress or as mechanically weak regions [e.g., Sykes, 1978].  
469 Various mechanisms have been proposed including stress concentration at intersecting  
470 faults [Talwani, 1999], around buried intrusions in the crust [Campbell, 1978; Zoback and  
471 Richardson, 1996; Pollitz *et al.*, 2001], or at the tip of a low velocity upper mantle seismic  
472 anomaly [Zhan *et al.*, 2016]. Other proposed mechanisms involve local weakening of the  
473 lower crust either thermally– [Grollmund and Zoback, 2001; Kenner and Segall, 2000]  
474 or geochemically–induced [Chen *et al.*, 2016] or bulk weakening in regions where the me-  
475 chanically strong mantle lithosphere is absent [Tesauro *et al.*, 2015]. Although all these  
476 mechanisms are plausible, they would persist over long geologic time intervals, whereas  
477 SCR seismicity does not. For instance, Van Arsdale [2000] show that the NMSZ was ac-

478 tivated around 10,000 yrs ago after millions of years of tectonic quiescence during which  
479 all the processes listed above would have been operating, had they existed. Therefore,  
480 although such mechanisms may locally perturb the long-term, static, tectonic stress field  
481 of a continental interior, they do not explain why SCR seismicity is episodic, with long  
482 and variable seismically quiet time intervals, and some faults rupturing only once.

483 Moreover, even if these processes were concentrating stress, the overall stress changes  
484 inside continents – including regions of stress concentration – arising from boundary and  
485 buoyancy forces, must occur at very low rates, as shown by the lack of detectable strain  
486 accumulation in continental interiors. The series of large earthquakes identified in the  
487 NMSZ in the past 3,000 yrs in the absence of detectable strain buildup ( $<0.2$  mm/yr  
488 over 100 km) argues against the notion that large SCR earthquakes release elastic strain  
489 energy that accumulates locally over short ( $\sim 1,000$  yrs) time scales, as described above.

490 In addition, the notion that SCR earthquakes preferentially occur within zones of crustal  
491 weakness is at odds with a number of observations *Zoback et al.* [1985]. The frictional  
492 strength of faults and unfaulted rock are similar, as shown by laboratory and in situ stud-  
493 ies, so that there is no reason why pre-existing fault zones in the crust would have low  
494 strength. Also, seismically active areas in SCRs show no evidence of the anomalous stress  
495 field expected near weak regions. In North America, for instance,  $S_{\max}$  directions are  
496 uniform throughout the central and eastern U.S., with little to no variations in the seismi-  
497 cally active NMSZ [*Hurd and Zoback, 2012b*]. Hence *Zoback and Zoback* [1981] conclude  
498 that “*seismicity in the central and eastern U.S. appears to be occurring in response to a*  
499 *broad, regionally uniform regional stress field*”.

## 6. Conclusion: a possible mechanism for large SCR earthquakes

500 It appears that SCR earthquakes release strain from a pre-stressed lithosphere where  
501 faults are at failure equilibrium and can be triggered by small transient stress changes  
502 caused, for instance, by surface load variations or fluid diffusion in the crust. If so, these  
503 earthquakes do not require a significant tectonic loading rate, which has not been observed  
504 in continental interiors [*Nocquet, 2012; Tregoning et al., 2013; Craig and Calais, 2014*],  
505 or long term strain localization on specific crustal structures. This mechanism requires  
506 the lithosphere to be accumulating and storing elastic strain over longer intervals than is  
507 observable by geodesy or paleoseismology.

508 Unfortunately, there is no present way to directly test the hypothesis of stored back-  
509 ground strain. However, the lack of evidence for localized interseismic strain accumu-  
510 lation in SCRs, together with observations that faults communicate over regions much  
511 larger than their length [*Liu et al., 2011*], suggest that they draw elastic energy from a  
512 broad, shared elastic strain reservoir. Similarly, the fact that strain from far-field motions  
513 is currently not accumulating fast enough to account for large earthquakes in the U.S.  
514 midcontinent indicates that large earthquakes there release elastic strain energy stored in  
515 the crust over long geologic time intervals [*Calais et al., 2010; Craig and Calais, 2014; Liu*  
516 *et al., 2014*].

517 Another indirect line of evidence is *Craig et al. [2016]*'s observation that the end-glacial,  
518 reverse-faulting earthquakes of Fennoscandia occurred while the horizontal strain-rate was  
519 extensional. Consequently, faulting did not release extensional strain accumulating at the  
520 time of failure, but instead released compressional strain that had accumulated through

521 long-term tectonic forcing (Figure 7). This forcing is likely due to the ridge-push force  
522 exerted by the cooling and thickening oceanic lithosphere formed at the Mid-Atlantic  
523 Ridge [Gölke and Coblenz, 1996; Pascal *et al.*, 2010], with a possible contribution from  
524 compressional stresses generated by ice loading during the last glacial period.

525 In addition to tectonic forcing, *Schrank et al.* [2012] show that thermal-elastic stresses  
526 in excess of 100 MPa can be stored in the crust during the burial of granite, placing the  
527 buried rock in highly pre-stressed state. Experiments show that below 400°C, expected in  
528 the brittle upper crust, only 10% of the total elastic energy is dissipated, with relaxation  
529 times of millions of years. Therefore, thermal elasticity may also bring the continental  
530 crust close to failure and contribute to a stress reservoir from which earthquakes can draw  
531 elastic energy.

532 Large earthquakes outside plate boundaries also occur within “stable oceanic regions”,  
533 as shown on Figure 1, but in much fewer numbers. Though this difference could be due  
534 to a lack of historical information on oceanic regions, it may also reflect the fact that the  
535 oceanic crust is more homogeneous than continental crust, if only because its age never  
536 exceeds 200 million years, and less subject to local or regional perturbations of stress or  
537 fault strength. For instance, hydrological loads do not change ocean bottom pressures, as  
538 shown by the very low secular/seasonal gravity changes derived from GRACE over the  
539 oceans compared to continental regions [e.g., *Wouters et al.*, 2014].

540 If faults in SCRs are at failure equilibrium in a pre-stressed crust able to sustain large  
541 differential stresses, then the occurrence of SCR earthquakes in time and space is better  
542 explained by transient perturbations of stress or fault strength than by the slow accu-

543 mulation of tectonic stress on long-lived active faults. In other words, while tectonic (or  
544 thermal) stress provides the energy that is released during large SCR earthquakes, earth-  
545 quake occurrence results from a local and temporary perturbation of stress or crustal  
546 strength near a fault that is favorably oriented relative to the regional tectonic stress  
547 field. These transient perturbations may result from fluid pore pressure increase at earth-  
548 quake nucleation depth, or from local changes in “secondary stresses” – for instance caused  
549 by surface loading/unloading. Hence, earthquake sequences appear episodic and clustered  
550 – sometimes involving a single rupture – rather than persistent.

551 If faults in SCRs are activated by transient loading stresses or fluid overpressures and  
552 draw energy from a long-lived and broadly distributed “strain reservoir”, they need not be  
553 steady-state systems, on any time scale. For the same reasons, large SCR earthquakes may  
554 occur in the absence of geodetic evidence for local strain accumulation around the faults  
555 that are activated. If so, geodetic measurements may contain limited information about  
556 the seismic potential of faults in SCRs, as shown by the lack of correlation between current  
557 strain accrual and seismicity in stable North America (Figure 6). The same holds for the  
558 location of past large earthquakes or current seismicity, which indicate where strain release  
559 occurred but not necessarily where it accrues today in preparation for future events. In  
560 that view, seismic hazard in SCRs is likely to be more spatially distributed than indicated  
561 by paleoearthquakes, current seismicity, or geodetic strain rates.

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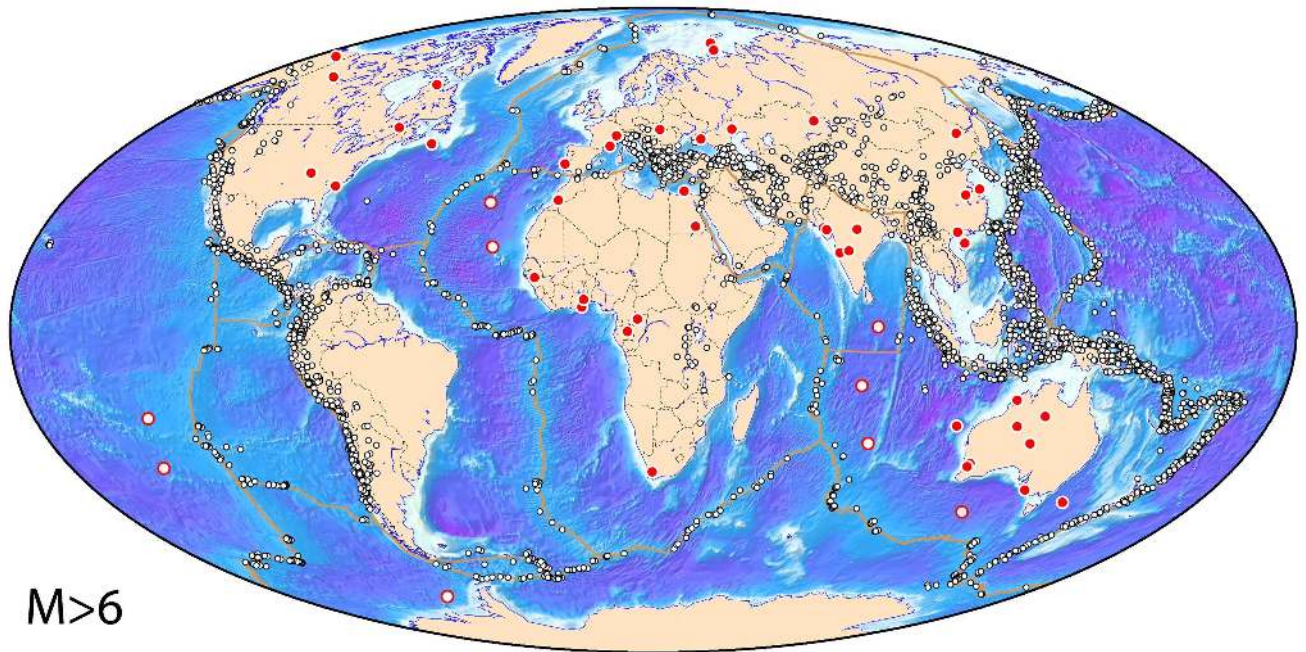
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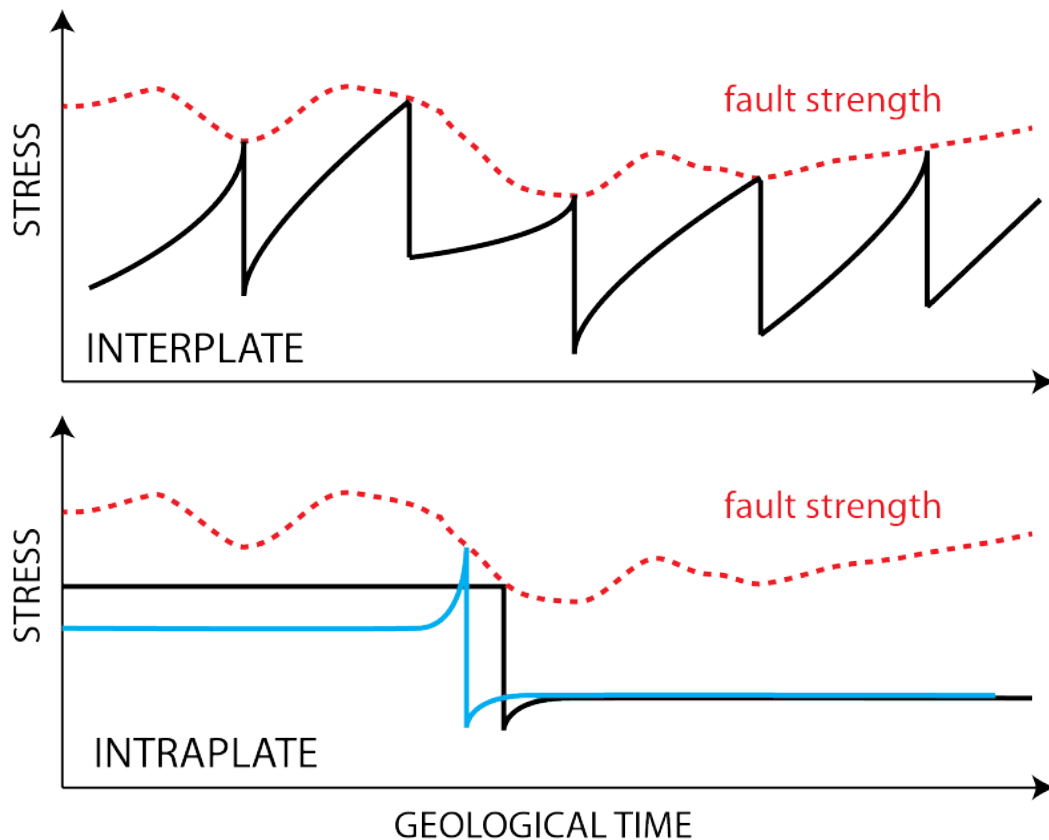
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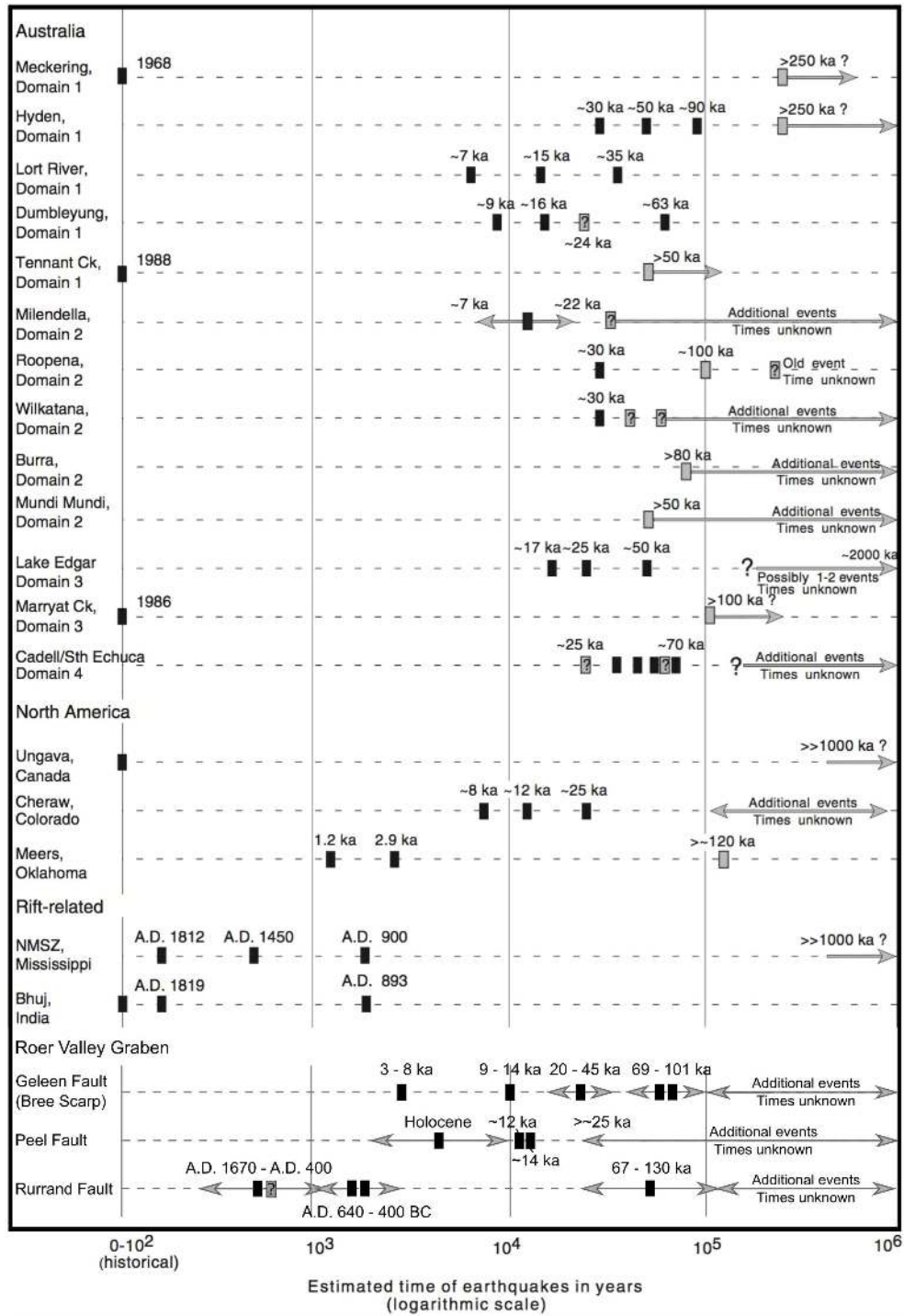
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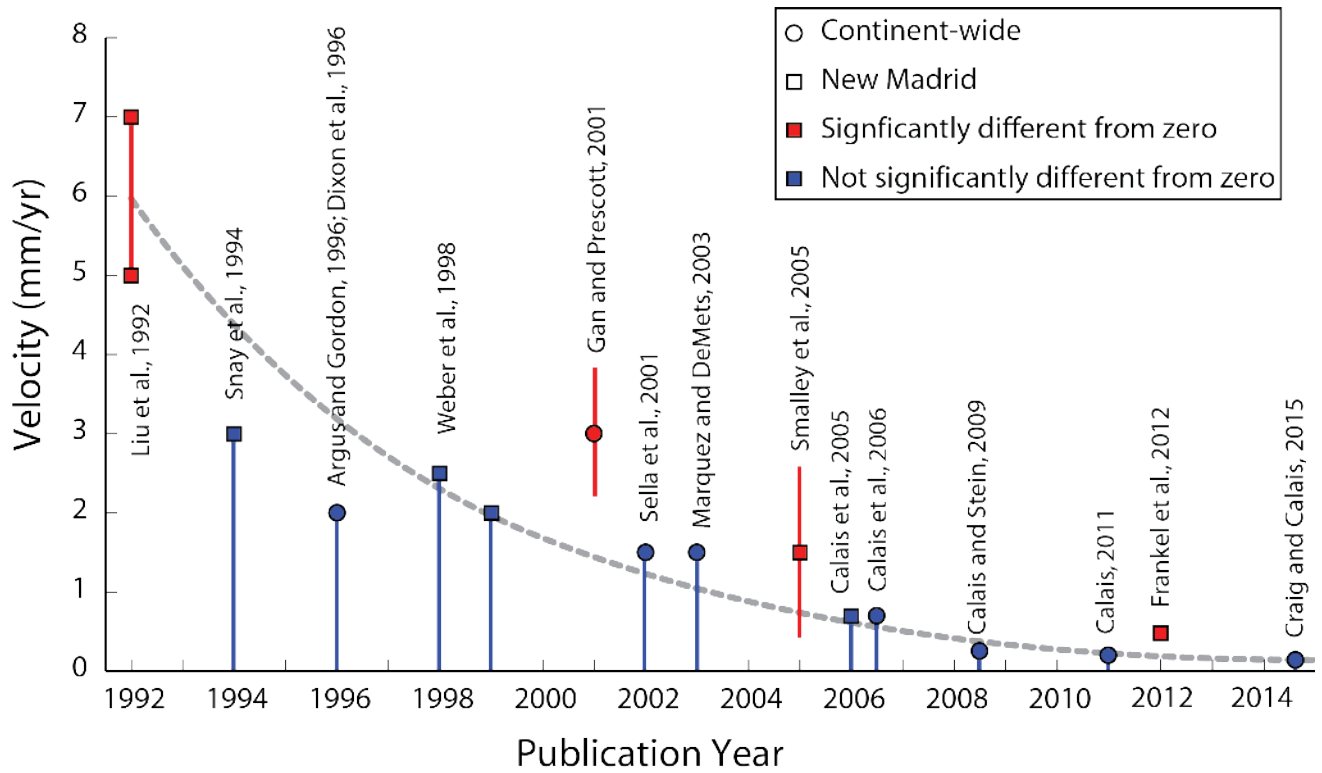
**Figure 1.** Worldwide seismic activity (<http://earthquake.usgs.gov/>). Large circles show  $M>6$  intraplate earthquakes: red for stable continental regions, white for stable oceanic regions.



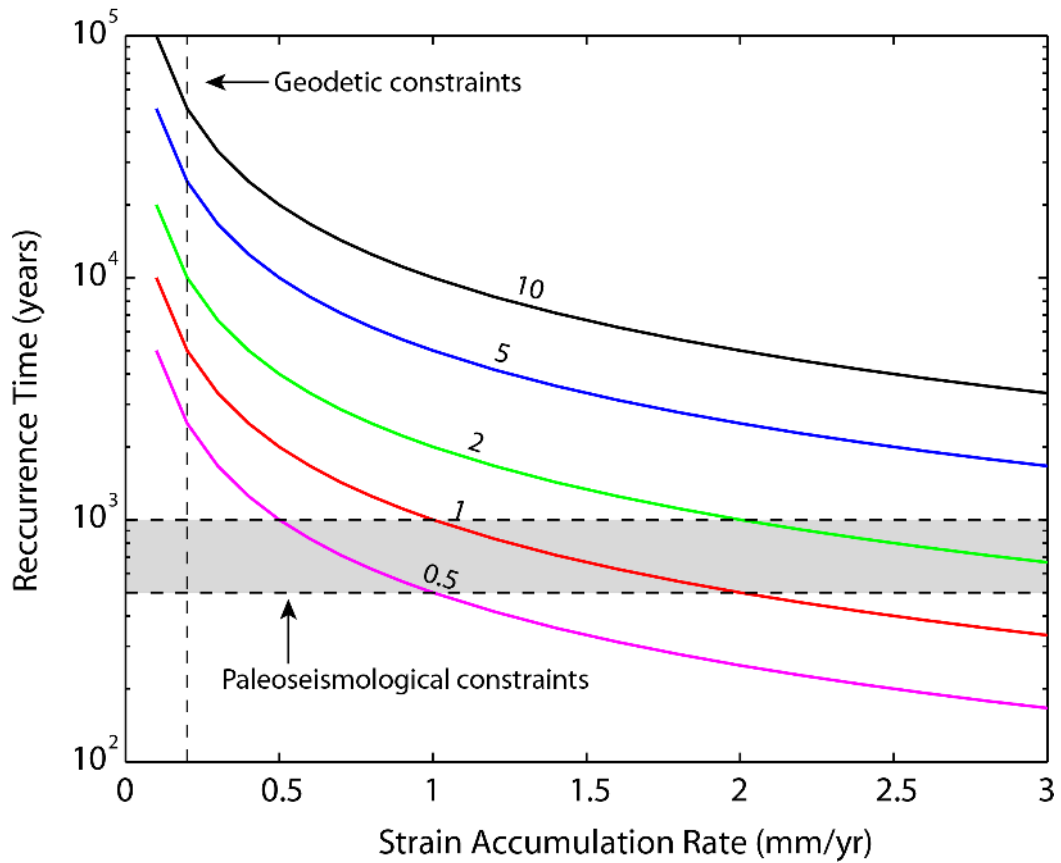
**Figure 2.** Stress changes and earthquake sequence. Top: a sequence of plate boundary earthquakes occurs as a result of tectonic loading, at a rate that may slightly vary with time, and temporal variations in fault strength [Kanamori and Brodsky, 2004]. Bottom: in SCR settings, stress accrues at very slow rates and earthquakes occur as a result of fault strength change (black line, e.g., fluid pore pressure increase at seismogenic depth) or of transient stress perturbations (blue line, e.g., hydrological or sedimentary load change).



**Figure 3.** Compilation of surface-breaking earthquake recurrence data for SCR settings updated from [Crone et al., 2003; Clark et al., 2012]. Data for the Roer Valley Graben are from Vanneste et al. [2001], Frechen et al. [2001], and van den Berg et al. [2002]

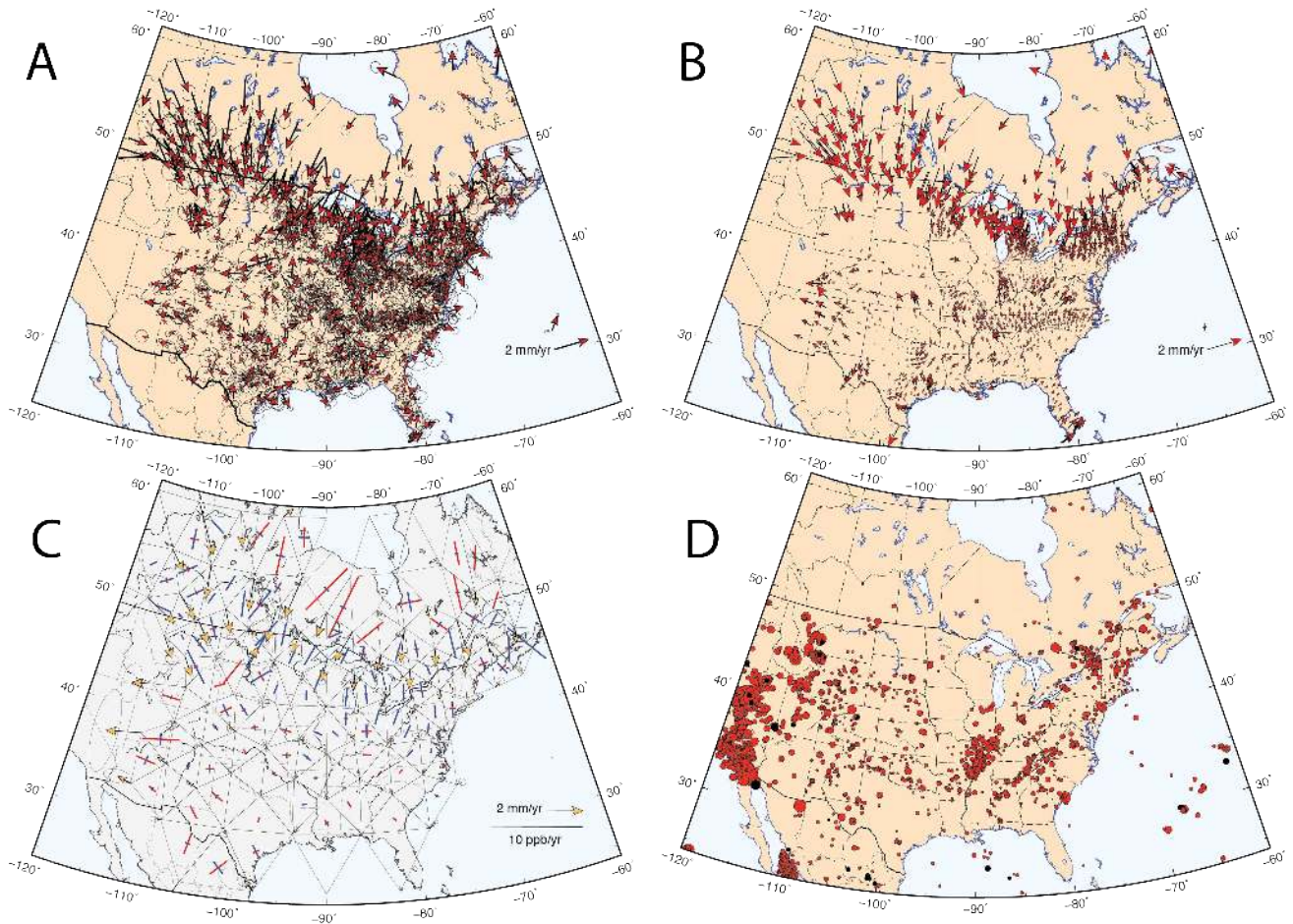


**Figure 4.** Maximum permissible deformation rates in the NMSZ as a function of publication year. Circles show continent-wide studies; squares show NMSZ studies. Red are publications claiming rates significantly different from zero; blue are upper bounds for publications claiming rates not significantly different from zero. The decrease in rates as a function of time reflects more precise site velocity estimates because of both more precise site positions and longer observation time spans.

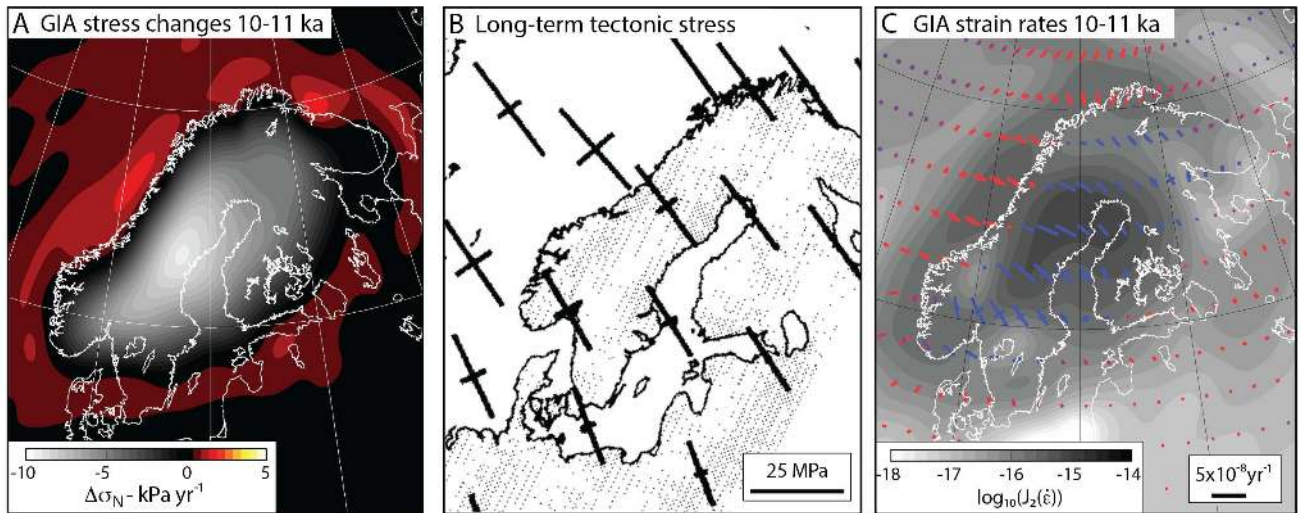


**Figure 5.** Earthquake recurrence interval as a function of slip rate across the New Madrid fault zone in steady-state, with two end-member values of coseismic slip for magnitude 7 (magenta and red curves) and magnitude 8 (blue and black curves) earthquakes. Numbers by each curve indicate the assumed coseismic slip in meters. Note that the GPS and paleoseismology domains do not overlap [Newman *et al.*, 1999; Craig and Calais, 2014].





**Figure 6.** Comparison between current geodetic horizontal strain rates and seismicity in the North American plate interior. (A) Raw GPS site velocities after the removal of a rigid plate rotation. Ellipses are 95% confidence. (B) Spatially averaged residual velocities calculated using a nearest neighbor scheme with a search radius of 800 km [Calais *et al.*, 2006]. (C) Residual velocity field interpolated to triangle vertices and corresponding principal strains. (D) Historical and instrumental seismicity, NEIC catalog (neic.usgs.gov).



**Figure 7.** Evidence for the release of long-term tectonic strain stored in continental interiors [Craig *et al.*, 2016]. (A) Rate of change of applied normal stress on a fault representative of the overall trend of the majority of known major end-glacial faults in Fennoscandia (strike =  $035^\circ$ , and dip =  $40^\circ$ ). (B). Long-term tectonic stress (principal directions) resulting to the ridge-push force exerted by the cooling and thickening oceanic lithosphere formed at the Mid-Atlantic Ridge [Gölke and Coblenz, 1996]. (C) Second invariant of the deviatoric strain rate tensor, overlain by the principle axes of the horizontal strain rate tensor (colored blue for extension and red for compression).