



Deposited via The University of Leeds.

White Rose Research Online URL for this paper:

<https://eprints.whiterose.ac.uk/id/eprint/104850/>

Version: Accepted Version

Article:

Calais, E, Camelbeeck, T, Stein, S et al. (2016) A New Paradigm for Large Earthquakes in Stable Continental Plate Interiors. *Geophysical Research Letters*, 43 (20). 10,621-10,637. ISSN: 0094-8276

<https://doi.org/10.1002/2016GL070815>

© 2016, American Geophysical Union. This is an author produced version of a paper published in *Geophysical Research Letters*. Uploaded with permission from the publisher.

Reuse

Items deposited in White Rose Research Online are protected by copyright, with all rights reserved unless indicated otherwise. They may be downloaded and/or printed for private study, or other acts as permitted by national copyright laws. The publisher or other rights holders may allow further reproduction and re-use of the full text version. This is indicated by the licence information on the White Rose Research Online record for the item.

Takedown

If you consider content in White Rose Research Online to be in breach of UK law, please notify us by emailing eprints@whiterose.ac.uk including the URL of the record and the reason for the withdrawal request.

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

E. Calais, T. Camelbeeck, S. Stein, M. Liu, and T.J. Craig

E. Calais, Department of Geosciences, Ecole normale supérieure, PSL Research University,
Paris, France

T. Camelbeeck, Royal Observatory of Belgium, Circulaire 3, 1180 Brussels, Belgium

S. Stein, Department of Earth and Planetary Sciences, Northwestern University, Evanston, IL
60208, USA

M. Liu, Department of Geological Sciences, University of Missouri, Columbia, MO 65211, USA

T.J. Craig, Institute of Geophysics and Tectonics, University of Leeds, Leeds, UK

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 ± 0.5 mm/yr over the past
115 10 ka [*Friedrich et al.*, 2003] are consistent with the 1.6 ± 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 ± 150 C.E., 900 ± 100 C.E., 300 ± 200 C.E., and 2350 ± 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 ± 269 B.C.E and
188 1620 ± 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 ~ 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⁻¹
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 χ^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 ~ 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 ~ 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.

562 **Acknowledgments.** E.C. acknowledges support from the USGS (grants 03HQGR0001,
563 07HQGR0049 and G10AP00022), from the French Investment Program SINAPS project
564 through the Commissariat à l’Energie Atomique (CEA/DASE/LDG) and the Institut de

565 Radioprotection et Sûreté Nucléaire (IRSN), and from the Yves Rocard Joint Laboratory
566 (ENS, CNRS, CEA/DASE). T.J.C. thanks the Royal Commission for the Exhibition of
567 1851 for financial support through a Research Fellowship. We thank R. Bilham and an
568 anonymous reviewer for their insightful comments, which significantly helped improve the
569 original manuscript.

References

- 570 Adams, J., R. J. Wetmiller, H. S. Hasegawa, and J. Drysdale (1991), The first surface
571 faulting from a historical intraplate earthquake in North America, *Nature Geoscience*,
572 *352*(6336), 617–619.
- 573 Ahorner, L. (1975), Present-day stress field and seismotectonic block movements along
574 major fault zones in Central Europe, *Tectonophysics*, *29*(1-4), 233–249.
- 575 Allman, A., and A. Smolka (2001), Increasing Loss Potential in Earthquake Risk-A Rein-
576 surance Perspective, in *Proceedings of the Workshop: Evaluation of the potential for*
577 *large earthquakes in regions of present-day low seismic activity in Europe*, edited by
578 T. Camelbeeck, pp. 1–4.
- 579 Argus, D. F., and R. G. Gordon (1996), Tests of the rigid-plate hypothesis and bounds
580 on intraplate deformation using geodetic data from very long baseline interferometry,
581 *Journal of Geophysical Research: Solid Earth*, *101*(B6), 13,555–13,572.
- 582 Arvidsson, R. (1996), Fennoscandian earthquakes: whole crustal rupturing related to
583 postglacial rebound, *Science*, *274*(5288), 744–736.
- 584 Beavan, J., P. Tregoning, M. Bevis, and T. Kato (2002), Motion and rigidity of the
585 Pacific Plate and implications for plate boundary deformation, *Journal of Geophysical*
586 *Research: Solid Earth*, *107*(B10), 2261.
- 587 Bent, A. L. (1994), The 1989 (MS 6.3) Ungava, Quebec, earthquake: a complex intraplate
588 event, *Bulletin of the Seismological Society of America*, *84*(4), 1075–1088.
- 589 Bettinelli, P., J.-P. Avouac, M. Flouzat, L. Bollinger, G. Ramillien, S. Rajaure, and
590 S. Sapkota (2008), Seasonal variations of seismicity and geodetic strain in the Himalaya

591 induced by surface hydrology, *Earth and Planetary Science Letters*, *266*(3-4), 332–344.

592 Bird, P., Z. Liu, and W. K. Rucker (2008), Stresses that drive the plates from below:
593 Definitions, computational path, model optimization, and error analysis, *Journal of*
594 *Geophysical Research: Solid Earth*, *113*(B11), B11,406–32.

595 Bollinger, L., F. Perrier, J. P. Avouac, S. Sapkota, U. Gautam, and D. R. Tiwari (2007),
596 Seasonal modulation of seismicity in the Himalaya of Nepal, *Geophysical Research Let-*
597 *ters*, *34*(8), L08,304–5.

598 Boyd, O. S., R. Smalley, and Y. Zeng (2015), Crustal deformation in the New Madrid
599 seismic zone and the role of postseismic processes, *Journal of Geophysical Research*
600 *(Solid Earth)*, *120*(8), 5782–5803.

601 Calais, E., and S. Stein (2009), Time-Variable Deformation in the New Madrid Seismic
602 Zone, *Science*, *323*(5920), 1442–1442.

603 Calais, E., G. Mattioli, C. DeMets, J. M. Nocquet, and S. Stein (2005), Tectonic strain
604 in plate interiors?, *Nature Geoscience*, *438*, E9–E10.

605 Calais, E., L. Dong, M. Wang, Z. Shen, and M. Vergnolle (2006), Continental deformation
606 in Asia from a combined GPS solution, *Geophysical Research Letters*, *33*(24), L24,319–
607 6.

608 Calais, E., A. M. Freed, R. Van Arsdale, and S. Stein (2010), Triggering of New Madrid
609 seismicity by late-Pleistocene erosion, *Nature Geoscience*, *466*(7306), 608–611.

610 Camelbeeck, T., and M. Meghraoui (1996), Large earthquake in northern Europe more
611 likely than once thought, *Eos, Transactions American Geophysical Union*, *77*(42), 405–
612 409.

- 613 Camelbeeck, T., and M. Meghraoui (1998), Geological and geophysical evidence for large
614 palaeo-earthquakes with surface faulting in the Roer Graben (northwest Europe), *Geo-*
615 *physical Journal International*, 132(2), 347–362.
- 616 Camelbeeck, T., K. Vanneste, P. Alexandre, K. Verbeeck, T. Petermans, P. Rosset, M. Ev-
617 eraerts, R. Warnant, and M. Van Camp (2007), Relevance of active faulting and seismic-
618 ity studies to assessments of long-term earthquake activity and maximum magnitude in
619 intraplate northwest Europe, between the Lower Rhine Embayment and the North Sea,
620 in *Special Paper 425: Continental Intraplate Earthquakes: Science, Hazard, and Policy*
621 *Issues*, pp. 193–224, Geological Society of America.
- 622 Campbell, D. L. (1978), Investigation of the stress-concentration mechanism for intraplate
623 earthquakes, *Geophysical Research Letters*, 5(6), 477–479.
- 624 Cappa, F., J. Rutqvist, and K. Yamamoto (2009), Modeling crustal deformation and
625 rupture processes related to upwelling of deep CO₂-rich fluids during the 1965–1967
626 Matsushiro earthquake swarm in Japan, *Journal of Geophysical Research: Solid Earth*,
627 114(B10), B10,304–20.
- 628 Chang, W.-L., R. B. Smith, C. M. Meertens, and R. A. Harris (2006), Contemporary
629 deformation of the Wasatch Fault, Utah, from GPS measurements with implications
630 for interseismic fault behavior and earthquake hazard: Observations and kinematic
631 analysis, *Journal of Geophysical Research: Solid Earth*, 111(B11), B11,405–19.
- 632 Chen, C., H. Gilbert, C. Andronicos, M. Hamburger, T. Larson, S. Marshak, G. Pavlis,
633 and X. Yang (2016), Shear velocity structure beneath the central United States: Im-
634 plications for the origin of the Illinois Basin and intraplate seismicity, *Geochemistry*,

- 635 *Geophysics, Geosystems, 17.*
- 636 Clark, D., A. McPherson, and R. Van Dissen (2012), Long-term behaviour of Australian
637 stable continental region (SCR) faults, *Tectonophysics, 566-567(C)*, 1–30.
- 638 Clark, D. J., and M. Leonard (2015), Do Australian intraplate faults generate character-
639 istic earthquakes?, in *Tenth Pacific Conference on Earthquake Engineering.*
- 640 Copley, A., S. Mitra, R. A. Sloan, and S. Gaonkar (2014), Active faulting in apparently
641 stable peninsular India: Rift inversion and a Holocene-age great earthquake on the
642 Tapti Fault, *Journal of Geophysical Research Solid Earth, 119(8)*, 6650–6666.
- 643 Costain, J. K., and G. A. Bollinger (2010), Review: Research Results in Hydroseismicity
644 from 1987 to 2009, *Bulletin of the Seismological Society of America, 100(5A)*, 1841–
645 1858.
- 646 Craig, T. J., and E. Calais (2014), Strain accumulation in the New Madrid and Wabash
647 Valley seismic zones from 14 years of continuous GPS observation, *Journal of Geophys-
648 ical Research: Solid Earth, 119(12)*, 9110–9129.
- 649 Craig, T. J., E. Calais, L. Fleitout, L. Bollinger, and O. Scotti (2016), Evidence for the
650 release of long-term tectonic strain stored in continental interiors through intraplate
651 earthquakes, *Geophysical Research Letters, 43.*
- 652 Crone, A. J., and K. V. Luza (1990), Style and timing of Holocene surface faulting on the
653 Meers fault, southwestern Oklahoma, *Geological Society of America Bulletin, 102(1)*,
654 1–17.
- 655 Crone, A. J., M. N. Machette, and J. R. Bowman (1992), Geological investigations of the
656 1988 Tennant Creek, Australia, earthquakes; implications for paleoseismicity in stable

- 657 continental regions, *U.S. Geological Survey Bulletin*, 2032-A, 46 p.
- 658 Crone, A. J., M. N. Machette, and J. R. Bowman (1997a), Episodic nature of earthquake
659 activity in stable continental regions revealed by palaeoseismicity studies of Australian
660 and North American quaternary faults, *Australian Journal of Earth Sciences*, 44(2),
661 203–214.
- 662 Crone, A. J., N. M. Machette, L. A. Bradley, and S. A. Mahan (1997b), Late Quaternary
663 surface faulting on the Cheraw fault, southeastern Colorado, U.S. Geological Survey.
- 664 Crone, A. J., P. M. De Martini, M. N. Machette, K. Okumura, and J. R. Prescott (2003),
665 Paleoseismicity of Two Historically Quiescent Faults in Australia: Implications for Fault
666 Behavior in Stable Continental Regions, *Bulletin of the Seismological Society of Amer-*
667 *ica*, 93(5), 1913–1934.
- 668 Dèzes, P., S. M. Schmid, and P. A. Ziegler (2004), Evolution of the European Cenozoic Rift
669 System: interaction of the Alpine and Pyrenean orogens with their foreland lithosphere,
670 *Tectonophysics*, 389(1-2), 1–33.
- 671 Dixon, T. H., A. Mao, and S. Stein (1996), How rigid is the stable interior of the North
672 American Plate?, *Geophysical Research Letters*, 23(21), 3035–3038.
- 673 Ellsworth, W. L., A. L. Llenos, A. F. McGarr, A. J. Michael, J. L. Rubinstein, C. S.
674 Mueller, M. D. Petersen, and E. Calais (2015), Increasing seismicity in the U. S. mid-
675 continent: Implications for earthquake hazard, *The Leading Edge*, 34(6), 618–626.
- 676 Feldl, N., and R. Bilham (2006), Great Himalayan earthquakes and the Tibetan plateau,
677 *Nature Geoscience*, 444(7116), 165–170.

- 678 Field, E. H., et al. (2014), Uniform California Earthquake Rupture Forecast, Version
679 3 (UCERF3)—The Time-Independent Model, *Bulletin of the Seismological Society of*
680 *America*, *104*(3), 1122–1180.
- 681 Fleitout, L., and C. Froidevaux (1983), Tectonic stresses in the lithosphere, *Tectonics*,
682 *2*(3), 315–324.
- 683 Frechen, M., K. Vanneste, K. Verbeeck, E. Paulissen, and T. Camelbeeck (2001), The
684 deposition history of the coversands along the Bree Fault Escarpment, NE Belgium,
685 *Geologie en Mijnbouw*, *80*(3-4), 171–185.
- 686 Friedrich, A. M., B. P. Wernicke, N. A. Niemi, R. A. Bennett, and J. L. Davis (2003),
687 Comparison of geodetic and geologic data from the Wasatch region, Utah, and impli-
688 cations for the spectral character of Earth deformation at periods of 10 to 10 million
689 years, *Journal of Geophysical Research: Solid Earth*, *108*(B4), 2199.
- 690 Fuhrmann, T., B. Heck, A. Knöpfler, F. Masson, M. Mayer, P. Ulrich, M. Westerhaus,
691 and K. Zippelt (2013), Recent surface displacements in the Upper Rhine Graben —
692 Preliminary results from geodetic networks, *Tectonophysics*, *602*(C), 300–315.
- 693 Garcia-Moreno, D., et al. (2015), Fault activity in the epicentral area of the 1580 Dover
694 Strait (Pas-de-Calais) earthquake (northwestern Europe), *Geophysical Journal Interna-*
695 *tional*, *201*(2), 528–542.
- 696 Gölke, M., and D. Coblenz (1996), Origins of the European regional stress field, *Tectono-*
697 *physics*, *266*(1-4), 11–24.
- 698 González, P. J., K. F. Tiampo, M. Palano, F. Cannavò, and J. Fernández (2012), The
699 2011 Lorca earthquake slip distribution controlled by groundwater crustal unloading,

- 700 *Nature Geoscience*, 5(11), 821–825.
- 701 Got, J.-L., V. Monteiller, J. Guilbert, D. Marsan, Y. Cansi, C. Maillard, and J.-P. Santoire
702 (2011), Strain localization and fluid migration from earthquake relocation and seismicity
703 analysis in the western Vosges (France), *Geophysical Journal International*, 185(1), 365–
704 384.
- 705 Greene, M., C. Godavitarne, and F. Kringold (2000), Overview of the Maharashtra, India
706 emergency earthquake rehabilitation program, *12th World Conference on Earthquake*
707 *Engineering*.
- 708 Grollmund, B., and M. D. Zoback (2001), Did deglaciation trigger intraplate seismicity
709 in the New Madrid seismic zone?, *Geology*, 29(2), 175–178.
- 710 Gupta, H. K. (1985), The present status of reservoir induced seismicity investigations
711 with special emphasis on Koyna earthquakes, *Tectonophysics*, 118(3-4), 257–279.
- 712 Hackl, M., R. Malservisi, U. Hugentobler, and R. Wonnacott (2011), Estimation of velocity
713 uncertainties from GPS time series: Examples from the analysis of the South African
714 TrigNet network, *Journal of Geophysical Research: Solid Earth*, 116(B11), B11,404–12.
- 715 Hainzl, S., T. Kraft, J. Wassermann, H. Igel, and E. Schmedes (2006), Evidence for
716 rainfall-triggered earthquake activity, *Geophysical Research Letters*, 33(19), L19,303–5.
- 717 Hanks, T. C., and A. C. Johnston (1992), Common features of the excitation and propa-
718 gation of strong ground motion for North American earthquakes, *Bulletin of the Seis-*
719 *mological Society of America*, 82(1), 1–23.
- 720 Heidbach, O., J. Reinecker, M. Tingay, and B. Müller (2007), Plate boundary forces are
721 not enough: second-and third-order stress patterns highlighted in the World Stress Map

- 722 database, *Geophysical Research Letters*, *26*, TC6014.
- 723 Heidbach, O., M. Tingay, A. Barth, J. Reinecker, D. Kurfeß, and B. Müller (2010), Global
724 crustal stress pattern based on the World Stress Map database release 2008, *Tectono-*
725 *physics*, *482*(1-4), 3–15.
- 726 Heki, K. (2003), Snow load and seasonal variation of earthquake occurrence in Japan,
727 *Earth and Planetary Science Letters*, *207*(1-4), 159–164.
- 728 Herrmann, R. B., H. Benz, and C. J. Ammon (2011), Monitoring the Earthquake Source
729 Process in North America, *Bulletin of the Seismological Society of America*, *101*(6),
730 2609–2625.
- 731 Hetzel, R., and A. Hampel (2005), Slip rate variations on normal faults during
732 glacial–interglacial changes in surface loads, *Nature Geoscience*, *435*(7038), 81–84.
- 733 Holbrook, J., W. J. Autin, T. M. RITTENOUR, S. Marshak, and R. J. Goble (2006),
734 Stratigraphic evidence for millennial-scale temporal clustering of earthquakes on a
735 continental-interior fault: Holocene Mississippi River floodplain deposits, New Madrid
736 seismic zone, USA, *Tectonophysics*, *420*(3-4), 431–454.
- 737 Hough, S. E., J. G. Armbruster, L. Seeber, and J. F. Hough (2000), On the Modified
738 Mercalli intensities and magnitudes of the 1811–1812 New Madrid earthquakes, *Journal*
739 *of Geophysical Research: Solid Earth (1978–2012)*, *105*(B10), 23,839–23,864.
- 740 Hurd, O., and M. D. Zoback (2012a), Regional Stress Orientations and Slip Compatibility
741 of Earthquake Focal Planes in the New Madrid Seismic Zone, *Seismological Research*
742 *Letters*, *83*(4), 672–679.

- 743 Hurd, O., and M. D. Zoback (2012b), Intraplate earthquakes, regional stress and fault
744 mechanics in the Central and Eastern U.S. and Southeastern Canada, *Tectonophysics*,
745 *581(C)*, 182–192.
- 746 Illies, J. H., H. Baumann, and B. Hoffers (1981), Stress pattern and strain release in the
747 Alpine foreland, *Tectonophysics*, *71*, 157–172.
- 748 Jakobsson, M., et al. (2014), Major earthquake at the Pleistocene-Holocene transition in
749 Lake Vattern, southern Sweden, *Geology*, *42(5)*, 379–382.
- 750 Johansson, J. M., J. L. Davis, and H. G. Scherneck (2002), Continuous GPS measurements
751 of postglacial adjustment in Fennoscandia 1. Geodetic results, *Journal of Geophysical*
752 *Research Solid Earth*, *107(B8)*.
- 753 Johnson, G. A., S. P. Horton, M. Withers, and R. Cox (2014), Earthquake Focal Mecha-
754 nisms in the New Madrid Seismic Zone, *Seismological Research Letters*, *85(2)*, 257–267.
- 755 Johnston, A. C. (1989), The Seismicity of ‘Stable Continental Interiors’, in *Earthquakes*
756 *at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound*, pp. 299–327,
757 Springer Netherlands, Dordrecht.
- 758 Johnston, A. C. (1996), Seismic moment assessment of earthquakes in stable continental
759 regions—II. Historical seismicity, *Geophysical Journal International*, *125*, 639–678.
- 760 Johnston, A. C., L. R. Kanter, K. J. Coppersmith, and C. A. Cornell (1994), The earth-
761 quakes of stable continental regions , *Tech. rep.*, Palo Alto.
- 762 Joskow, P. L., and J. E. Parsons (2012), The Future of Nuclear Power After Fukushima,
763 *MIT Center for Energy and Environmental Policy Research*, p. 29 p.

- 764 Kanamori, H., and E. E. Brodsky (2004), The physics of earthquakes, *Reports on Progress*
765 *in Physics*, 67(8), 1429–1496.
- 766 Kenner, S. J., and P. Segall (2000), A mechanical model for intraplate earthquakes: Ap-
767 plication to the New Madrid seismic zone, *Science*, 289, 2329–2331.
- 768 Keranen, K. M., H. M. Savage, G. A. Abers, and E. S. Cochran (2013), Potentially induced
769 earthquakes in Oklahoma, USA: Links between wastewater injection and the 2011 Mw
770 5.7 earthquake sequence, *Geology*, 41(6), 699–702.
- 771 Keranen, K. M., M. Weingarten, G. A. Abers, B. A. Bekins, and S. Ge (2014), Sharp
772 increase in central Oklahoma seismicity since 2008 induced by massive wastewater in-
773 jection, *Science*, 345(6195), 448–451.
- 774 Kierulf, H. P., H. Steffen, and M. Simpson (2014), A GPS velocity field for Fennoscandia
775 and a consistent comparison to glacial isostatic adjustment models, *Journal of Geo-*
776 *physical Research Solid Earth*, 119(8), 6613–6629.
- 777 Kuhn, T. S. (1962), The structure of scientific revolutions, 2(2), 1–174.
- 778 Lagerbäck, R. (1992), Dating of Late Quaternary faulting in northern Sweden, *Journal of*
779 *the Geological Society*, 149(2), 285–291.
- 780 Lagerbäck, R., and M. Sundh (2008), Early Holocene faulting and paleoseismicity in
781 northern Sweden, Geological Survey of Sweden, Uppsala.
- 782 Lambeck, K., and A. Purcell (2003), Glacial Rebound and Crustal Stress in Finland, *Tech.*
783 *rep.*, Oliluoto, Finland.
- 784 Leclère, H., G. Daniel, O. Fabbri, F. Cappa, and F. Thouvenot (2013), Tracking fluid
785 pressure buildup from focal mechanisms during the 2003-2004 Ubaye seismic swarm,

- 786 France, *Journal of Geophysical Research: Solid Earth*, 118(8), 4461–4476.
- 787 Lecocq, T., T. Petermans, P. Alexandre, and T. Camelbeeck (2008), Earthquake reloca-
788 tion in the Ardenne (Belgium): identification of active structures in intraplate context,
789 in *Seismic Risk - Earthquakes in North-Western Europe*.
- 790 Lidberg, M., J. M. Johansson, H.-G. Scherneck, and G. A. Milne (2010), Recent re-
791 sults based on continuous GPS observations of the GIA process in Fennoscandia from
792 BIFROST, *Journal of Geodynamics*, 50(1), 8–18.
- 793 Lindblom, E., B. Lund, A. Tryggvason, M. Uski, R. Bodvarsson, C. Juhlin, and R. Roberts
794 (2015), Microearthquakes illuminate the deep structure of the endglacial Parvie fault,
795 northern Sweden, *Geophysical Journal International*, 201(3), 1704–1716.
- 796 Liu, L., M. D. Zoback, and P. Segall (1992), Rapid Intraplate Strain Accumulation in the
797 New Madrid Seismic Zone, *Science*, 257(5077), 1666–1669.
- 798 Liu, M., S. Stein, and H. Wang (2011), 2000 years of migrating earthquakes in North
799 China: How earthquakes in midcontinents differ from those at plate boundaries, *Litho-
800 sphere*, 3(2), 128–132.
- 801 Liu, M., G. Luo, H. Wang, and S. Stein (2014), Long aftershock sequences in North China
802 and Central US: implications for hazard assessment in mid-continent, *Earthquake Sci-
803 ence*, 27(1), 27–35.
- 804 Long, L. T. (1988), A Model for Major Intraplate Continental Earthquakes, *Seismological
805 Research Letters*, 59(4), 273–278.
- 806 Luttrell, K., and D. Sandwell (2010), Ocean loading effects on stress at near shore plate
807 boundary fault systems, *Journal of Geophysical Research: Solid Earth*, 115, B08,411–14.

- 808 Luttrell, K., D. Sandwell, B. Smith-Konter, B. Bills, and Y. Bock (2007), Modulation
809 of the earthquake cycle at the southern San Andreas fault by lake loading, *Journal of*
810 *Geophysical Research: Solid Earth*, *112*, B08,411–15.
- 811 Machette, M. N. (2000), Active, capable, and potentially active faults - a paleoseismic
812 perspective, *Journal of Geodynamics*, *29*(3-5), 387–392.
- 813 Machette, M. N., A. J. Crone, and J. R. Bowman (1993), Geologic investigations of the
814 1986 Marryat Creek, Australia, earthquake; implications for paleoseismicity in stable
815 continental regions, *U.S. Geological Survey Bulletin*, *2032-B*, 27 p.
- 816 Madole, R. F. (1988), Stratigraphic evidence of Holocene faulting in the mid-continent:
817 The Meers fault, southwestern Oklahoma, *Geological Society of America Bulletin*,
818 *100*(3), 392–401.
- 819 Maurer, R., and K. Oblitas (2001), Gujarat Earthquake Recovery Program Assessment
820 Report, *Tech. rep.*
- 821 Mazzotti, S., T. S. James, J. Henton, and J. Adams (2005), GPS crustal strain, postglacial
822 rebound, and seismic hazard in eastern North America: The Saint Lawrence valley
823 example, *Journal of Geophysical Research: Solid Earth*, *110*(B11), B11,301–16.
- 824 McCalpin, J. P., and S. P. Nishenko (1996), Holocene paleoseismicity, temporal clustering,
825 and probabilities of future large ($M \geq 7$) earthquakes on the Wasatch fault zone, Utah,
826 *Journal of Geophysical Research: Solid Earth (1978–2012)*, *101*(B3), 6233–6253.
- 827 McNamara, D. E., H. M. Benz, R. B. Herrmann, E. A. Bergman, P. Earle, A. Holland,
828 R. Baldwin, and A. Gassner (2015), Earthquake hypocenters and focal mechanisms in
829 central Oklahoma reveal a complex system of reactivated subsurface strike-slip faulting,

- 830 *Geophysical Research Letters*, 42(8), 2742–2749.
- 831 Mörner, N.-A. (2005), An interpretation and catalogue of paleoseismicity in Sweden,
832 *Tectonophysics*, 408(1-4), 265–307.
- 833 Mueller, K., S. E. Hough, and R. Bilham (2004), Analysing the 1811–1812 New Madrid
834 earthquakes with recent instrumentally recorded aftershocks, *Nature Geoscience*, 49,
835 284–288.
- 836 Muir-Wood, R. (1989), Extraordinary Deglaciation Reverse Faulting in Northern
837 Fennoscandia, in *Earthquakes at North-Atlantic Passive Margins: Neotectonics and*
838 *Postglacial Rebound*, pp. 141–173, Springer Netherlands, Dordrecht.
- 839 Müller, B., M. L. Zoback, K. Fuchs, L. Mastin, S. Gregersen, N. Pavoni, O. Stephansson,
840 and C. Ljunggren (1992), Regional patterns of tectonic stress in Europe, *Journal of*
841 *Geophysical Research: Solid Earth (1978–2012)*, 97(B8), 11,783–11,803.
- 842 Newman, A., S. Stein, J. Weber, J. Engeln, A. Mao, and T. Dixon (1999), Slow deforma-
843 tion and lower seismic hazard at the New Madrid seismic zone, *Science*, 284, 619–621.
- 844 Nocquet, J. M. (2012), Present-day kinematics of the Mediterranean: A comprehensive
845 overview of GPS results, *Tectonophysics*, 578, 220–242.
- 846 Nocquet, J. M., and E. Calais (2003), Crustal velocity field of western Europe from
847 permanent GPS array solutions, 1996–2001, *Geophysical Journal International*, 154(1),
848 72–88.
- 849 Nocquet, J. M., and E. Calais (2004), Geodetic Measurements of Crustal Deformation in
850 the Western Mediterranean and Europe, *Pure and Applied Geophysics*, 161(3), 661–681.

- 851 Nocquet, J. M., E. Calais, and B. Parsons (2005), Geodetic constraints on glacial isostatic
852 adjustment in Europe, *Geophysical Research Letters*, *32*, L06,308.
- 853 Nuttli, O. W. (1973), The Mississippi Valley earthquakes of 1811 and 1812: Intensities,
854 ground motion and magnitudes, *Bulletin of the Seismological Society of America*, *63*(1),
855 227–248.
- 856 Olesen, O., L. H. Blikra, and A. Braathen (2004), Neotectonic deformation in Norway
857 and its implications: a review, *Norwegian Journal of Geology*, *84*, 3–34.
- 858 Olesen, O., H. Bungum, J. Dehls, C. Lindholm, and C. Pascal (2013), Neotectonics,
859 seismicity and contemporary stress field in Norway—mechanisms and implications, in
860 *Quaternary Geology of Norway, Geological Survey of Norway Special Publication*, edited
861 by L. Olsen, O. Fredin, and O. Olesen, pp. 145–174, NGU Special Publication.
- 862 Page, M. T., and S. E. Hough (2014), The New Madrid Seismic Zone: Not Dead Yet,
863 *Science*, *343*(6172), 762–764.
- 864 Pascal, C., D. Roberts, and R. H. Gabrielsen (2010), Tectonic significance of present-day
865 stress relief phenomena in formerly glaciated regions, *Journal of the Geological Society*,
866 *167*(2), 363–371.
- 867 Pechmann, J. C., and W. J. Arabasz (1995), The problem of the random earthquake
868 in seismic hazard analysis: Wasatch Front region, Utah, *Utah Geological Association*
869 *Publication*, *24*, 77–93.
- 870 Peltier, W. R., D. F. Argus, and R. Drummond (2015), Space geodesy constrains ice
871 age terminal deglaciation: The global ICE-6GC (VM5a) model, *Journal of Geophysical*
872 *Research: Solid Earth*, *120*(1), 450–487.

- 873 Petersen, M. D., M. P. Moschetti, and P. M. Powers (2015), The 2014 United States
874 National Seismic Hazard Model, *Earthquake Spectra*, *31*, S1–S30.
- 875 Pollitz, F. F., L. Kellogg, and R. Bürgmann (2001), Sinking mafic body in a reactivated
876 lower crust: a mechanism for stress concentration at the New Madrid seismic zone,
877 *Bulletin of the Seismological Society of America*, *91*, 1882—1897.
- 878 Raleigh, C. B., J. H. Healy, and J. D. Bredehoeft (1972), *Faulting and Crustal Stress*
879 *at Rangely, Colorado, Geophysical Monograph Series*, vol. 16, American Geophysical
880 Union, Washington, D. C.
- 881 Rigo, A., et al. (2015), Present-day deformation of the Pyrenees revealed by GPS surveying
882 and earthquake focal mechanisms until 2011, *Geophysical Journal International*, *201*(2),
883 947–964.
- 884 Rothé, J. (1968), Fill a lake, start an earthquake, *Ekistics*, *26*(156), 432–435.
- 885 Saria, E., E. Calais, Z. Altamimi, P. Willis, and H. Farah (2013), A new velocity field for
886 Africa from combined GPS and DORIS space geodetic Solutions: Contribution to the
887 definition of the African reference frame (AFREF), *Journal of Geophysical Research:*
888 *Solid Earth*, *118*(4), 1677–1697.
- 889 Sbar, M. L., and L. B. Sykes (1977), Seismicity and lithospheric stress in New York and
890 adjacent areas, *Journal of Geophysical Research: Solid Earth*, *82*(36), 5771–5786.
- 891 Schrank, C. E., F. Fousseis, A. Karrech, and K. Regenauer-Lieb (2012), Thermal-elastic
892 stresses and the criticality of the continental crust, *Geochemistry, Geophysics, Geosys-*
893 *tems*, *13*(9), 1–21.

- 894 Schulte, S. M., and W. D. Mooney (2005), An updated global earthquake catalogue for
895 stable continental regions: reassessing the correlation with ancient rifts, *Geophysical*
896 *Journal International*, 161(3), 707–721.
- 897 Schwartz, D. P., and K. J. Coppersmith (1984), Fault behavior and characteristic earth-
898 quakes: Examples from the Wasatch and San Andreas Fault Zones, *Journal of Geo-*
899 *physical Research: Solid Earth*, 89(B7), 5681–5698.
- 900 Sella, G. F., S. Stein, T. H. Dixon, M. Craymer, T. S. James, S. Mazzotti, and R. K.
901 Dokka (2007), Observation of glacial isostatic adjustment in “stable” North America
902 with GPS, *Geophysical Research Letters*, 34(2), L02,306–6.
- 903 Sibson, R. H. (1990), Conditions for fault-valve behavior, Deformation mechanisms, Rhe-
904 ology and Tectonics RJ Knipe, EH Rutter, 15–28, in *Deformation Mechanisms, Rheology*
905 *and Tectonics*, edited by R. J. Knipe and E. H. Rutter, pp. 15–28, Geol. Soc. of London.
- 906 Simpson, D. W. (1976), Seismicity changes associated with reservoir loading, *Engineering*
907 *Geology*, 10(2-4), 123–150.
- 908 Simpson, D. W., W. S. Leith, and C. H. Scholz (1988), Two types of reservoir-induced
909 seismicity, *Bulletin of the Seismological Society of America*, 78(6), 2025–2040.
- 910 Smalley, R., M. A. Ellis, J. Paul, and R. B. Van Arsdale (2005), Space geodetic evidence
911 for rapid strain rates in the New Madrid seismic zone of central USA, *Nature Geoscience*,
912 435(7045), 1088–1090.
- 913 Smit, L., A. Fagereng, B. Braeuer, and J. Stankiewicz (2015), Microseismic Activity and
914 Basement Controls on an Active Intraplate Strike-Slip Fault, Ceres-Tulbagh, South
915 Africa, *Bulletin of the Seismological Society of America*, 105, 1–8.

- 916 Smith, C. A., M. Sundh, and H. Mikko (2014), Surficial geology indicates early Holocene
917 faulting and seismicity, central Sweden, *International Journal of Earth Sciences*, *103*(6),
918 1711–1724.
- 919 Snay, R., J. F. Ni, and H. C. Neugebauer (1994), Geodetically Derived Strain Across
920 the Northern New Madrid Seismic Zone, *U.S. Geological Survey Professional Paper*,
921 *1538-F-G*, F1–F6.
- 922 Spencer, B. F. J., et al. (2008), Impact of Earthquakes on the Central USA, *Tech. Rep.*
923 *08-02*, Mid-America Earthquake Center, Urbana, IL.
- 924 Špičák, A., and J. Horálek (2001), Possible role of fluids in the process of earthquake
925 swarm generation in the West Bohemia/Vogtland seismoactive region, *Tectonophysics*,
926 *336*(1-4), 151–161.
- 927 Steffen, R., P. Wu, H. Steffen, and D. W. Eaton (2014), The effect of earth rheology
928 and ice-sheet size on fault slip and magnitude of postglacial earthquakes, *Earth and*
929 *Planetary Science Letters*, *388*(C), 71–80.
- 930 Stein, R. S., G. C. P. King, and J. Lin (1992), Change in Failure Stress on the Southern
931 San Andreas Fault System Caused by the 1992 Magnitude = 7.4 Landers Earthquake,
932 *Science*, *258*(5086), 1328–1332.
- 933 Stein, R. S., J. H. Dieterich, and A. A. Barka (1996), Role of stress triggering in earthquake
934 migration on the North Anatolian Fault, *Physics and Chemistry of the Earth*, *21*(4),
935 225–230.
- 936 Stein, S., and M. Liu (2009), Long aftershock sequences within continents and implications
937 for earthquake hazard assessment, *Nature Geoscience*, *462*(7269), 87–89.

- 938 Stein, S., S. Cloetingh, N. H. Sleep, and R. Wortel (1989), Passive Margin Earthquakes,
939 Stresses and Rheology, in *Earthquakes at North-Atlantic Passive Margins: Neotectonics*
940 *and Postglacial Rebound*, pp. 231–259, Springer Netherlands, Dordrecht.
- 941 Stein, S., A. Friedrich, and A. Newman (2005), Dependence of Possible Characteristic
942 Earthquakes on Spatial Sampling: Illustration for the Wasatch Seismic Zone, Utah,
943 *Seismological Research Letters*, *76*(4), 432–436.
- 944 Stein, S., M. Liu, E. Calais, and Q. Li (2009), Mid-Continent Earthquakes as a Complex
945 System, *Seismological Research Letters*, *80*(4), 551–553.
- 946 Sykes, L. R. (1978), Intraplate seismicity, reactivation of preexisting zones of weakness, al-
947 kaline magmatism, and other tectonism postdating continental fragmentation, *Reviews*
948 *of Geophysics*, *16*(4), 621–688.
- 949 Sykes, L. R., and M. L. Sbar (1973), Intraplate earthquakes, lithospheric stresses and the
950 driving mechanism of plate tectonics, *Nature Geoscience*, *245*(5424), 298–302.
- 951 Talwani, P. (1999), Fault geometry and earthquakes in continental interiors, *Tectono-*
952 *physics*, *305*(1-3), 371–379.
- 953 Tesauro, M., M. K. Kaban, and W. D. Mooney (2015), Variations of the lithospheric
954 strength and elastic thickness in North America, *Geochemistry, Geophysics, Geosys-*
955 *tems*, *16*(7), 2197–2220.
- 956 Tong, X., B. Konter-Smith, and D. T. Sandwell (2014), Is there a discrepancy between
957 geological and geodetic slip rates along the San Andreas Fault System?, *Journal of*
958 *Geophysical Research (Solid Earth)*, *119*, 2518–2538.

- 959 Townend, J., and M. D. Zoback (2000), How faulting keeps the crust strong, *Geology*,
960 *28*(5), 399–402.
- 961 Tregoning, P., R. Burgette, S. C. McClusky, S. Lejeune, C. S. Watson, and H. McQueen
962 (2013), A decade of horizontal deformation from great earthquakes, *Journal of Geo-*
963 *physical Research: Solid Earth*, *118*(5), 2371–2381.
- 964 Turpeinen, H., A. Hampel, T. Karow, and G. Maniatis (2008), Effect of ice sheet growth
965 and melting on the slip evolution of thrust faults, *Earth and Planetary Science Letters*,
966 *269*(1-2), 230–241.
- 967 Tuttle, M. P., E. S. Schweig, J. D. Sims, R. H. Lafferty, L. W. Wolf, and M. L. Haynes
968 (2002), The Earthquake Potential of the New Madrid Seismic Zone, *Bulletin of the*
969 *Seismological Society of America*, *92*(6), 2080–2089.
- 970 Van Arsdale, R. (2000), Displacement history and slip rate on the Reelfoot fault of the
971 New Madrid seismic zone, *Engineering Geology*, *55*(4), 219–226.
- 972 van den Berg, M., K. Vanneste, B. Dost, A. Lokhorst, M. van Eijk, and K. Verbeeck
973 (2002), Paleoseismic investigations along the Peel Boundary Fault: geological setting,
974 site selection and trenching results, *Netherlands Journal of Geosciences*, *81*(01), 39–60.
- 975 Vanneste, K., K. Verbeeck, T. Camelbeeck, E. Paulissen, M. Meghraoui, F. Renardy,
976 D. Jongmans, and M. Frechen (2001), Surface-rupturing history of the Bree fault scarp,
977 Roer Valley graben: Evidence for six events since the late Pleistocene, *Journal of Seis-*
978 *mology*, *5*(3), 329–359.
- 979 Vanneste, K., T. Camelbeeck, and K. Verbeeck (2013), A Model of Composite Seismic
980 Sources for the Lower Rhine Graben, Northwest Europe, *Bulletin of the Seismological*

- 981 *Society of America*, 103(2A), 984–1007.
- 982 Viola, G., M. Andreoli, Z. Ben-Avraham, I. Stengel, and M. Reshef (2005), Offshore mud
983 volcanoes and onland faulting in southwestern Africa: neotectonic implications and
984 constraints on the regional stress field, *Earth and Planetary Science Letters*, 231(1-2),
985 147–160.
- 986 Wang, C.-Y., and M. Manga (2009), Earthquakes Influenced by Water, in *Earthquakes*
987 *and Water*, pp. 125–139, Springer Berlin Heidelberg, Berlin, Heidelberg.
- 988 Weise, S. M., K. Bräuer, H. Kämpf, G. Strauch, and U. Koch (2001), Transport of mantle
989 volatiles through the crust traced by seismically released fluids: a natural experiment in
990 the earthquake swarm area Vogtland/NW Bohemia, Central Europe, *Tectonophysics*,
991 336(1-4), 137–150.
- 992 White, S., H. Stollhofen, I. G. Stanistreet, and V. Lorenz (2009), Pleistocene to Recent
993 rejuvenation of the Hebron Fault, SW Namibia, *Geological Society, London, Special*
994 *Publications*, 316(1), 293–317.
- 995 Wolin, E., S. Stein, F. Pazzaglia, A. Meltzer, A. Kafka, and C. Berti (2012), Mineral,
996 Virginia, earthquake illustrates seismicity of a passive-aggressive margin, *Geophysical*
997 *Research Letters*, 39(2), L02,305.
- 998 Wouters, B., J. A. Bonin, D. P. Chambers, R. E. M. Riva, I. Sasgen, and J. Wahr (2014),
999 GRACE, time-varying gravity, Earth system dynamics and climate change, *Reports on*
1000 *Progress in Physics*, 77(11), 116,801–42.
- 1001 Wu, P., and P. Johnston (2000), Can deglaciation trigger earthquakes in N. America?,
1002 *Geophysical Research Letters*, 29(9), 1321–1326.

- 1003 Wu, P., P. Johnston, and K. Lambeck (1999), Postglacial rebound and fault instability in
1004 Fennoscandia, *Geophysical Journal International*, *139*(3), 657–670.
- 1005 Zhan, Y., G. Hou, T. Kusky, and P. M. Gregg (2016), Stress development in heteroge-
1006 netic lithosphere: Insights into earthquake processes in the New Madrid Seismic Zone,
1007 *Tectonophysics*, *671*, 56–62.
- 1008 Zhao, B., Y. Huang, C. Zhang, W. Wang, K. Tan, and R. Du (2015), Crustal defor-
1009 mation on the Chinese mainland during 1998–2014 based on GPS data, *Geodesy and*
1010 *Geodynamics*, *6*(1), 7–15.
- 1011 Zoback, M. D., and M. L. Zoback (1981), State of Stress and Intraplate Earthquakes in
1012 the United States, *Science*, *213*(4503), 96–104.
- 1013 Zoback, M. D., W. H. Prescott, and S. W. Krueger (1985), Evidence for lower crustal
1014 ductile strain localization in southern New York, *Nature Geoscience*, *317*(6039), 705–
1015 707.
- 1016 Zoback, M. D., et al. (1993), Upper-crustal strength inferred from stress measurements
1017 to 6 km depth in the KTB borehole, *Nature Geoscience*, *365*(6447), 633–635.
- 1018 Zoback, M. L., and R. M. Richardson (1996), Stress perturbation associated with the
1019 Amazonas and other ancient continental rifts, *Journal of Geophysical Research: Solid*
1020 *Earth (1978–2012)*, *101*(B3), 5459–5475.
- 1021 Zoback, M. L., and M. D. Zoback (1989), Tectonic stress field of the continental United
1022 States, *Geological Society of America Memoirs*, *172*, 523–540.

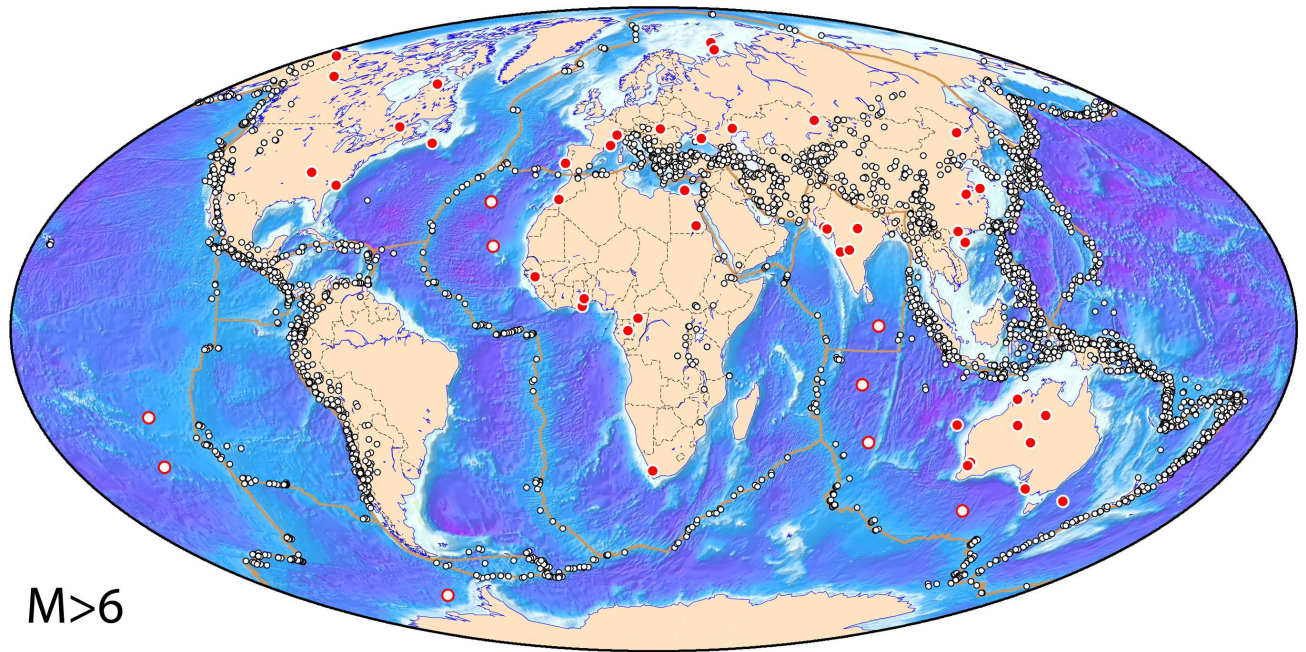


Figure 1. Worldwide seismic activity (<http://earthquake.usgs.gov/>). Large circles show $M>6$ intraplate earthquakes: red for stable continental regions, white for stable oceanic regions.

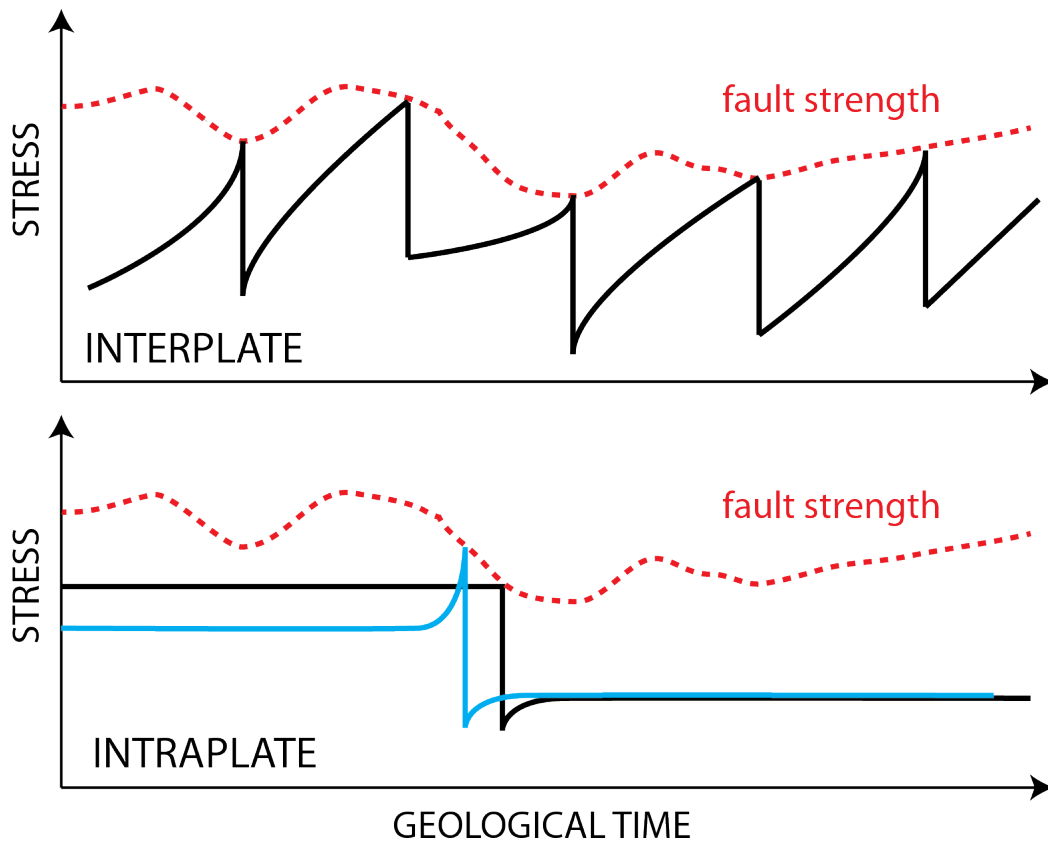


Figure 2. Stress changes and earthquake sequence. Top: a sequence of plate boundary earthquakes occurs as a result of tectonic loading, at a rate that may slightly vary with time, and temporal variations in fault strength [Kanamori and Brodsky, 2004]. Bottom: in SCR settings, stress accrues at very slow 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).

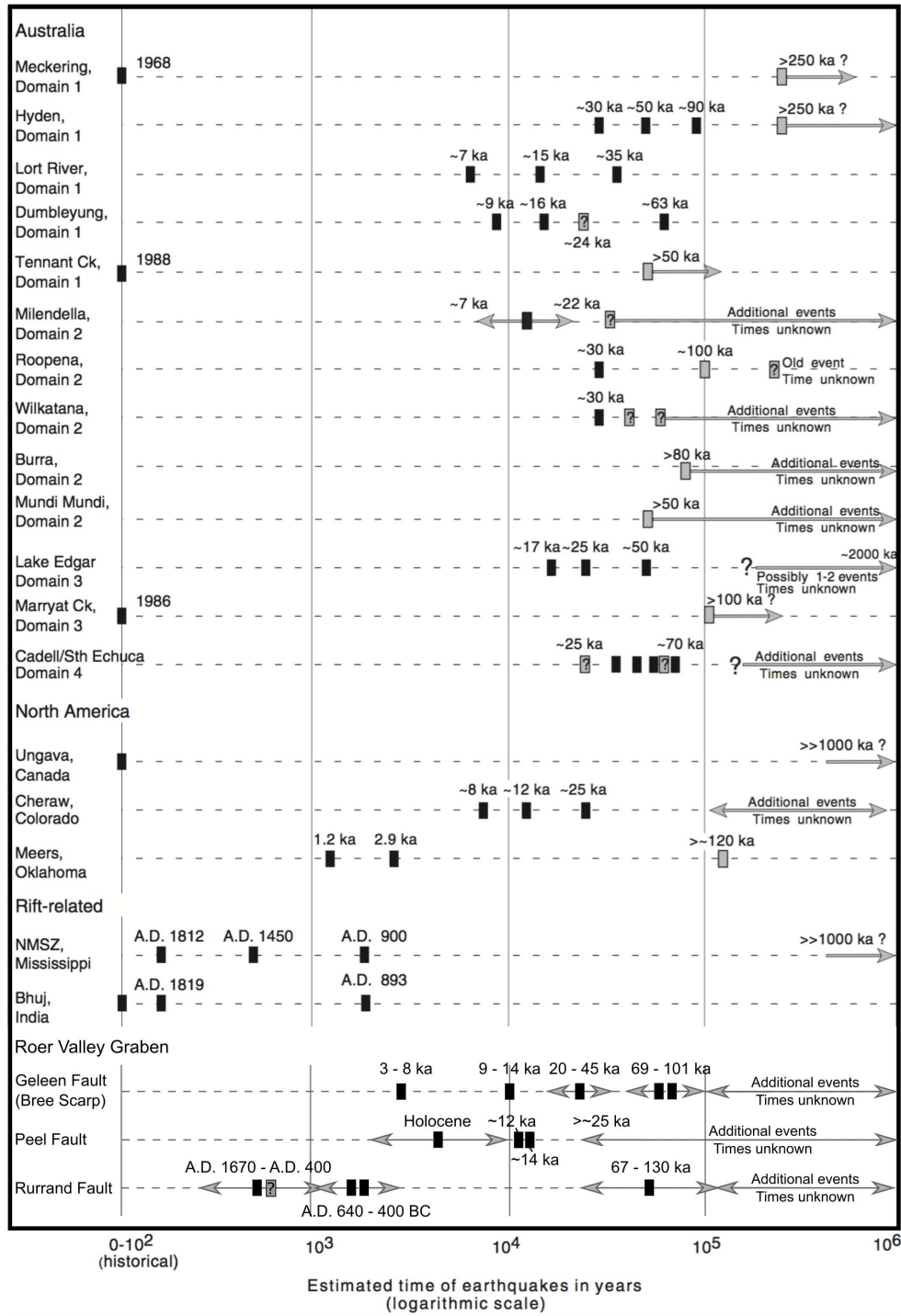


Figure 3. Compilation of surface-breaking earthquake recurrence data for SCR settings updated from [Crone et al., 2003; Clark et al., 2012]. Data for the Roer Valley Graben are from Vanneste et al. [2001], Frechen et al. [2001], and van den Berg et al. [2002]

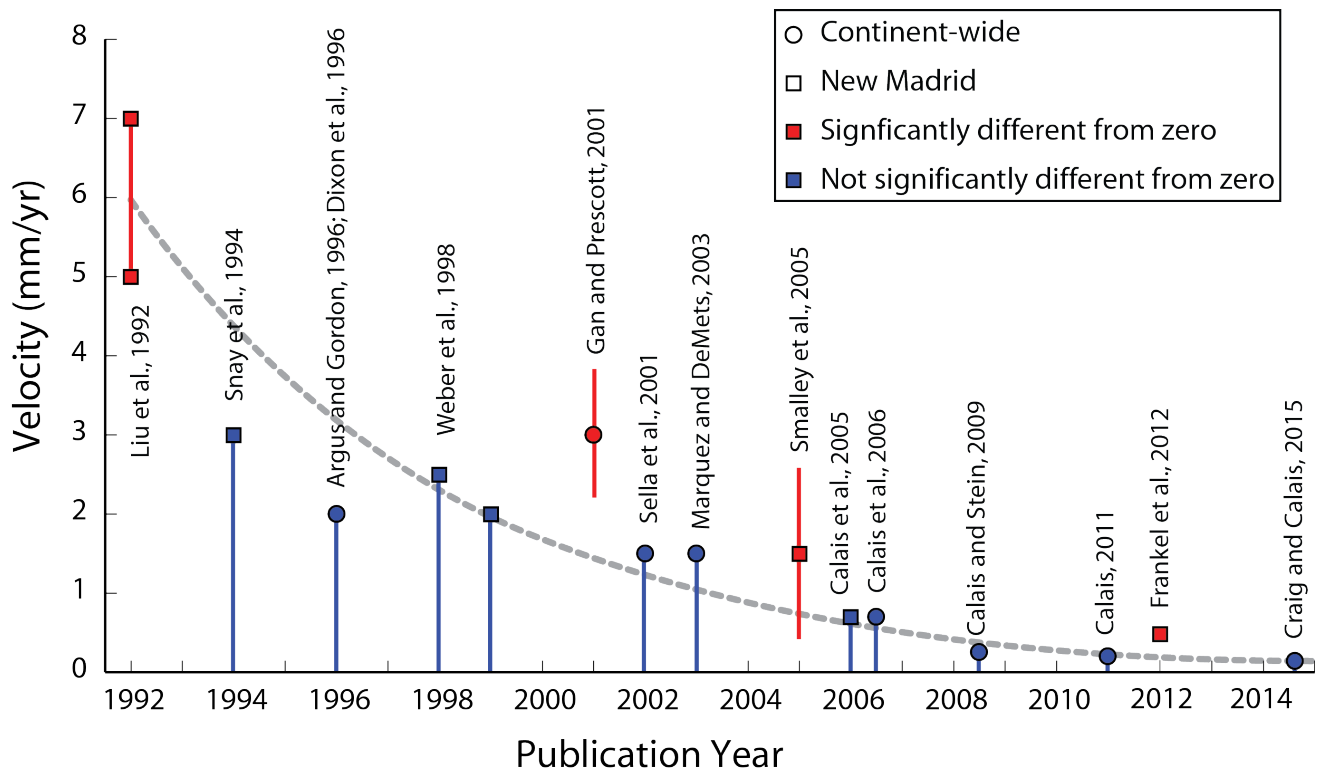


Figure 4. Maximum permissible deformation rates in the NMSZ as a function of publication year. Circles show continent-wide studies; squares show NMSZ studies. Red are publications claiming rates significantly different from zero; blue are upper bounds for publications claiming rates not significantly 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.

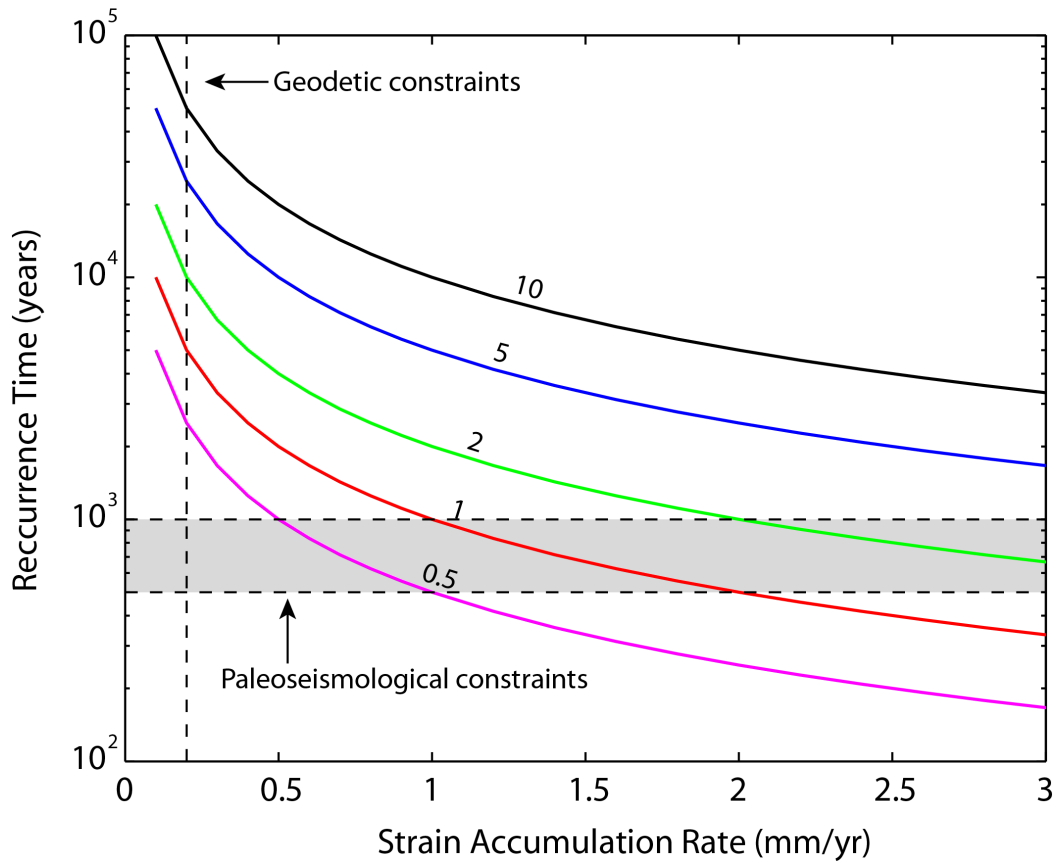


Figure 5. Earthquake recurrence interval as a function of slip rate across the New Madrid fault zone in steady-state, with two end-member values of coseismic slip for magnitude 7 (magenta and red curves) and magnitude 8 (blue and black curves) earthquakes. Numbers by each curve indicate the assumed coseismic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

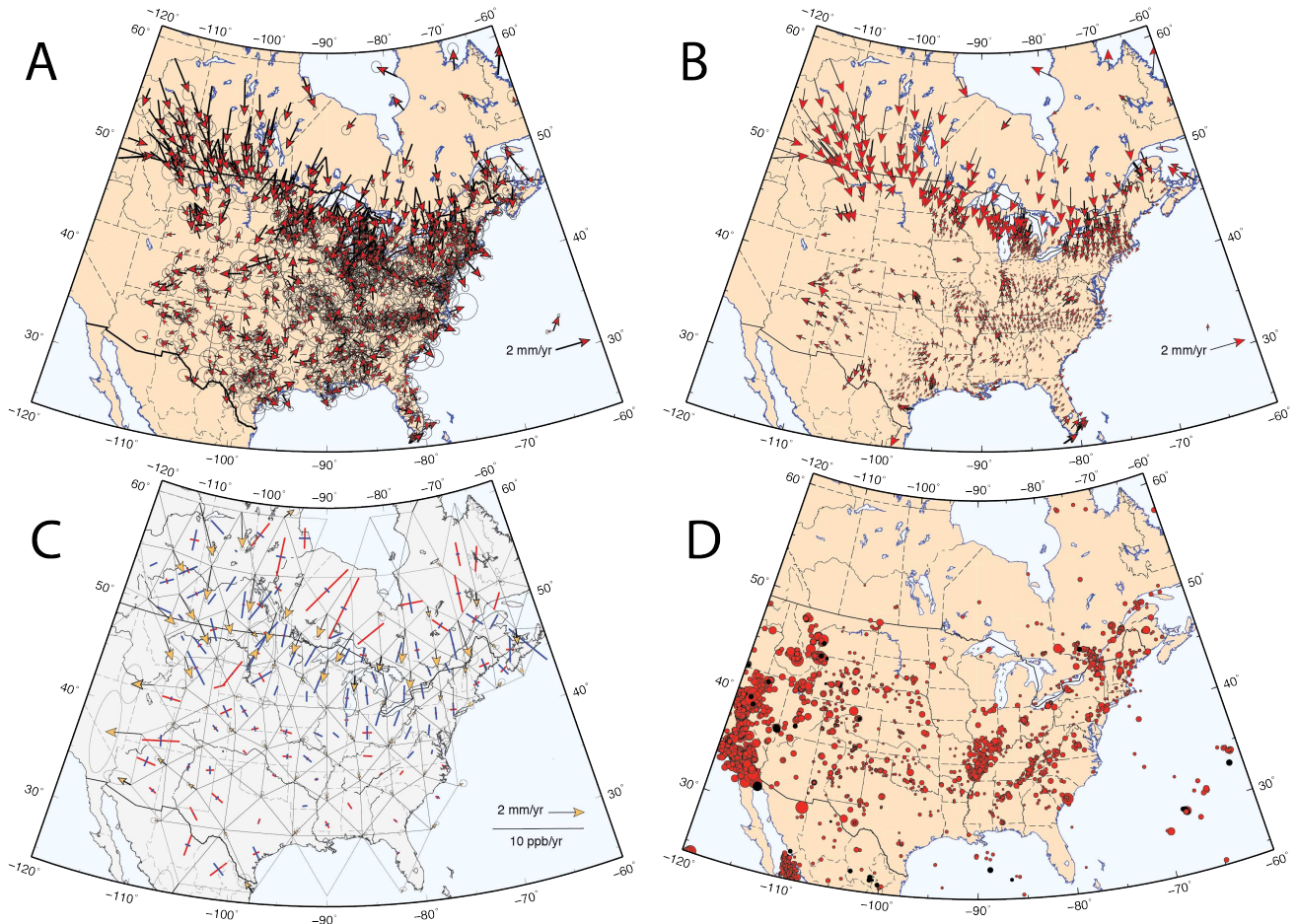


Figure 6. Comparison between current geodetic horizontal strain rates and seismicity in the North American plate interior. (A) Raw GPS site velocities after the removal of a rigid plate rotation. Ellipses are 95% confidence. (B) Spatially averaged residual velocities calculated using a nearest neighbor 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).

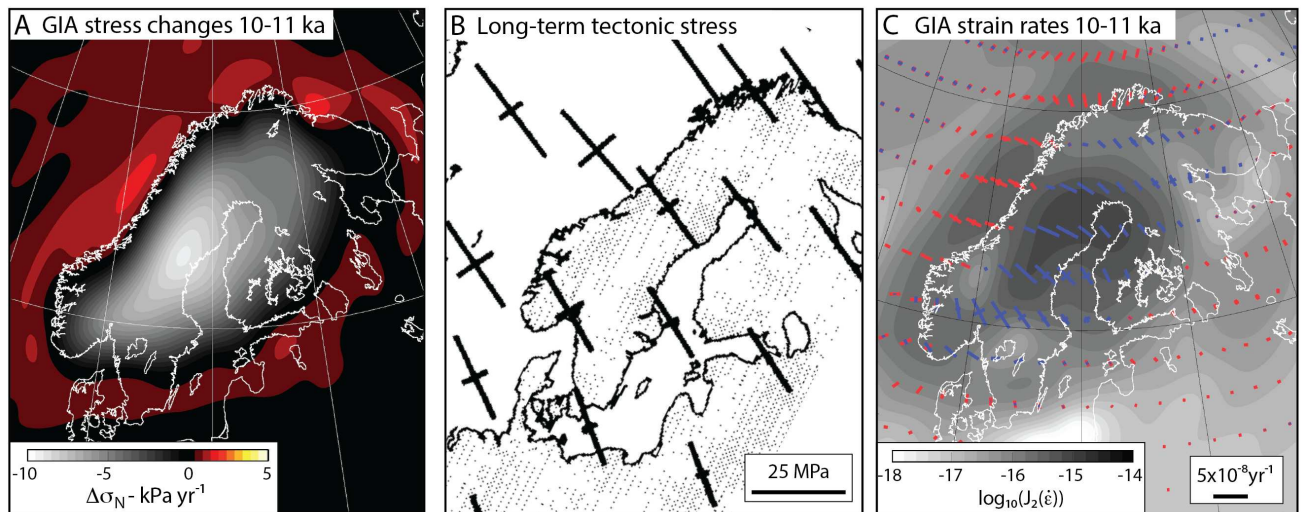


Figure 7. Evidence for the release of long-term tectonic strain stored in continental interiors [Craig *et al.*, 2016]. (A) Rate of change of applied normal stress on a fault representative of the overall trend of the majority of known major end-glacial faults in Fennoscandia (strike = 035° , and dip = 40°). (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).