A New Paradigm for Large Earthquakes in Stable Continental Plate Interiors

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Large earthquakes within stable continental regions (SCR) show that significant amounts of elastic strain can be released on geological structures far from plate boundary faults, where the vast majority of the Earth's seismic activity takes place. SCR earthquakes show spatial and temporal patterns that differ from those at plate boundaries and occur in regions where tectonic loading rates are negligible. However, in the absence of a more appropriate model, they are traditionally viewed as analogous to their plate boundary counterparts, occuring when the accrual of tectonic stress localized at long-lived active faults reaches failure threshold. Here we argue that SCR earthquakes are better explained by transient perturbations of local stress or fault strength that release elastic energy from a pre-stressed lithosphere. As a result, SCR earthquakes can occur in regions with no previous seismicity and no surface evidence for strain accumulation. They need not repeat, since the tectonic loading rate is close

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

1. Introduction

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

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

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

Although rare, SCR earthquakes can cause widespread damage because attenuation of seismic energy with distance is typically low in plate interiors [*Hanks and Johnston*, 1992] so that even moderate-size events can be devastating. Examples are the M_w6.2 1993 Latur earthquake (India) that caused over 8,000 fatalities and 300 million dollars in property damage [*Greene et al.*, 2000] or the Mw7.7 2001 Bhuj earthquake (India) that caused more than 20,000 fatalities and more than 4 billion dollars in total damage [*Maurer and Oblitas*, 2001]. A repeat of the 1811-1812 M7-7.5 New Madrid earthquakes in the central U.S. today is estimated to cost up to 300 billion dollars in damage [*Spencer et al.*, 2008] and similar figures would likely result from repeats of the 1756 Düren (Germany) or 1356 Basel (Switzerland) earthquakes in these highly populated regions of western Europe [*Allman and Smolka*, 2001]. Even though SCR earthquakes release only a few percent of the total seismic energy of the planet, they strike regions where the population and the infrastructures are often ill-prepared, even in developed countries.

Because SRC earthquakes are infrequent, occur in regions where present-day strain rates are very low, and rupture faults that are difficult to identify geologically, quantifying the associated hazard is a challenge [*Ellsworth et al.*, 2015; *Petersen et al.*, 2015]. Yet, accurate hazard estimates in SCRs are important for engineering design, in particular in the "post–Fukushima" era [*Joskow and Parsons*, 2012] for nuclear infrastructure that are designed for safety on a 10,000 yr timescale. Because our understanding of the earthquake process in SCRs leaves much to be desired, hazard calculations often implicitly use concepts and methods developed for plate boundaries.

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

2. The paradigm

The current paradigm for the earthquake process is well established, at least in general [e.g., *Kanamori and Brodsky*, 2004] (Figure 2, top). Stress builds up on faults over time as a result of steady plate motions, until their frictional strength is exceeded – at which point they rupture in an earthquake with a certain stress drop. Once unloaded by the earthquake, the fault takes a variable amount of time to be reloaded to the point of rupture depending on its strength, the magnitude of the stress drop, and the reloading stressing rate. Hence strain - the associated measurable quantity - continues building up and the fault will eventually rupture again so that the cycle repeats, regularly or irregularly. Over several such cycles, a balance results between the rates at which stress (or strain) accrues and is released. As a result, "geological slip rates" (actually strain release rates) should agree with "geodetic slip rates" (actually strain accumulation rates) so that a fault system conserves energy. As a corollary, past earthquakes, strain accrual rates, and fault segmentation contain some predictive information for long to medium-range forecasting of future earthquakes [e.g., *Field et al.*, 2014].

This view is supported by geodetic studies at plate boundaries showing that steady plate motions are accommodated by localized elastic deformation of the crust that accumulates at steady rates close to active faults. Fault slip during large earthquakes episodically releases this elastic strain so that, over a few hundred years, the rates of strain accumulation and release balance. For instance, *Tong et al.* [2014] show that geodetic and geologic slip rates agree within uncertainties along the San Andreas fault, one of the best known active plate boundary fault systems. This reasoning also holds for slower systems within broad regions of continental deformation such as the Wasatch fault separating the 3600 m high Wasatch range from the Great Salt Lake basin. Paleoseismological data showing that the major normal fault strand has slipped at an average rate of 1.7±0.5 mm/yr over the past 10 ka [*Friedrich et al.*, 2003] are consistent with the 1.6±0.4 mm/yr strain loading rate determined from GPS measurements [*Chang et al.*, 2006]. In addition, the Gutemberg-Richter relation derived from the instrumental earthquake catalog, consistent with the rate of paleoearthquakes identified on the Wasatch fault [*Schwartz and Coppersmith*, 1984; *McCalpin and Nishenko*, 1996] with one M7 event per ∼1,000 years [*Pechmann and Arabasz*, 1995; *Stein et al.*, 2005], which requires a strain accumulation rate on the order of 1.5 mm/yr, consistent with the geodetic observations.

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

Alternatively, intraplate faults may be releasing strain stored in the elastic crust over long intervals but not necessarily localizing observable interseismic strain at their time of failure [*Calais and Stein*, 2009]. Transient variations in crustal stress or fault strength, if large enough compared to the background tectonic stressing rates, may trigger rupture [*Long*, 1988; *Calais et al.*, 2010], releasing elastic energy from a pre-stressed lithosphere [*Feldl and Bilham*, 2006]. Once the available stresses on a fault segment have been released in an earthquake, the low background tectonic stressing-rate is insufficient to reload that segment to failure threshold on an observable timescale. Faults may consequently appear to fail only once, as described below for a number of SCR ruptures [*Crone et al.*, 2003]. Thus some clusters of present-day intraplate seismicity are long aftershock sequences of large events [*Stein and Liu*, 2009; *Boyd et al.*, 2015]. In this view, SCR seismicity is predominantly a transient feature triggered or inhibited by secondary, non-tectonic sources of stress change rather than a steady-state response of faults to constant tectonic loading.

3. Some characteristics of large SCR earthquakes

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

Such sporadic occurrence of large earthquakes, sometimes in the form of a single event on an old fault lacking evidence of Quaternary or recent activity, is a characteristic shared by other SCRs (Figure 3). The M6.3 1969 Ceres (South Africa) earthquake, for instance, which ruptured a 20 km–long strike-slip fault segment that did not reach the surface, occurred in a region with no evidence for previous earthquake activity [*Smit et al.*, 2015]. The Hebron fault in western Namibia shows a well–preserved, 40 km–long scarp with up to 10 m–throw that formed during a single Holocene event with no evidence for other Quaternary rupture [*Viola et al.*, 2005; *White et al.*, 2009]. In central India, the 40-50 km–long Tarpi fault scarp formed in the Holocene in one (or more) thrust-faulting earthquakes with no evidence for previous events or additional rupture since then [*Copley et al.*, 2014]. In Australia, one of the most arid and slowly eroding SCRs where scarps may be preserved for tens of thousands to millions of years, *Clark et al.* [2012] identified 300 small scarps and noted their poor spatial correlation with contemporary seismicity [*Clark and Leonard*, 2015].

In the Central and Eastern U.S., where environmental conditions are less favorable for the preservation of scarps, *Madole* [1988] and *Crone and Luza* [1990] report two surface–breaking events 1,200–1,300 years ago that ruptured a $\sim$60 km–long segment of the Paleozoic Meers fault in Oklahoma, with no evidence for other events in at least the past 120,000 years. *Crone et al.* [1997b] report evidence for three large earthquakes during the past 25 ka on the 44 km–long Cheraw fault in Colorado, with the most recent event in the early Holocene. Large historical earthquakes occurred in 1755 (M>6.0, near Cape Ann, MA), 1811-1812 (M7–7.5, near New Madrid, MO), 1843 (M6.3, northeast Arkansas), 1895 (M6.6, near New Madrid, MO), and 1886 ($\sim$M7 near Charleston, SC) [e.g., *Johnston et al.*, 1994].

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

Contrary to other SCRs, Western Europe underwent relatively recent, large-scale tectonic activity with the emplacement of the Cenozoic rift system of western and central Europe [*Illies et al.*, 1981; *Dèzes et al.*, 2004]. Some of the most seismically active regions today follow the overall trace of that structure, such as the Roer Valley Graben, bounded by Quaternary scarps related to earthquake activity [*Ahorner*, 1975; *Camelbeeck and Meghraoui*, 1998; *Vanneste et al.*, 2013]. Paleoseismological investigations along its western border have identified a 12 km–long by 15–20 m–high scarp along the Bree fault that offsets the 350–700 ka main terrace of the Maas River by about 40 m [*Camelbeeck and Meghraoui*, 1996, 1998]. One of the trenches across the Bree scarp shows evidence for five earthquakes in the past 100 ka, the most recent ∼3 ka B.P. associated with a 10 km–long by 0.5 m–high rupture scarp [*Vanneste et al.*, 2001]. On the opposite side of the Roer graben (Peel fault), similar investigations identified three large earthquakes within the past 25 ka [*van den Berg et al.*, 2002]. Detailed geomorphic analyses of the Bree and Peel scarps show that these border faults of the Roer Valley Graben were continuously active since the Middle Pleistocene, with earthquake magnitudes likely ranging from 6.3 to 7.0 [*Camelbeeck et al.*, 2007; *Vanneste et al.*, 2013].

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

The largest documented paleo-earthquakes in stable Europe likely occurred in the tectonically stable Fennoscandian craton in the late Pleistocene/early Holocene between 11 and 9 ka [*Muir-Wood*, 1989; *Olesen et al.*, 2004; *Mörner*, 2005; *Lagerbäck and Sundh*, 2008; *Jakobsson et al.*, 2014; *Olesen et al.*, 2013; *Smith et al.*, 2014]. They formed numerous scarps ranging from small-scale fractures to the 150 km–long Pärvie fault scarp, with offsets exceeding 15 m in places [*Muir-Wood*, 1989; *Lagerbäck and Sundh*, 2008]. Some of these earthquakes may have been larger than M8 [*Muir-Wood*, 1989; *Lagerbäck*, 1992; *Arvidsson*, 1996; *Lindblom et al.*, 2015] whereas the historic and instrumental seismicity of Fennoscandia rarely exceeds M_w5. There is little field evidence for on-going or repeated ruptures, but trenches across some faults indicate that faulting occurred in a single event [*Lagerbäck and Sundh*, 2008]. The clustering of these events 11-9 ka ago is a strong indication of a link with the last deglaciation [*Muir-Wood*, 1989; *Mörner*, 2005], as demonstrated by mechanical modelling studies [*Wu et al.*, 1999; *Wu and Johnston*, 2000; *Lambeck and Purcell*, 2003; *Turpeinen et al.*, 2008; *Steffen et al.*, 2014].

These examples illustrate the diversity of faults capable of generating large earthquakes in SCRs. Some occur in regions devoid of current seismicity or evidence for Quaternary ruptures. Some appear to have ruptured only once in recent times, while others show evidence for multiple events, sometimes clustered in time, separated by quiescent intervals of 10,000 to more than 100,000 years. Steady-state earthquake activity does not appear to persist in the long-term on any single fault. Hence seismic activity in SCRs appears to be episodic and sometimes clustered on faults that are active during relatively short time intervals, and then migrates to other structures [*Crone et al.*, 1997a; *Stein et al.*, 2009; *Crone et al.*, 2003; *Clark et al.*, 2012].

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

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

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

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

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

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

4. Is there measurable strain within SCRs?

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

Early geodetic measurements combining space and terrestrial data [*Liu et al.*, 1992] claimed 5-7 mm/yr of relative motion across the southern branch of the NMSZ (Figure 4). It was argued that this rate was in agreement with a steady-state fault system releasing one M_w8 earthquake every 500 to 1000 years, as expected then [*Johnston*, 1996]. However, similar observations in the northern part of the NMSZ led to inconclusive results, showing motions less than 3 mm/yr across the fault system [*Snay et al.*, 1994]. Similarly, episodic GPS measurements over the entire NMSZ reported no motion within uncertainties, placing an upper bound on deformation of 2.5 mm/yr [*Newman et al.*, 1999]. *Argus and Gordon* [1996] and *Dixon et al.* [1996] used continuously recording GPS stations throughout the plate interior to establish an upper bound of 2 mm/yr for residual motions across the Central–Eastern U.S., that was later reduced to 0.5 mm/yr thanks to longer time series and a much larger number of measurement sites [*Calais et al.*, 2006].

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

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

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

The search for tectonic strain accumulation within plate interiors has so far failed to identifiy seismically active regions where strain currently accrues at a measurable rate. However, horizontal deformation caused by GIA in plate interiors is easily captured by space geodetic measurements, at least in the current uplift areas. The BIFROST permanent GPS network in Sweden and Finland provided the first three-dimensional map of GIA over Fennoscandia [*Johansson et al.*, 2002; *Lidberg et al.*, 2010; *Kierulf et al.*, 2014]. Larger-scale studies have also identified horizontal motions outside of the uplifting areas in Western Europe [*Nocquet et al.*, 2005] and North America [*Calais et al.*, 2006; *Sella et al.*, 2007] indicating that large parts of those two continental interiors are experiencing long-wavelength strain caused by GIA.

Figure 6 shows a recently updated geodetic solution for stable North America following the methods described in *Calais et al.* [2006]. The velocities are residuals with respect to a subset of GPS stations located south of 42°N whose velocities fit a rigid rotation model with a reduced χ^2 close to unity. The regional pattern is consistent with that expected from GIA [*Peltier et al.*, 2015], with extension (up to 10^{-8} yr^{-1}) coincident with the uplift areas and shortening ($1–3\times10^{-9}$ yr^{-1}) associated with the subsiding forebuldge. A comparison with instrumental seismicity shows an interesting anticorrelation with GIA strain rates, except perhaps in the Lower St Lawrence area [*Mazzotti et al.*, 2005]. In other words, intraplate areas that are being strained as a result of GIA are not the ones that experience seismicity today. In addition, if GIA strain accrual was responsible for NMSZ earthquakes, the $\sim$1 mm/yr N–S shortening observed between the Great Lakes and the Gulf of Mexico would cause left-lateral and normal slip on the NE-SW–oriented New Madrid faults [*Craig and Calais*, 2014]. This is opposite to observations that show right-lateral and reverse motion consistent with large-scale tectonic stresses [*Hurd and Zoback*, 2012a]. These observations indicate that GIA strain accrual did not trigger New Madrid earthquakes, as also infered by *Wu and Johnston* [2000] on the basis of a modeling study.

5. The state of stress in SCRs

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

Similarly, stresses within continents vary only slowly with distance. Stress indicators, notably earthquake focal mechanisms, show broad areas with consistent maximum compressive horizontal stress (Shmax) directions consistent with plate-driving forces, locally modified by lithospheric properties in some regions [*Zoback and Zoback*, 1989; *Müller et al.*, 1992; *Heidbach et al.*, 2007, 2010]. In North America, Shmax shows a very consistent WSW-ENE direction across the central and eastern U.S., all the way to southeastern Canada [*Hurd and Zoback*, 2012a; *Herrmann et al.*, 2011]. This consistency is visible both in "natural" earthquakes and in the human–induced seismicity currently widespread thoughout Oklahoma and part of Texas. *McNamara et al.* [2015] show that well–induced earthquakes in central Oklahoma, some with magnitudes reaching 5.7 [*Keranen et al.*, 2013, 2014], occur on faults that are favorably oriented in a ENE-WSW compressive stress field, with focal mechanisms consistent with this background tectonic stress field. The same observation holds for the NMSZ, where *Hurd and Zoback* [2012b] show focal mechanism P-axes consistent with ENE-WSW Shmax orientation over much of the central and eastern U.S.

That the crust breaks on pre-existing faults favorably oriented with respect to the regional tectonic stress field with source mechanisms consistent with that stress field does not necessarily mean that this background tectonic stress is responsible for bringing individual faults to failure. For instance, the mechanism triggering recent seismicity in Oklahoma and Texas is wastewater injections following oil recovery, which increase pore pressure at depth, lowering effective normal stress on faults and bringing them closer to failure [*Keranen et al.*, 2014]. This mechanism was likely the cause of the M_w5.7, November 2011 earthquake in central Oklahoma, which was broadly felt and caused damage in the epicentral region [*Keranen et al.*, 2013]. Ample evidence shows that earthquakes are sometimes triggered by fluid injections during oil recovery or mining operations and by the filling of water reservoirs [*Rothé*, 1968; *Raleigh et al.*, 1972; *Simpson*, 1976; *Gupta*, 1985]. Two mechanisms have been invoked to explain the latter, either the increase of elastic stresses due to the flexure of the crust under the load, or the lowering of effective normal stress on faults as water diffuses down to hypocentral depths [*Simpson et al.*, 1988].

Seismic swarms of natural origin are also attributed to fluid overpressure following the diffusion of mantle volatiles [*Weise et al.*, 2001; *Špičák and Horálek*, 2001; *Cappa et al.*, 2009] or meteoric water [*Hainzl et al.*, 2006; *Costain and Bollinger*, 2010; *Got et al.*, 2011; *Leclère et al.*, 2013] to seismogenic depth. Many such studies argue that fluid overpressure at depth plays a key role in earthquake nucleation by lowering effective stress on fault segments that are nearly critically stressed for shear failure [*Sibson*, 1990; *Cappa et al.*, 2009; *Wang and Manga*, 2009].

Alternately, earthquakes can be triggered by changes in elastic stresses driven by the loading or unloading of the crust by surface or ground water. *González et al.* [2012] showed that stress changes caused by water extraction from a shallow aquifer likely triggered a M_w5.1 earthquake near Lorca, Spain, in 2011. Its source mechanism indicates reverse faulting on the SW-NE–oriented Murcia fault and reflects the regional stress field imparted by the oblique convergence between Nubia and Eurasia [*Nocquet and Calais*, 2004]. *Heki* [2003] explains seasonal cycles in earthquake occurrence in northern Japan as a result of the modulation of the regional stress field by stresses of a few kPa caused by snow loading. *Bollinger et al.* [2007] and *Bettinelli et al.* [2008] report seasonal strain and stress variations in the Nepal Himalaya that correlate with seasonal variations in seismicity, with summer seismicity suppressed by stress-loading accompanying monsoon rains.

Though most of the triggered earthquakes referred to above are small, some may be much larger. In northern Sweden and Finland, the series of M7–8 end-glacial earthquakes around 9,500 years ago has been interpreted as a result of decreased normal stresses on steeply dipping reverse faults as the Fennoscandian ice sheet was rapidly melting [*Wu et al.*, 1999; *Turpeinen et al.*, 2008; *Steffen et al.*, 2014]. In the Basin and Range province of the Western U.S., *Hetzel and Hampel* [2005] show that the increased slip rate on the Wasatch fault since about 17 ka – more specifically the "strain release rate", determined from the paleoearthquake record [e.g., *Friedrich et al.*, 2003] – could be explained by the stress changes induced by a regression of Lake Bonneville and the melting of glaciers in the Uinta and Wasatch mountains. In the upper Mississippi embayment of the Central U.S., *Calais et al.* [2010] showed that an intense erosional event between 16 and 10 ka caused upward flexure of the lithosphere and a reduction of normal stresses in the upper crust sufficient to unclamp pre-existing faults close to critical failure, possibly triggering the sequence of large Holocene earthquakes in the region. Once a large earthquake has occurred, stress changes may trigger additional regional events via elastic (coseismic) or viscoelastic (postseismic) stress transfer and a clustered sequence of events may develop [*Kenner and Segall*, 2000; *Mueller et al.*, 2004].

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

In SCRs, however, tectonic stressing occurs at rates that are at least 100 times lower than at major plate boundaries, so external forcings may dominate and localize earthquake activity in space and time. Hence, the timing and location of SCR earthquakes may be largely independent of long-term tectonic loading under a regional, essentially time-invariant, tectonic stress field, but instead be determined by small transient stress changes in a crust close to failure equilibrium. Regardless of the specific transient stress change that brings a fault to failure, the resulting rupture mechanism will however be consistent with the background static tectonic stress field, which defines the style of faulting.

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

Seismically active areas within SCRs are sometimes interpreted as the result of local concentrations of tectonic stress or as mechanically weak regions [e.g., *Sykes*, 1978]. Various mechanisms have been proposed including stress concentration at intersecting faults [*Talwani*, 1999], around buried intrusions in the crust [*Campbell*, 1978; *Zoback and Richardson*, 1996; *Pollitz et al.*, 2001], or at the tip of a low velocity upper mantle seismic anomaly [*Zhan et al.*, 2016]. Other proposed mechanisms involve local weakening of the lower crust either thermally– [*Grollimund and Zoback*, 2001; *Kenner and Segall*, 2000] or geochemically–induced [*Chen et al.*, 2016] or bulk weakening in regions where the mechanically strong mantle lithosphere is absent [*Tesauro et al.*, 2015]. Although all these mechanisms are plausible, they would persist over long geologic time intervals, whereas SCR seismicity does not. For instance, *Van Arsdale* [2000] show that the NMSZ was activated around 10,000 yrs ago after millions of years of tectonic quiescence during which all the processes listed above would have been operating, had they existed. Therefore, although such mechanisms may locally perturb the long-term, static, tectonic stress field of a continental interior, they do not explain why SCR seismicity is episodic, with long and variable seismically quiet time intervals, and some faults rupturing only once.

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

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

6. Conclusion: a possible mechanism for large SCR earthquakes

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

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

Another indirect line of evidence is *Craig et al.* [2016]'s observation that the end-glacial, reverse-faulting earthquakes of Fennoscandia occurred while the horizontal strain-rate was extensional. Consequently, faulting did not release extensional strain accumulating at the time of failure, but instead released compressional strain that had accumulated through long-term tectonic forcing (Figure 7). This forcing is likely due to the ridge-push force exerted by the cooling and thickening oceanic lithosphere formed at the Mid-Atlantic Ridge [*Gölke and Coblentz*, 1996; *Pascal et al.*, 2010], with a possible contribution from compressional stresses generated by ice loading during the last glacial period.

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

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

If faults in SCRs are at failure equilibrium in a pre-stressed crust able to sustain large differential stresses, then the occurrence of SCR earthquakes in time and space is better explained by transient perturbations of stress or fault strength than by the slow accumulation of tectonic stress on long-lived active faults. In other words, while tectonic (or thermal) stress provides the energy that is released during large SCR earthquakes, earthquake occurrence results from a local and temporary perturbation of stress or crustal strength near a fault that is favorably oriented relative to the regional tectonic stress field. These transient perturbations may result from fluid pore pressure increase at earthquake nucleation depth, or from local changes in "secondary stresses" – for instance caused by surface loading/unloading. Hence, earthquake sequences appear episodic and clustered – sometimes involving a single rupture – rather than persistent.

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

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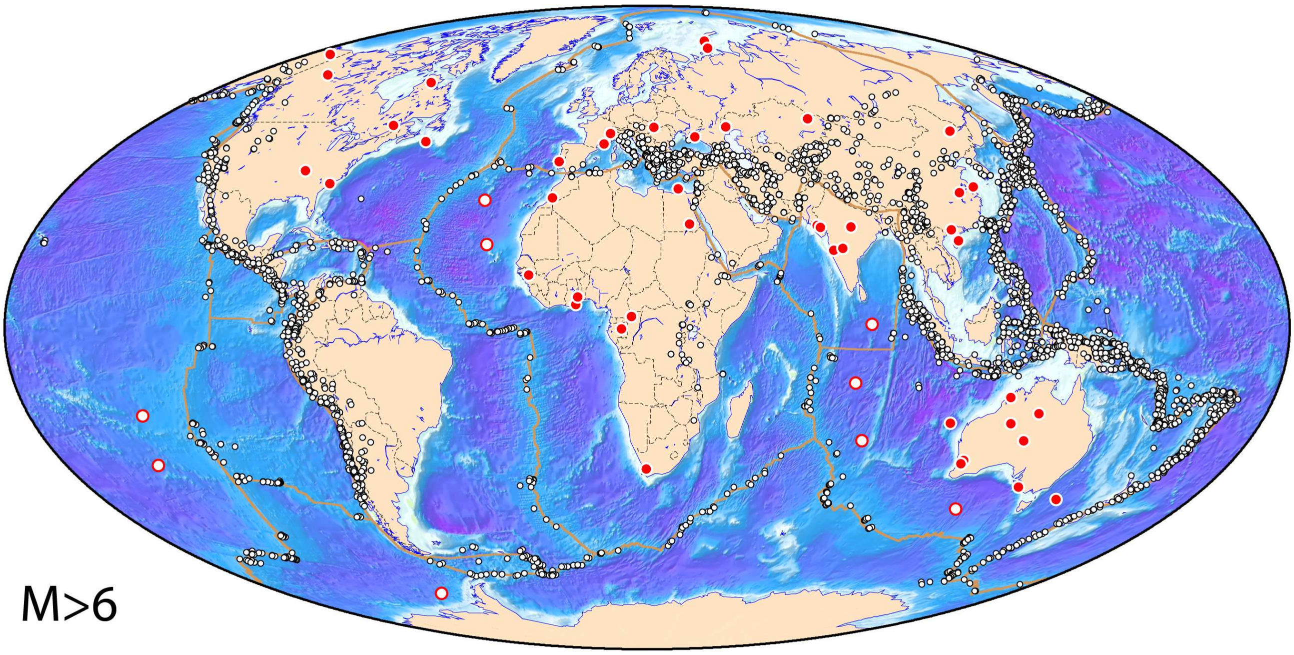

Figure 1. Worldwide seismic activity (`http://earthquake.usgs.gov/`). Large circles show M>6 intraplate earthquakes: red for stable continental regions, white for stable oceanic regions.

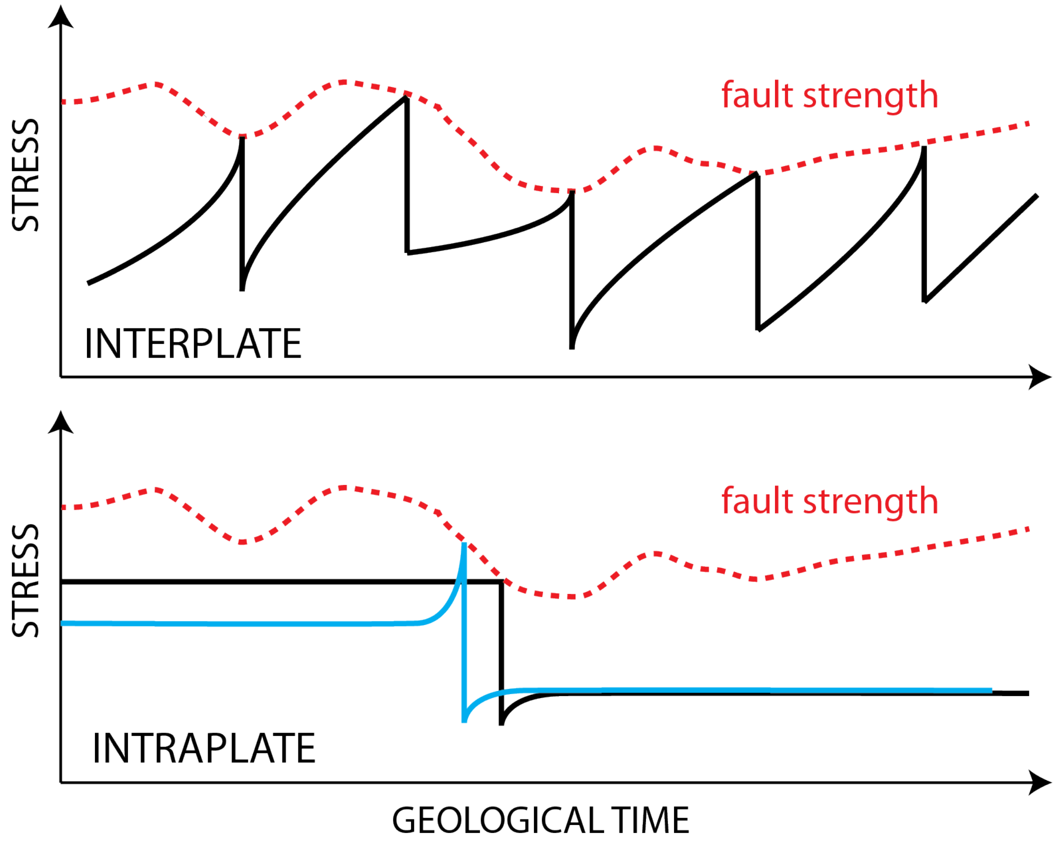

Figure 2. Stress changes and earthquake sequence. Top: a sequence of plate boundary earthquakes occurs as a result of tectonic loading, at a rate that may slightly vary with time, and temporal variations in fault strength [*Kanamori and Brodsky*, 2004]. Bottom: in SCR settings, stress accrues at very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

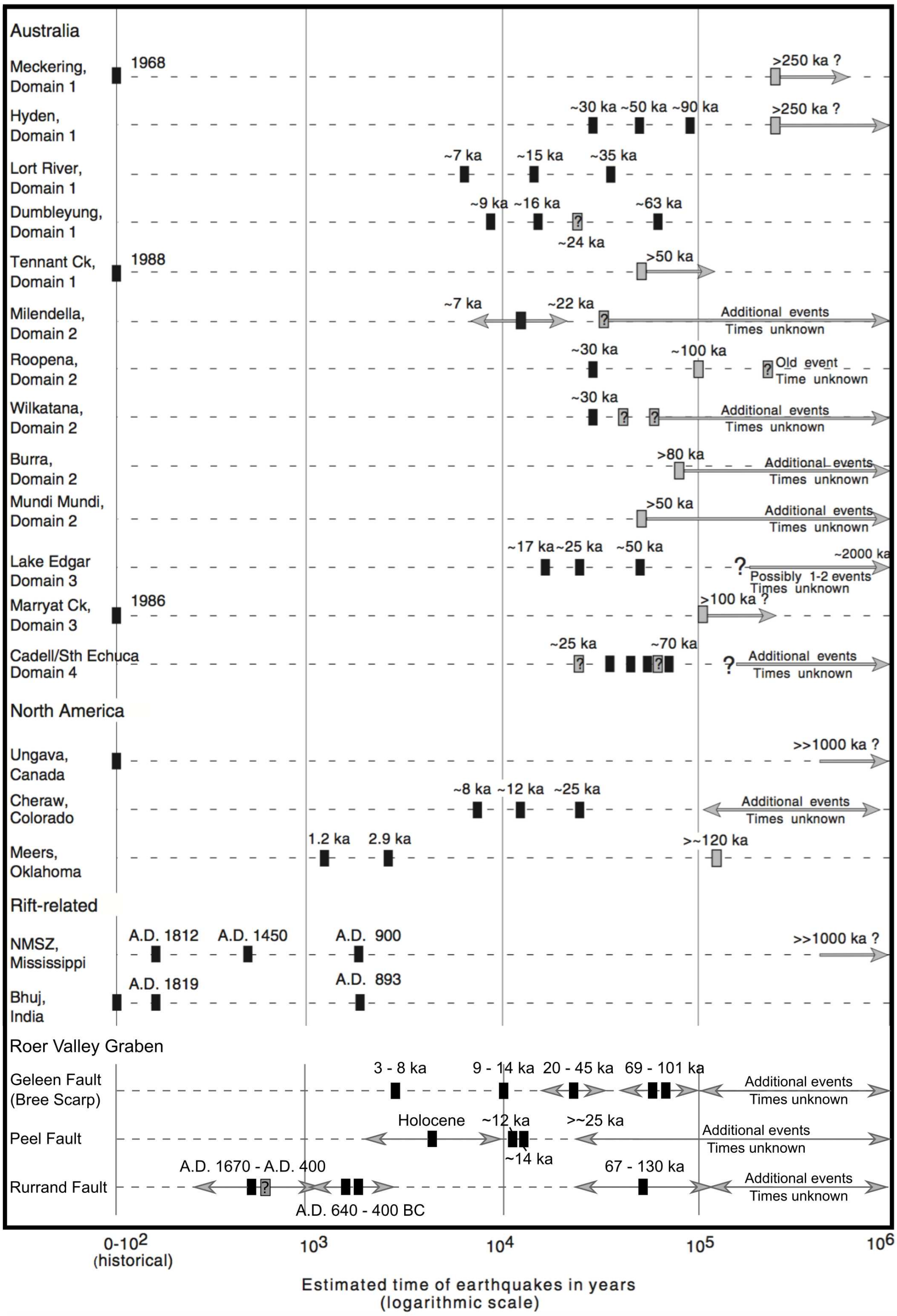

Figure 3. Compilation of surface-breaking earthquake recurrence data for SCR settings updated from [*Crone et al.*, 2003; *Clark et al.*, 2012]. Data for the Roer Valley Graben are from *Vanneste et al.* [2001], *Frechen et al.* [2001], and *van den Berg et al.* [2002]

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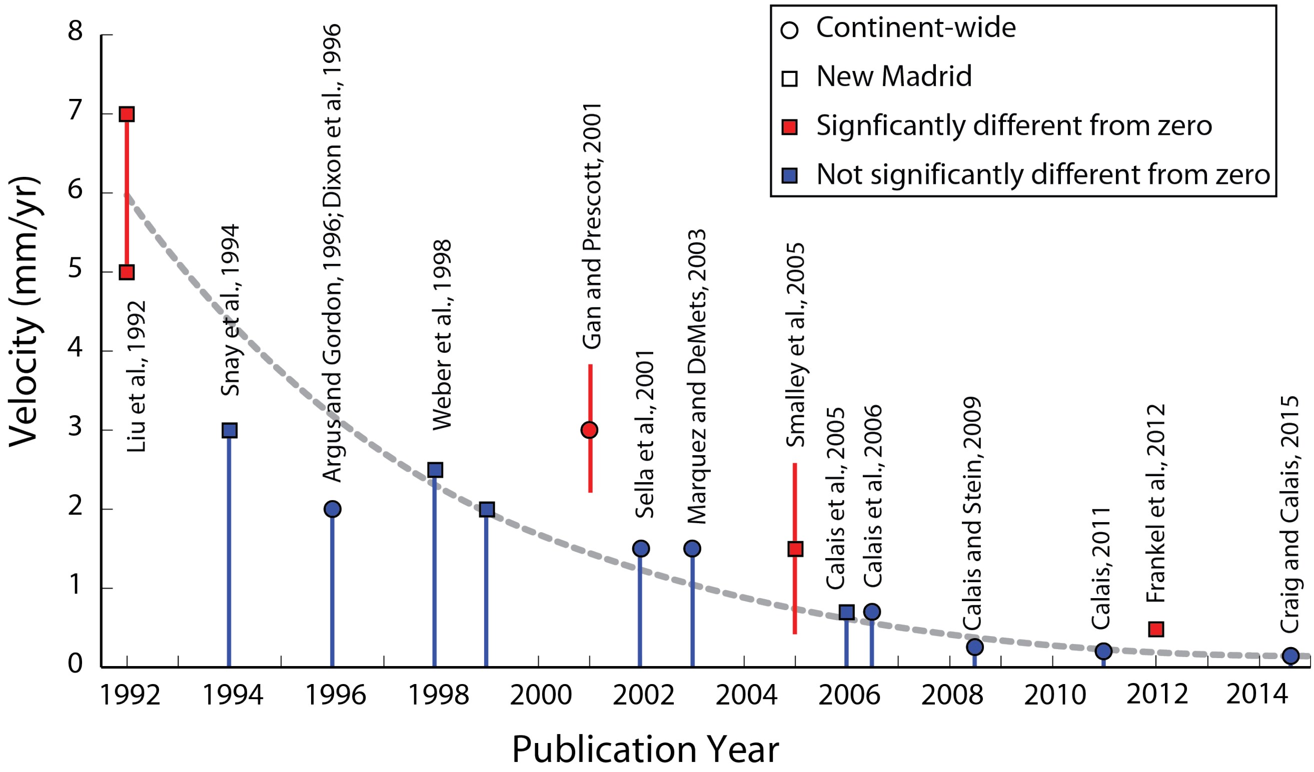

Figure 4. Maximum permissible deformation rates in the NMSZ as a function of publication year. Circles show continent-wide studies; squares show NMSZ studies. Red are publications claiming rates significantly different from zero; blue are upper bounds for publications claiming rates not significantly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

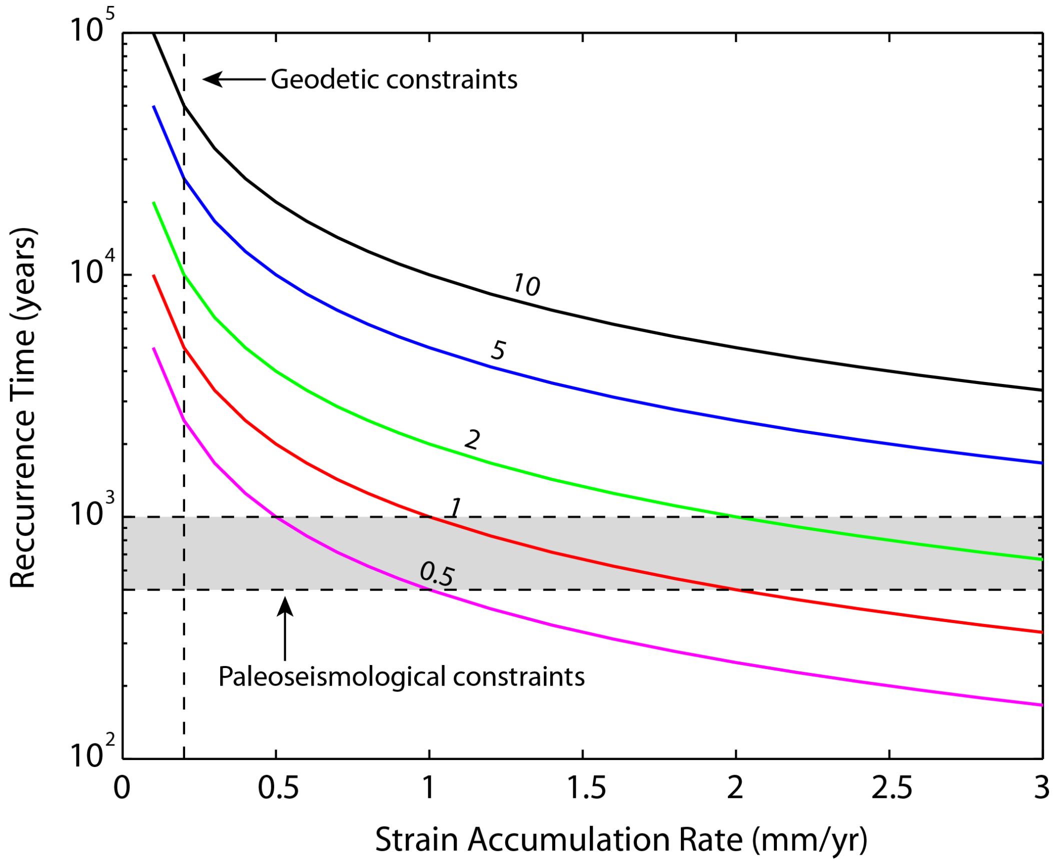

Figure 5. Earthquake recurrence interval as a function of slip rate across the New Madrid fault zone in steady-state, with two end-member values of coseismic slip for magnitude 7 (magenta and red curves) and magnitude 8 (blue and black curves) earthquakes. Numbers by each curve indicate the assumed coseismic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

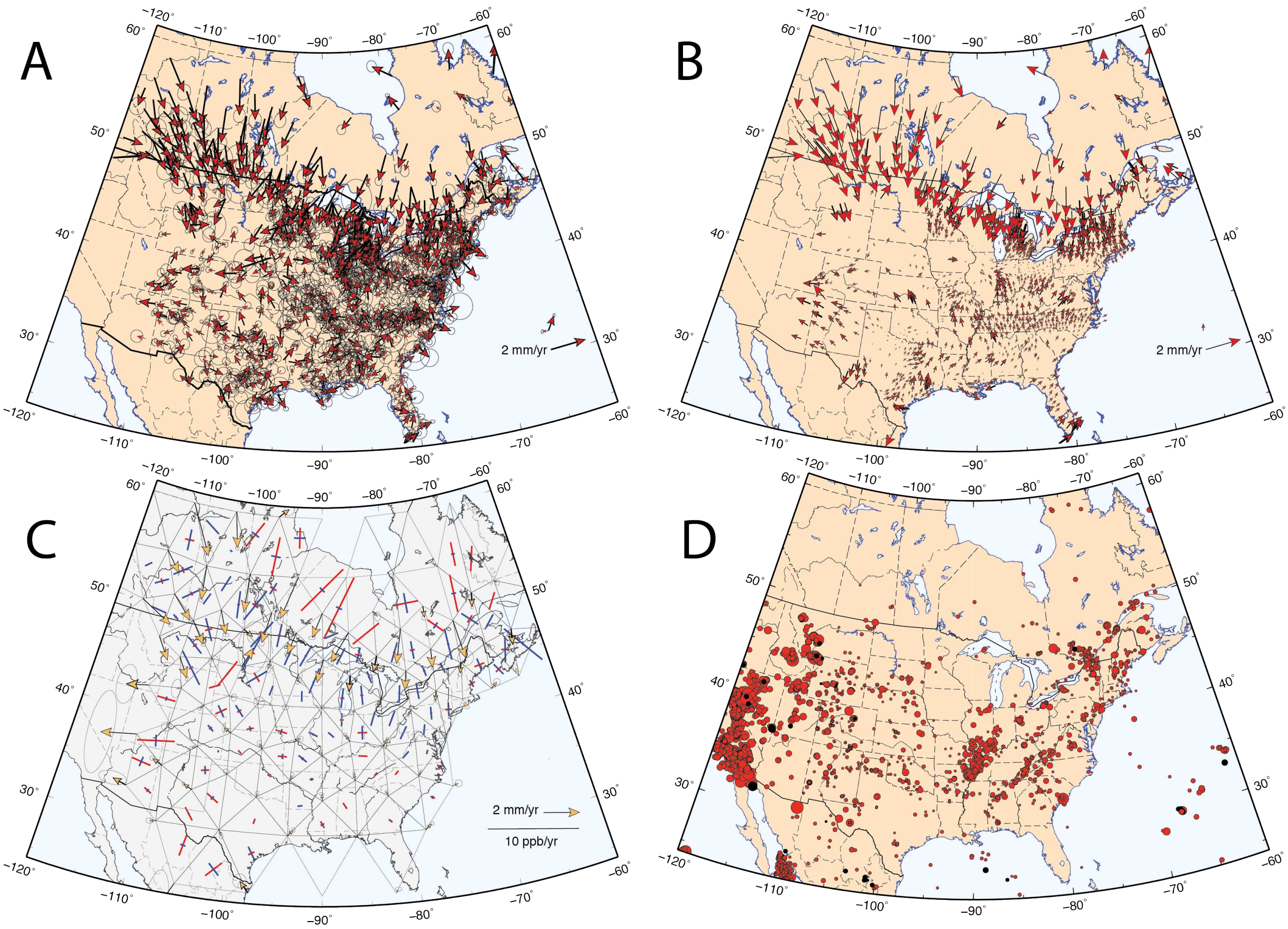

Figure 6. Comparison between current geodetic horizontal strain rates and seismicity in the North American plate interior. (A) Raw GPS site velocities after the removal of a rigid plate rotation. Ellipses are 95% confidence. (B) Spatially averaged residual velocities calculated using a nearest neighbor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

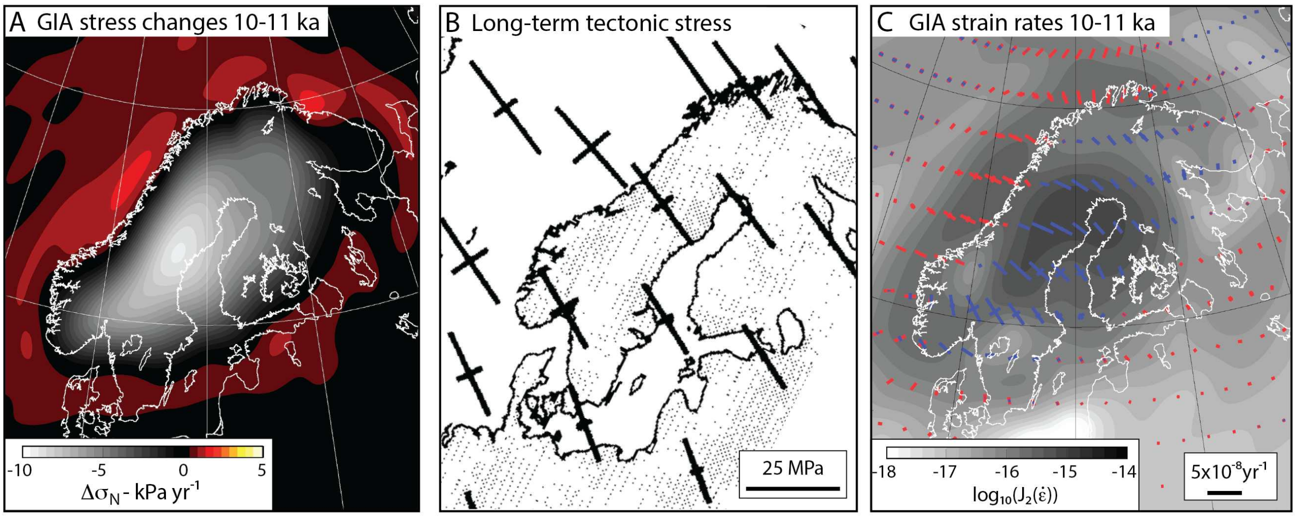

Figure 7. Evidence for the release of long-term tectonic strain stored in continental interiors [*Craig et al.*, 2016]. (A) Rate of change of applied normal stress on a fault representative of the overall trend of the majority of known major end-glacial faults in Fennoscandia (strike = 035°, and dip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