# Mineral, Virginia, earthquake illustrates seismicity of a passive-aggressive margin

Emily Wolin,[1] Seth Stein,[1] Frank Pazzaglia,[2] Anne Meltzer,[2] Alan Kafka,[3] and Claudio Berti[2]

[1] The August 2011 M 5.8 Mineral, Virginia, earthquake that shook much of the northeastern U.S. dramatically demonstrated that passive continental margins sometimes have large earthquakes. We illustrate some general aspects of such earthquakes and outline some of the many unresolved questions about them. They occur both offshore and onshore, reaching magnitude 7, and are thought to reflect reactivation of favorably-oriented, generally margin-parallel, faults created during one or more Wilson cycles by the modern stress field. They pose both tsunami and shaking hazards. However, their specific geologic setting and causes are unclear because large magnitude events occur infrequently, microseismicity is not well recorded, and there is little, if any, surface expression of repeated ruptures. Thus presently active seismic zones may be areas associated with higher seismicity over the long term, the present loci of activity that migrates, or aftershock zones of large prehistoric earthquakes. The stresses causing the earthquakes may result from platewide driving forces, glacial isostatic adjustment, localized margin stresses, and/or dynamic topography. The resulting uncertainties make developing cost-effective mitigation strategies a major challenge. Progress on these issues requires integrating seismic, geodetic, and geological techniques. **Citation:** Wolin, E., S. Stein, F. Pazzaglia, A. Meltzer, A. Kafka, and C. Berti (2012), Mineral, Virginia, earthquake illustrates seismicity of a passive-aggressive margin, *Geophys. Res. Lett.*, *39*, L02305, doi:10.1029/2011GL050310.

Passive-Aggressive behavior is a form of covert abuse. Covert abuse is subtle and veiled or disguised by actions that appear to be normal.

[(http://divorcesupport.about.com/od/abusiverelationships/a/Pass_Agg.htm)]

## 1. Introduction

[2] On August 23, 2011, residents of the U.S.'s east coast may have felt that the earth had turned passive-aggressive, when they were surprised by shaking from the M 5.8 Mineral, Virginia earthquake. Many had assumed that earthquakes large enough to damage structures like the Washington Monument did not occur in the region. From a tectonic standpoint, their assumption made sense. Eastern

North America is a "passive" continental margin, along which the continent and seafloor are part of the same plate. The fundamental tenet of plate tectonics, articulated by *Wilson* [1965, p. 344], is that "plates are not readily deformed except at their edges." Hence if plates were the ideal rigid entities assumed in many applications, such as calculating plate motions, passive continental margins should be seismically passive.

[3] In reality, however, large and damaging earthquakes occur along passive margins worldwide [*Stein et al.*, 1979, 1989; *Schulte and Mooney*, 2005] (Figure 1). These earthquakes release a disproportionate share, about 25%, of the net seismic moment release in nominally-stable continental regions [*Schulte and Mooney*, 2005]. They were recognized even prior to the formulation of plate tectonics by *Gutenberg and Richter* [1954, p. 82], who noted, "Nearly all stable masses exhibit marginal features which are seismically active." Hence although passive margin earthquakes are only a very minor component of global seismicity, they are not uncommon, and can be thought of as analogous to the medical case of a "common rare disease."

[4] However, no comprehensive model exists to explain these earthquakes. We know little about their causes and the possible hazards they pose, partly because they are relatively rare, due to the slow deformation at these margins. They are generally thought to reflect reactivation of ancient, favorably-oriented, faults created by previous continental collision and breakup. This view is consistent with geological observations that passive margins are often reactivated in compression [*Cloetingh et al.*, 2008]. However, verifying and using this generalization is difficult. Thus our goal here is to illustrate some general observations about passive margin earthquakes and outline some of the many unresolved questions about them.

[5] Eastern North America has impressive examples of passive margin seismicity. The best known in modern times, the 1929 M 7.2 earthquake on the Grand Banks of Newfoundland [*Hasegawa and Kanamori*, 1987; *Bent*, 1995], caused a large ($\sim10^{11}$ m³) landslide of thick continental slope sediments, which cut trans-Atlantic telegraph cables and generated a tsunami. The resulting 28 fatalities are either all or most of Canada's known deaths from earthquakes. The earthquake was felt as far away as New York and Montreal (http://earthquakescanada.nrcan.gc.ca/histor/20th-eme/1929/1929-eng.php) and thus gives a useful comparison with the similar intensity distribution for the largest of the 1811-12 New Madrid earthquake sequence [*Hough*, 2008]. Other notable events include the 1933 M 7.3 Baffin Bay [*Bent*, 2002], 1755 M $\sim$6 Cape Ann, Massachusetts [*Ebel*, 2006], and 1886 M $\sim$7 Charleston [*Hough*, 2008] earthquakes. Hence the 2011 Virginia

[1]Department of Earth and Planetary Sciences, Northwestern University, Evanston, Illinois, USA.
[2]Department of Earth and Environmental Sciences, Lehigh University, Bethlehem, Pennsylvania, USA.
[3]Weston Observatory, Department of Earth and Environmental Sciences, Boston College, Weston, Massachusetts, USA.
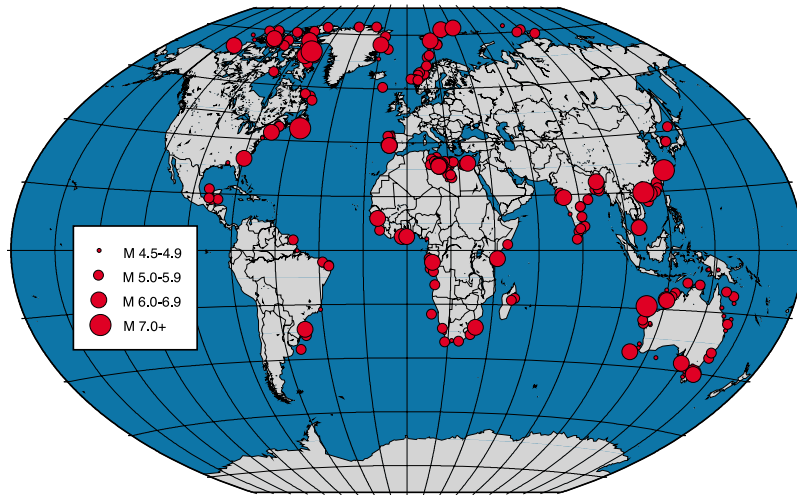
**Figure 1.** Map of passive continental margin earthquakes. Data from *Schulte and Mooney* [2005].

earthquake is part of a diffuse seismic zone spanning the continental margin (Figure 2).

## 2. Offshore Earthquake Hazards

[6] Among the intriguing challenges posed by these earthquakes is assessing their societal hazard. For this purpose, the offshore and onshore earthquakes differ. For onshore earthquakes, the primary hazard is shaking. For large offshore earthquakes like the Grand Banks, the primary hazard can be a resulting tsunami [*ten Brink*, 2009], because the intensity of strong shaking decreases rapidly with distance. To assess this hazard, a crucial question is whether the slumps and tsunamis require an earthquake trigger, or can arise from slope instability alone. In the former case, the hazard can be assessed using earthquake frequency-magnitude data [*Swafford and Stein*, 2007].

[7] Studies of the Grand Banks earthquake, the only well studied tsunamigenic earthquake off eastern North America, have come to opposing conclusions. *Hasegawa and Kanamori* [1987] inferred from first motions and surface wave spectra that the source was a single force, e.g. a landslide. In contrast, *Bent* [1995] used waveform modeling to infer a double couple or earthquake source.

[8] An alternative approach is to consider aftershocks, which result from changes of stress and fault properties induced by the main shock. At plate boundaries, plate motion at rates of tens of mm/yr quickly reloads a fault after a large earthquake and overwhelms the effects of the main shock within about a decade. However, faults within continental interiors are reloaded at rates significantly less than 1 mm/yr, allowing aftershocks to continue, potentially, for hundreds of years. A model based on laboratory experiments [*Dieterich*, 1994], which predicts that the length of aftershock sequences varies inversely with the rate at which faults are loaded, accords with observations from a range of faults [*Stein and Liu*, 2009]. For example, aftershocks continue today from the 1959 Hebgen Lake, Montana, earthquake, and seismicity in the areas of past large intracontinental earthquakes, including those in New Madrid, Missouri (1811-1812) and Charlevoix, Quebec (1663) appear to be aftershocks. Many small earthquakes in the eastern U.S. may be aftershocks of strong earthquakes that

took place hundreds or thousands of years ago [*Ebel et al.*, 2000].

[9] Because a large landslide should not be followed by a long aftershock sequence, studies of aftershocks could make it possible to distinguish landslides triggered by earthquakes from those due to slope instability alone. We thus examined the seismicity of the Grand Banks and Baffin Bay areas between 1920 and 2009 (Figure 3, top). Seismicity in the Grand Banks exhibits a rough linear trend along the fault plane proposed by *Bent* [1995], although several of the largest aftershocks have been fixed at the main shock
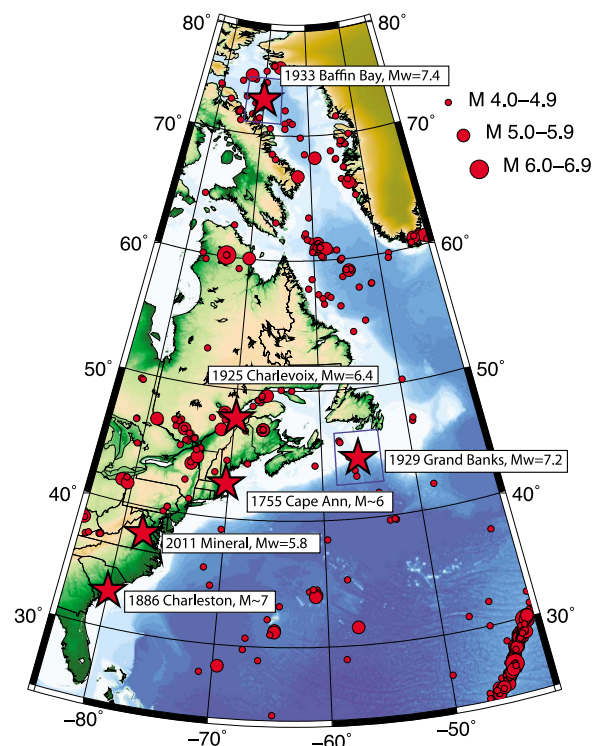


**Figure 2.** Seismicity of the eastern North American continental margin since 1980 taken from ANSS and Earthquakes Canada catalogs. Major historical events are also shown. Boxes denote aftershock areas used in Figure 3.
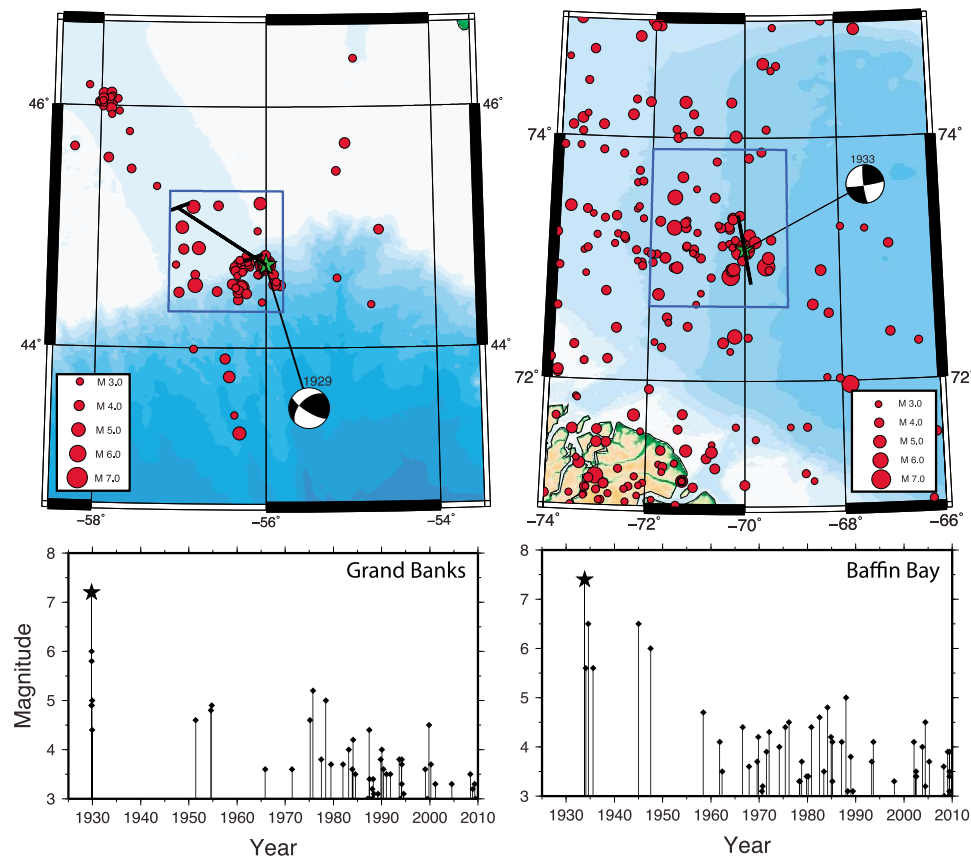
**Figure 3.** Aftershock locations and histories for the 1929 Grand Banks and 1933 Baffin Bay earthquakes.

epicenter. Seismicity in Baffin Bay is more scattered, either intrinsically or due to location uncertainties [*Qamar*, 1974; *Bent*, 2002]. In both areas, the largest (M~5-6) events occur within 30-40 years of the main shock and within 1-2 fault lengths of its epicenter.

[10] Assuming that these larger events outline the region where most aftershocks occur, we treat as aftershocks any event in a rectangular region enclosing the main shock and subsequent large events. Choosing such a region was straightforward near the Grand Banks due to the linear trend of epicenters. In Baffin Bay, a scattered distribution of the largest earthquakes and an abundance of recent, small events made selecting an aftershock region more difficult. Our selection was based on the location of the largest aftershocks in the 40 years after the main shock and the geometry of the fault plane [*Bent*, 2002]. The box shown excludes several M~6-6.5 quakes from 1933-1957 as well as the 1963 M 6.3 Baffin Island earthquake. Most of these events occurred too far from the 1933 epicenter to qualify as aftershocks, although many locations for earlier events have considerable uncertainty. We discount the 1963 earthquake as an aftershock due to its thrust mechanism and WNW-striking fault geometry [*Stein et al.*, 1979]. Although selecting these aftershock regions is subjective, our results are robust to variations in size and location of the chosen regions.

[11] Figure 3 (bottom) shows event magnitudes versus time for the two regions. Despite the incomplete and non-uniform catalogs, both show decay characteristic of aftershock sequences. The Grand Banks experienced a number of aftershocks of variable magnitude in the same year as the

main shock, and several magnitude 5 s occur in the next 40 years. In contrast, three aftershocks of magnitude 6 or greater followed the Baffin Bay earthquake, but seismicity decreases between 1950 and 1975. Despite these differences, both series show overall decay for several decades that seems to continue today. This similarity presumably reflects the fact that the main shocks occurred in similar tectonic settings, have comparable magnitude, and occurred only four years apart. It thus seems likely that both events were earthquakes.

[12] A resulting question is where to map the hazard from similar earthquakes. Figure 4 illustrates this issue by comparing hazard maps for Canada made in 1985 and 2005. The older map shows concentrated high hazard bull's-eyes at the sites of the Grand Banks and Baffin Bay earthquakes, assuming there is something especially hazardous about these locations. The alternative is to assume that similar earthquakes can occur anywhere along the margin, possibly on faults remaining from the most recent phase of rifting. This possibility seems more plausible geologically, and is suggested by the seismicity between the Grand Banks and Baffin Bay, some of which may be aftershocks of prehistoric earthquakes. The 2005 map makes this assumption, and thus shows a "ribbon" of high hazard along the coast, while retaining the bull's-eyes. The same issues apply to the U.S. coast, where present maps do not consider offshore events.

## 3. Onshore Earthquake Hazard

[13] The hazard due to onshore continental margin earthquakes is illustrated by the 60 deaths caused by the
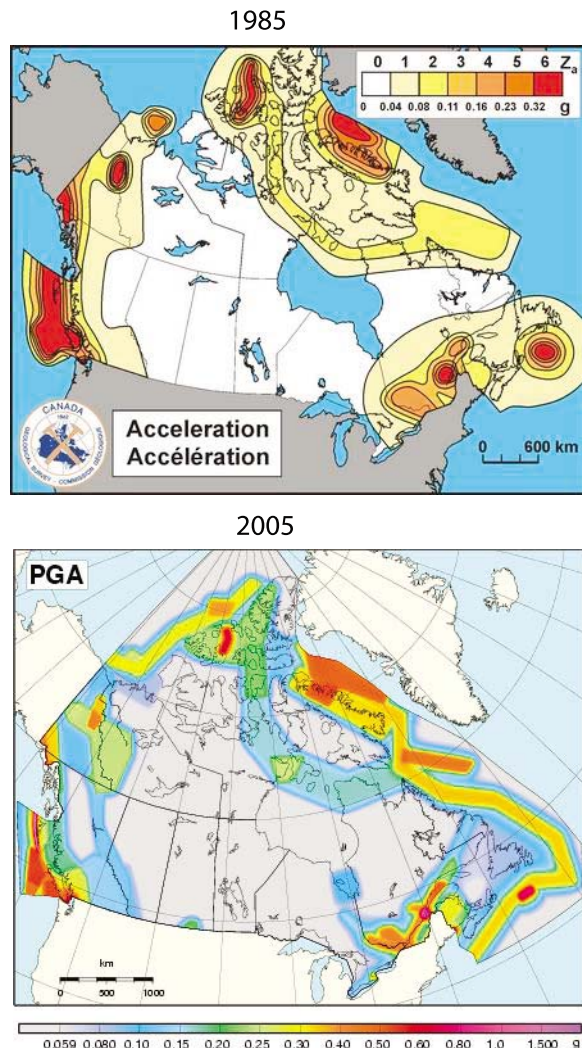
**Figure 4.** Comparison of the 1985 and 2005 Geological Survey of Canada hazard maps. The older map shows concentrated high hazard bull's-eyes along the east coast at the sites of the 1929 Grand Banks and 1933 Baffin Bay earthquakes, whereas the new map assumes that similar earthquakes can occur anywhere along the margin.

Charleston earthquake and the damage caused by the 2011 Virginia earthquake. A major challenge in assessing this hazard is that the tectonic settings and causes of such earthquakes are unclear.

[14] The focal mechanism and aftershock locations (R. Hermann, personal communication, 2011) for the Virginia earthquake are consistent with reverse faulting on a SE-dipping NE-SW striking fault (Figure 5). The fault trend is roughly parallel to the margin, mapped structures in the Virginia Piedmont [*Hughes*, 2011], and the Stafford fault system [*Mixon and Newell*, 1977]. However, the earthquake occurred on the northern edge of the central Virginia seismic zone (CVSZ), a seismic trend normal to the fault plane, margin, and associated structures, that has no presently recognized geologic or geomorphic expression. Moreover, there is no obvious topography related to faster rock uplift or differential deformation in this or nearby seismic zones than in adjacent areas that appear largely aseismic.

[15] This observation suggests that the CVSZ and similar seismic zones along the margin may not be long-lived concentrations of deformation. Instead, they may be the recent loci of seismicity that migrates [*Stein et al.*, 2009], or aftershock zones of large prehistoric earthquakes. It has been suggested [*Sykes*, 1978] that seismicity correlates with extensions of Atlantic Ocean fracture zones. However, the larger subsequent earthquake location dataset indicates that this correlation is weak (Figure 6).

## 4. Causes

[16] The forces driving the seismicity are also unclear. On average, stress indicators for the eastern U.S. show compression oriented ENE [*Zoback*, 1992]. This direction is similar to that predicted by models of intraplate stress due to platewide forces including "ridge push" caused by cooling oceanic lithosphere [*Richardson et al.*, 1979], mantle flow beneath the continent [*Forte et al.*, 2007], and combinations of these and other topographic forces (A. Ghosh and W. E. Holt, Plate motions and stresses from global dynamic models, submitted to *Science*, 2011). The observation that seismicity occurs especially along the margins suggests a contribution from local mechanisms causing approximately margin-normal localized stresses [*Stein et al.*, 1989]. These include "spreading" of lower density continental crust over oceanic lithosphere [*Bott*, 1971], the load of offshore sediment [*Walcott*, 1972; *Turcotte et al.*, 1977; *Cloetingh et al.*, 1983], and stresses due to the removal of glacial loads [*Stein et al.*, 1979, 1989; *Quinlan*, 1984].

[17] The possible role of deglaciation in triggering seismicity by perturbing the background stress state has been a subject of interest because two of the largest passive margin earthquakes, 1929 Grand Banks and 1933 Baffin Bay, occur along the deglaciated coast. A challenge in assessing this effect has been that the predicted rates of glacial isostatic adjustment (GIA) and thus the area over which this effect is significant depend crucially on the assumed viscosity structure. For example, *Wu and Johnston* [2000] find that deglaciation may be significant for earthquakes in the St. Lawrence valley. However, because GIA effects decay rapidly away from the ice margin, they should have little effect in the New Madrid area, unless the viscosities of the crust and upper mantle there are an order of magnitude weaker than surrounding areas [*Grollimund and Zoback*, 2001], which seems not to be the case [*McKenna et al.*, 2007].

[18] The availability of GPS data has recently made it possible to map the rate of GIA [*Sella et al.*, 2007]. The results (Figure 7) show that these motions are small south of the "hinge line" (approximately at the latitude of the Great Lakes) separating uplift to the north from subsidence to the south. In general, seismic moment release decreases southward along the margin, consistent with the variation in vertical motion rates observed by GPS, suggesting that north of the hinge line GIA is an important contributor to intraplate seismicity.

[19] However, south of the hinge line, other stress sources should be more significant. The occurrence of large earthquakes and on other margins that have not been recently glaciated (Figure 1) indicates that GIA cannot be the only mechanism at work. A similar conclusion emerges from geological observations of other vertical motions. In particular, in the mid-Atlantic region deformed stratigraphic and
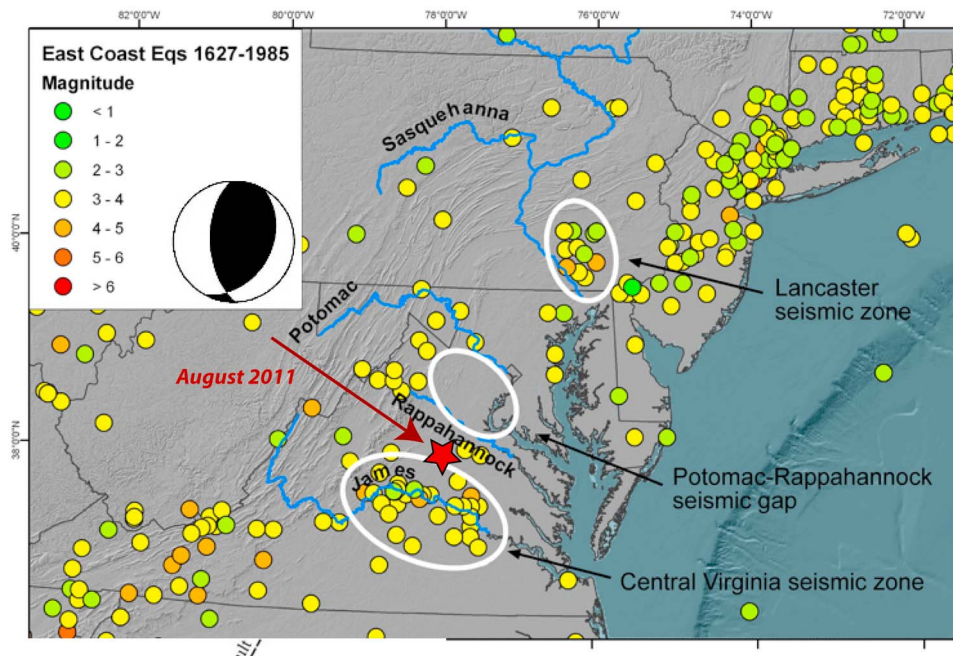
**Figure 5.** Location of the Mineral earthquake and regional seismicity. Focal mechanism from USGS.

geomorphic markers, localized high-relief topography, and rapid river incision show uplift of the Piedmont and Appalachians relative to the Coastal Plain for the past 10 Ma, suggesting that the seismicity reflects active and long-term deformation [*Pazzaglia and Gardner*, 1994, 2000; *Pazzaglia et al.*, 2010]. These motions may reflect dynamic topography resulting from mantle flow [*Moucha et al.*, 2008].

## 5. Challenges

[20] The Mineral earthquake illustrates our limited knowledge of passive margin earthquakes. The issue is how forces that we do not understand cause motion on faults that have not been identified and hazards that we cannot easily assess.

[21] First, we incompletely understand the earthquakes' geologic setting. Most have not yet been associated with specific structures whose geological and paleoseismic history can be studied. Similarly, because these earthquakes are relatively rare, the instrumental seismic record is generally inadequate to map long-lived faults. Thus most of what we know comes from studying the fault planes and aftershock geometries of recent large magnitude earthquakes, which may give little insight into future ones.

[22] Second, it is hard to assess the recurrence of large passive margin earthquakes. Except in the Charleston area [*Talwani and Schaeffer*, 2001], little paleoseismic data exist [*Wheeler*, 2006]. As a result, one can only use regional frequency-magnitude data.

[23] Third, we are unclear about the force systems loading faults and causing motion on them. Although various models predict stress directions generally consistent with the inferred stress field, it is difficult to discriminate between models.

[24] As a result, hazard mappers must chose between drawing high-hazard bull's-eyes at the locations of past earthquakes, which are often not useful predictors of future hazard [*Swafford and Stein*, 2007; *Liu et al.*, 2011; *Stein et al.*, 2011], or mapping a uniform regional hazard. As Figure 4 shows, the different assumptions yield quite different maps, and a long time will be needed to see which was more useful. As a result, developing a cost-effective mitigation policy is a major challenge.

[25] Progress on these issues seems most likely to come slowly via an integrated approach using various techniques. High-precision geodesy can resolve crustal motions as slow as 1 mm/yr [*Calais and Stein*, 2009]. GPS and InSAR studies will thus be able to identify regions where resolvable strain is accumulating to be released in future earthquakes, or show that any strain accumulation is much slower.
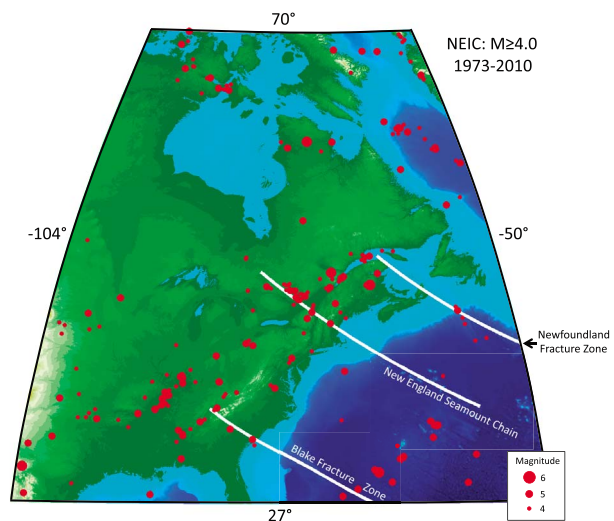


**Figure 6.** Seismicity of eastern North America compared to extensions of fracture zones proposed as controlling mechanism by *Sykes* [1978].
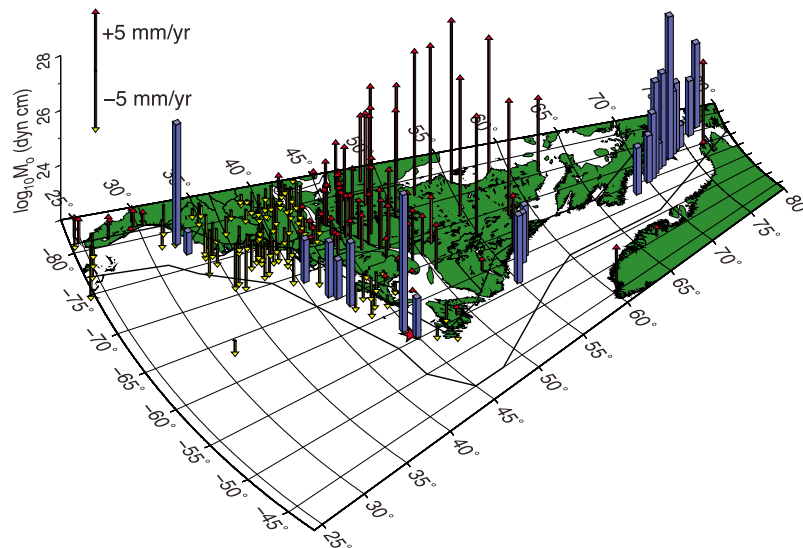
**Figure 7.** Seismic moment release (1568–2003) from *Schulte and Mooney* [2005] (blue bars) compared to vertical GPS velocities (red and yellow arrows) [*Sella et al.*, 2007].

Enhancing regional permanent seismological networks and the EarthScope program's USArray will provide better data on large earthquakes and the locations of microseismicity. Seismological, gravity, and magnetic data can be used to identify and map buried and potentially seismogenic faults. Geomorphology and stratigraphy can constrain the long-term deformation of geomorphic and stratigraphic markers, and paleoseismic studies may succeed in identifying the locations of past earthquakes. Once enough data are compiled to develop a space-time history of seismicity, modeling studies can explore the dynamics of the fault systems [*Li et al.*, 2009].

[26] A natural question to ask is whether, given these challenges, passive margin earthquakes are a minor curiosity about which research is unlikely to yield results commensurate with the effort involved. Our sense is that the problem is worthy of study precisely because we do not understand how large earthquakes occur where idealized plate tectonics predicts they should not. The eastern U.S.'s population density and the need to expand the nation's energy portfolio, perhaps via offshore drilling and further development of nuclear power, make understanding continental margin earthquakes even more significant.

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C. Berti, A. Meltzer, and F. Pazzaglia, Department of Earth and Environmental Sciences, Lehigh University, Bethlehem, PA 18015, USA.

A. Kafka, Weston Observatory, Department of Earth and Environmental Sciences, Boston College, Weston, MA 02493, USA.

S. Stein and E. Wolin, Department of Earth and Planetary Sciences, Northwestern University, Evanston, IL 60208, USA. (seth@earth.northwestern.edu)