frontiers in
EARTH SCIENCE

OPINION ARTICLE
published: 19 August 2014
doi: 10.3389/feart.2014.00020

Pulverization provides a mechanism for the nucleation of earthquakes at low stress on strong faults

Karen R. Felzer*

U. S. Geological Survey, Pasadena, CA, USA
**Correspondence: kfelzer@usgs.gov*

Edited by:
Jean-Philippe Avouac, California Institute of Technology, USA

Reviewed by:
Harsha S. Bhat, CNRS, France
Mai-Linh Doan, Université de Grenoble-Alpes, France

Keywords: earthquake forecasting, san andreas heat flow anomaly, aftershocks, pulverization, time predictable model

An earthquake occurs when rock that has been deformed under stress rebounds elastically along a fault plane (Gilbert, 1884; Reid, 1911), radiating seismic waves through the surrounding earth. Rupture along the entire fault surface does not spontaneously occur at the same time, however. Rather the rupture starts in one tiny area, the rupture nucleation zone, and spreads sequentially along the fault. Like a row of dominoes, one bit of rebounding fault triggers the next. This triggering is understood to occur because of the large dynamic stresses at the tip of an active seismic rupture. The importance of these crack tip stresses is a central question in earthquake physics. The crack tip stresses are minimally important, for example, in the time predictable earthquake model (Shimazaki and Nakata, 1980), which holds that prior to rupture stresses are comparable to fault strength in many locations on the future rupture plane, with bits of variation. The stress/strength ratio is highest at some point, which is where the earthquake nucleates. This model does not require any special conditions or processes at the nucleation site; the whole fault is essentially ready for rupture at the same time. The fault tip stresses ensure that the rupture occurs as a single rapid earthquake, but the fact that fault tip stresses are high is not particularly relevant since the stress at most points does not need to be raised by much. Under this model it should technically be possible to forecast earthquakes based on the stress-renewaql concept, or estimates of when the fault as a whole will reach the critical stress level, a practice used in official hazard mapping (Field, 2008). This model also indicates that physical precursors may be present and detectable, since stresses are unusually high over a significant area before a large earthquake.

The fact that fault tip stresses are high is critical in a second earthquake model, however, which argues that stress conditions over the the fault are in fact generally much lower than static fault strength. Failure occurs only because of the high fault tip stresses followed by the drop in fault strength that is known to occur at fast sliding speeds (Di Toro et al., 2011). In this case there would not be any special conditions over the fault before rupture, making the prediction of large earthquakes practically impossible. Likewise because the whole fault plane does not need to be at any particular stress before rupture, the time-predictable stress renewal model would not be effective. That is, with rupture time not closely tied to fault stress, a long quiescence since the last earthquake would not foretell an imminent quake. Because stress over most of the fault is initially far below failure strength special processes to decrease the stress/strength ratio at the nucleation point would likely be required to enable rupture initiation. As will be explained below, although this second model is more complicated, it is overwhelmingly supported by the observed data.

The observed data shows that there is compelling evidence both that seismogenic faults are strong and that the average static stress on faults at the time of rupture is far below this strength. Borehole measurements adjacent to faults, for example, have shown a critically stressed crust with hydrostatic pore pressure and friction coefficients of 0.6–0.9 (Townend and Zoback, 2000), indicating that the total strength on well oriented faults should be on the order of 50–100 MPa. Multiple lines of evidence indicate, however, that deviatoric shear stress resolved onto the fault plane at the time of rupture is rarely more than 10 MPa. This evidence includes the lack of a heat flow anomaly around the San Andreas Fault (Lachenbruch and Sass, 1980; Fulton et al., 2004) which cannot be explained by fluid heat transport (Fulton et al., 2004), lack of heat flow anomalies around other major faults (Kano et al., 2006), slip striation rotations on mainshock fault planes that show that the initial stress was low (Spudich et al., 1998), regional rotations of focal mechanisms after large earthquakes that limit the amount of stress that could have been on the surrounding faults prior to the mainshock (Hardebeck and Hauksson, 2001; Hasegawa et al., 2011), and self healing pulses during rupture which require low stress/high strength conditions (Heaton, 1990; Noda et al., 2009). Direct borehole measurements of resolved shear stress along active faults have also found low values at the time of measurement (Zoback and Healy, 1992; Brudy et al., 1997). Furthermore, while there are some outliers (Allman and Shearer, 2009), earthquake stress drops generally fall in the range of 1–10 MPa (Abercrombie and Leary, 1993). This would tell us little if each earthquake released a small portion of the total stress on the fault, but observational seismic (Michael et al., 1990; Beroza and Zoback, 1993; Hasegawa et al., 2011) and borehole studies (Barton and Zoback, 1994) indicate that earthquake stress drops are near complete. Thus the 10 MPa maximum usually seen in stress drops and the 10 MPa maximum usually seen in more direct measures of fault prestress are not an accidental coincidence; both are measures of the average fault pre-stress at rupture initiation.

In order for a low fault stress/high fault strength model to be completely satisfactory, however, we must be able to explain how earthquakes nucleate. I propose that the nucleation results from a multi-step process that starts with high frequency acceleration coming from the fault tip of a previous earthquake. The culmination of this process is a small area of severely weakened fault which allows for local stress/strength parity. While this hypothesis, in supporting the low fault pre-stress model, indicates that the stress renewal model cannot work, it suggests that the areas of most likely earthquake occurrence can be forecast because many earthquakes will nucleate near triggering earthquakes that occurred recently. This concept has been borne out experimentally in the results of the prospective 5 year RELM California earthquake forecasting experiment (Schorlemmer et al., 2010). In contrast, repeated efforts to use variations of the time predictable model, including the Parkfield earthquake experiment (Bakun and McEvilly, 1984) which was officially sanctioned by the U.S. Geological Survey, and various seismic gap global forecasts, have met with repeated failure (Rong et al., 2003; Jackson and Kagan, 2006).

The key that high frequency acceleration may be important in earthquake nucleation comes from their unique presence before the nucleation of most aftershocks. Types of events that are depleted in high frequency energy, such as aseismic events and nuclear test blasts (Hough and Anderson, 1989) produce relatively few aftershocks (Pollitz and Johnston, 2006; Llenos et al., 2009; Parsons and Velasco, 2009), even if the accompanying static stress change and low to intermediate frequency dynamic stress change, respectively, are comparable to tectonic events. Another compelling observation is that total aftershock productivity scales linearly with mainshock area, or with mainshock magnitude M as 10^M (Felzer et al., 2002). Total high frequency energy release scales the same way (Hanks and McGuire, 1981; Boatwright, 1982; Zeng et al., 1994), but static and lower frequency dynamic stress changes do not. The decay of high frequency acceleration energy with distance from the mainshock fault plane also matches the decay of aftershock density with distance from the fault plane, a feature which is again not matched by the decay of static stress change amplitude (Gomberg and Felzer, 2008). There may be concern that high frequency acceleration decays too rapidly with distance from the mainshock to trigger aftershocks at distances out to 50 km, where they have been observed to occur (Felzer and Brodsky, 2006). Rapid attenuation of high frequencies does occur at the surface, but attenuation is much slower at seismogenic depths. Frequencies up to 200 Hz have been observed 100 km away from earthquakes in borehole seismometers (Leary, 1995).

The unique power of high frequency acceleration is that above 150/s it can pulverize cracked rock; higher frequency acceleration may pulverize solid rock (Doan and Gary, 2009). Pulverization occurs at the point when energy enters the system too rapidly to be supported by the growth of a limited number of cracks. Whether or not this point is reached is a function of the frequency of the incoming energy and the geometry of existing cracks, with the amplitude apparently playing a minimal role (Doan and Gary, 2009). Near a mainshock fault the amount of energy is very high, and pulverization has been observed at the surface and subsurface near major faults (Doan and Gary, 2009; Mitchell et al., 2011; Wechsler et al., 2011). Wide swaths of pulverization that can be readily recognized in the field may require extreme events, but recent nanometer-scale observations have demonstrated that the thin concentrated veins of fault gouge that are universally found within the principal slip zone (PSZ) of earthquakes contain particles so small that they are consistent with creation by pulverization, not abrasion (Sammis and Ben-Zion, 2008). Other evidence from exposed earthquake faults in mines and in the field further support the idea that pulverization is central to creating the gouge in the PSZ (Wilson et al., 2005). This indicates that pulverization may be a routine feature of earthquake occurrence. There is no direct evidence that pulverization occurs outside of the narrow PSZ for most earthquakes or outside of the near fault zone for more extreme events. The measured existence of frequencies at depth high enough to pulverize out to 100 km, however, suggests that pulverization could occur at depth out to these distances in optimally cracked and oriented locations. Pulverization would not be expected to be common at these distances, but we know that with distance from the fault plane the incidence of aftershocks also drops of rapidly. Furthermore, as noted above, the region that needs to be weakened to nucleate an earthquake can be very small. Therefore away from the immediate fault plane region pulverization only needs to occur at sparse and tiny optimal locations for the mechanism to be viable. Once an earthquake occurs the nucleation site becomes part of the PSZ, so there is no way to prove from field observations that a part of the pulverization occurred during the nucleation phase. My primary argument is that the occurrence of pulverization at the hypocenter of every aftershock is entirely possible and may be further investigated with modeling studies.

When fresh gouge is created by pulverization the direct effect is to change the frictional regime from velocity weakening to velocity strengthening (Beeler et al., 1996). Velocity strengthening areas that are above the brittle/ductile transition are technically in a regime of "brittle creep" (Perfettini and Avouac, 2004). When these areas are subject to static or long period dynamic stress change they experience a sharp increase in velocity (Perfettini and Avouac, 2004). Laboratory results further show that after several to 10 mm of slip, an amount of afterslip that can accumulate rapidly after an earthquake (Hsu et al., 2002; Freed, 2007), slip localization will occur within fresh gouge (Beeler et al., 1996). This localization caused a 2–3 times reduction in fault strength in an experimental surface subjected to 2.4 MPa of normal stress (Reches and Lockner, 2010); under more realistic higher normal stresses the strength reduction should be even higher, if experiments with strength reduction at larger slip speeds serve as a guide (Di Toro et al., 2011). The loss of strength is accompanied by a shift back to velocity weakening friction (Beeler et al., 1996). Thus the area is both weakened and returned to the proper frictional regime for earthquake nucleation. The pulverized, weakened point may now be pushed to final failure by static pressure from the fault tip, rapidly becoming part of the same earthquake, or by surrounding afterslip, occurring later as part of the formal aftershock sequence (Perfettini and Avouac, 2004), or by later fluid flow, slow earthquake slip, a long period wave arrival from a distant earthquake, or other stress transferring event. In the latter case the event could appear to be an independent earthquake.

It is important to note that while the above hypothesis nullifies the time predictable model as commonly presented it does not completely negate the concept of a seismic cycle. Because stress drop is expected to be near complete, fault segments that rupture should require some minimum recovery time before they can rupture again, and this has been observed in the field (Rubin et al., 1999; McGuire, 2008). Once a minimum recovery time has passed, however, there will be a long period of time during which fault stress is far below fault strength, but the fault may fail if pulverization occurs at one tiny point. Hence the wait time between ruptures becomes the sum of a predictable recovery time and an unpredictable fault interaction time, in agreement with observation that repeat times between paleoseismic ruptures is not periodic but is also not entirely random (Scharer et al., 2010).

ACKNOWLEDGMENTS

I am grateful for helpful reviews from Morgan Page, Jeanne Hardebeck, Nancy King, Kate Scharer, Sue Hough, John Townend, and the two reviewers, and for the support of my family.

REFERENCES

Abercrombie, R., and Leary, P. (1993). Source parameters of small earthquakes recorded at 2.5 km depth, cajon pass, southern California: implications for earthquake scaling. *Geophys. Res. Lett.* 20, 1511–1514. doi: 10.1029/93GL00367

Allman, B. P., and Shearer, P. M. (2009). Global variations of stress drops for moderate to large earthquakes. *J. Geophys. Res.* 114:B01310. doi: 10.1029/2008JB005821

Bakun, W. H., and McEvilly, T. V. (1984). Recurrence models and Parkfield, California, earthquakes. *J. Geophys. Res.* 89, 3051–3058. doi: 10.1029/JB089iB05p03051

Barton, C. A., and Zoback, M. D. (1994). Stress perturbations associated with active faults penetrated by boreholes: possible evidence for near-complete stress drop and a new technique for stress magnitude measurement. *J. Geophys. Res.* 99, 9373–9390. doi: 10.1029/93JB03359

Beeler, N. M., Tullis, T. E., Blanpied, M. L., and Weeks, J. D. (1996). Frictional behavior of large displacement experimental faults. *J. Geophys. Res.* 101, 8697–8715. doi: 10.1029/96JB00411

Beroza, G. C., and Zoback, M. D. (1993). Mechanism diversity of the Loma Prieta aftershocks and mechanics of mainshock-aftershock interaction. *Science* 259, 210–213. doi: 10.1126/science.259.5092.210

Boatwright, J. (1982). A dynamic model for far-field acceleration. *Bull. Seis. Soc. Am.* 72, 1049–1068.

Brudy, M., Zoback, M. D., Fuchs, K., Rummel, F., and Baumgartner, J. (1997). Estimation of the complete stress tensor to 8 km depth in the [ktb] scientific drill holes: implication for crustal strength. *J. Geophys. Res.* 102, 18453–18475. doi: 10.1029/96JB02942

Di Toro, G., Han, R., Hirose, T., Paola, N. D., Nielsen, S., Mizoguchi, K., et al. (2011). Fault lubrication during earthquakes. *Nature* 471, 494–499. doi: 10.1038/nature09838

Doan, M.-L., and Gary, G. (2009). Rock pulverization at high strain rate near the San Andreas fault. *Nat. Geosci.* 2, 709–712. doi: 10.1038/ngeo640

Felzer, K. R., Becker, T. W., Abercrombie, R. E., Göran Ekström, and Rice, J. R. (2002). Triggering of the 1999 M_W7.1 Hector Mine earthquake by aftershocks of the 1992 M_W7.3 Landers earthquake. *J. Geophys. Res.* 107:2190. doi: 10.1029/2001JB000911

Felzer, K. R., and Brodsky, E. E. (2006). Decay of aftershock density with distance indicates triggering by dynamic stress. *Nature* 441, 735–738. doi: 10.1038/nature04799

Field, E. H. (2008). Overview of the working group for the development of Regional Earthquake Likelihood Models (RELM). *USGS Open File Report* 78, 7–16. doi: 10.1785/gssrl.78.1.7

Freed, A. M. (2007). Afterslip (and only afterslip) following the 2004 Parkfield, California earthquake. *Geophys. Res. Lett.* 34:L06312. doi: 10.1029/2006GL029155

Fulton, P. M., Saffer, D. M., Harris, R. N., and Bekins, B. A. (2004). Re-evaluation of heat flow data near parkfield, ca: evidence for a weak San Andreas fault. *Geophys. Res. Lett.* 31:L15S15. doi: 10.1029/2003GL019378

Gilbert, G. K. (1884). A theory of earthquakes of the Great Basin, with a practical application. *Am. J. Sci* 27, 49–53. doi: 10.2475/ajs.s3-27.157.49

Gomberg, J., and Felzer, K. (2008). A model of earthquake triggering probabilities and application to dynamic deformations constrained by ground motion observations. *J. Geophys. Res.* 113:B10317. doi: 10.1029/2007JB005184

Hanks, T. C., and McGuire, R. K. (1981). The character of high frequency strong ground motion. *Bull. Seis. Soc. Am.* 71, 2071–2095.

Hardebeck, J. L., and Hauksson, E. (2001). Crustal stress field in southern California and its implications for fault mechanics. *J. Geophys. Res.* 106, 21859–21882. doi: 10.1029/2001JB000292

Hasegawa, A., Yoshida, K., and Okada, T. (2011). Nearly complete stress drop in the 2011 mw 9.0 off the Pacific coast of Tohoku earthquake. *Earth Planets Space* 63, 703–707. doi: 10.5047/eps.2011.06.007

Heaton, T. H. (1990). Evidence and implications of self-healing pulses of slip in earthquake rupture. *Phys. Earth Planet. Int.* 64, 1–20. doi: 10.1016/0031-9201(90)90002-F

Hough, S. E., and Anderson, J. G. (1989). Attenuation in western Nevada: preliminary results from earthquakes and explosion sources. *Geophys. Res. Lett.* 16, 207–210. doi: 10.1029/GL016i002p00207

Hsu, Y.-J., Bechor, N., Segall, P., Yu, S.-B., Kuo, L.-C., and Ma, K.-F. (2002). Rapid afterslip following the 1999 Chi-Chi, Taiwan earthquake. *Geophys. Res. Lett.* 29, 1–4. doi: 10.1029/2002GL014967

Jackson, D. D., and Kagan, Y. Y. (2006). The 2004 Parkfield earthquake, the 1985 prediction, and characteristic earthquakes: lessons for the future. *Bull. Seis. Soc. Am.* 96, S397–S409. doi: 10.1785/0120050821

Kano, Y., Mori, J., Fujio, R., Ito, H., Yanagidani, T., Nakao, S., et al. (2006). Heat signature on the chelungpu fault associated with the 1999 Chi-Chi, Taiwan, earthquake. *Geophys. Res. Lett.* 33:L14306. doi: 10.1029/2006GL026733

Lachenbruch, A. H., and Sass, J. H. (1980). Heat flow and energetics of the San Andreas Fault zone. *J. Geophys. Res.* 85, 6185–6222. doi: 10.1029/JB085iB11p06185

Leary, P. C. (1995). The cause of frequency-dependent seismic absorption in crustal rock. *Geophys. J. Int.* 122, 143–151. doi: 10.1111/j.1365-246X.1995.tb03542.x

Llenos, A. L., McGuire, J. J., and Ogata, Y. (2009). Modeling seismic swarms triggered by aseismic transients. *Earth Planet. Sci. Lett.* 281, 59–69. doi: 10.1016/j.epsl.2009.02.011

McGuire, J. J. (2008). Seismic cycles and earthquake predictability on East Pacific Rise transform faults. *Bull. Seis. Soc. Am.* 98, 1067–1084. doi: 10.1785/0120070154

Michael, A. J., Ellsworth, W. L., and Oppenheimer, D. H. (1990). Coseismic stress changes induced by the 1989 Loma Prieta, California earthquake. *Geophys. Res. Lett.* 17, 1441–1444. doi: 10.1029/GL017i009p01441

Mitchell, T. M., Zion, Y. B., and Shimamoto, T. (2011). Pulverized fault rocks and damage asymmetry along the arima-takatsuki tectonic line, japan. *Earth Planet. Sci. Lett.* 308, 284–297. doi: 10.1016/j.epsl.2011.04.023

Noda, H., Dunham, E. M., and Rice, J. R. (2009). Earthquake ruptures with thermal weakening and the operation of major faults at low overall stress levels. *J. Geophys. Res.* 114:B07302. doi: 10.1029/2008JB006143

Parsons, T., and Velasco, A. A. (2009). On near-source earthquake triggering. *J. Geophys. Res.* 114:B10307. doi: 10.1029/2008JB006277

Perfettini, H., and Avouac, J. P. (2004). Postseismic relaxation driven by brittle creep: a possible mechanism to reconcile geodetic measurements and the decay rate of aftershocks, application to the Chi-Chi earthquake, Taiwan. *J. Geophys. Res.* 109:B02304. doi: 10.1029/2003JB002488

Pollitz, F. F., and Johnston, M. J. S. (2006). Direct test of static stress versus dynamic stress triggering of aftershocks. *Geophys. Res. Lett.* 33:L15318. doi: 10.1029/2006GL026764

Reches, Z., and Lockner, D. A. (2010). Fault weakening and earthquake instability by powder lubrication. *Nature* 467, 452–456. doi: 10.1038/nature09348

Reid, H. F. (1911). The elastic-rebound theory of earthquakes. *Univ. Calif. Publ. Bull. Dept. Geol. Sci.* 6, 413–444.

Rong, Y. F., Jackson, D. D., and Kagan, Y. Y. (2003). Seismic gaps and earthquakes. *J. Geophys. Res.* 108:2471. doi: 10.1029/2002JB002334

Rubin, A. M., Gillard, D., and Got, J. (1999). Streaks of microearthquakes along creeping faults. *Nature* 400, 635–641. doi: 10.1038/23196

Sammis, C. G., and Ben-Zion, Y. (2008). Mechanics of grain-size reduction in fault zones. *J. Geophys. Res.* 113:B02306. doi: 10.1029/2006JB004892

Scharer, K. M., Biasi, G. P., Weldon, R. J. II., and Fumal, T. (2010). Quasi-periodic recurrence of large earthquakes on the southern San Andreas fault. *Geology* 38, 555–558. doi: 10.1130/G30746.1

Schorlemmer, D., Zechar, J. D., Werner, M. J., Field, E. H., Jackson, D. D., and Jordan, T. H. (2010). First results of the Regional Earthquake Likelihood models experiment. *Pure Appl. Geophys.* 167, 859–876. doi: 10.1007/s00024-010-0081-5

Shimazaki, K., and Nakata, T. (1980). Time-predictable recurrence model for large earthquakes. *Geophys. Res. Lett.* 7, 279–282. doi: 10.1029/GL007i004p00279

Spudich, P., Guatteri, M., Otsuki, K., and Minagawa, J. (1998). Use of fault striations and dislocation models to infer tectonic shear stress during the 1995 hyogo-ken nanbu (kobe) earthquake. *Bull. Seis. Soc. Am.* 88, 413–427.

Townend, J., and Zoback, M. D. (2000). How faulting keeps the crust strong. *Geology* 28, 399–402. doi: 10.1130/0091-7613(2000)28<399:HFKTCS>2.0.CO;2

Wechsler, N., Allen, E. E., Rockwell, T. K., Girty, G., Chester, J. S., and Ben-Zion, Y. (2011). Characterization off pulverized granitoids in a shallow core along the San Andreas Fault, Littlerock, CA. *Geophys. J. Int.* 186, 401–417. doi: 10.1111/j.1365-246X.2011.05059.x

Wilson, B., Dewers, T., Reches, Z., and Brune, J. (2005). Particle size and energetics of gouge from earthquake rupture zones. *Nature* 434, 749–752. doi: 10.1038/nature03433

Zeng, Y., Anderson, G., and Yu, G. (1994). A compromise source model for computing realistic synthetic strong ground motions. *Geophys. Res. Lett.* 8, 725–728.

Zoback, M. D., and Healy, J. H. (1992). *In situ* stress measurements to 3.5 km depth in the Cajon pass scientific research borehole: implications for the mechanics of crustal faulting. *J. Geophys. Res.* 97, 5039–5057. doi: 10.1029/91JB02175

Conflict of Interest Statement: The author declares that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

Received: 14 January 2014; accepted: 28 July 2014; published online: 19 August 2014.

Citation: Felzer KR (2014) Pulverization provides a mechanism for the nucleation of earthquakes at low stress on strong faults. Front. Earth Sci. ***2****:20. doi: 10.3389/feart.2014.00020*

This article was submitted to Structural Geology and Tectonics, a section of the journal Frontiers in Earth Science.