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Creep events slip less than ordinary earthquakes

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[1] We find that slow seismic events have smaller fault slips compared to ordinary earthquakes with similar dimensions. For ordinary earthquakes, the ratio of slip to fault length is largely consistent, yet the physical controls on this ratio are unknown. Recently discovered slow slip or creep events in which faults move quasi-statically over periods of days to years shed new light on this old conundrum. For example, large slow events that extend over 100 km have slips of centimeters, while ordinary earthquakes that rupture a comparable length of fault typically slip meters. The small slips of quasi-static events compared to the large slips of earthquakes show that dynamic processes significantly control the rupture growth of ordinary earthquakes. We propose a model where slip on a heterogeneous fault accounts for the difference. Inertial overshoot may result in larger final slips for earthquakes than comparable creep events. **Citation:** Brodsky, E. E., and J. Mori (2007), Creep events slip less than ordinary earthquakes, *Geophys. Res. Lett.*, 34, L16309, doi:10.1029/2007GL030917.

1. Introduction and Observations

[2] Average slip and rupture length are two of the most fundamental quantities defining the size of an earthquake. Somewhat mysteriously, their ratio has long been observed to be largely independent of size for a wide variety of earthquakes [Kanamori and Anderson, 1975; Scholz, 1990; Abercrombie, 1995]. Any situation in which the slip-rupture length ratio varies systematically would provide a window into a major feature of earthquake rupture.

[3] Slow slip events open up just such a window. We use the terms “slow” or “creep” for events that have durations of days to years. Creep events have been recently shown to be a major feature of tectonic systems [e.g., Fujii, 1993; Linde et al., 1996; Hirose et al., 1999; Dragert et al., 2001; Kostoglodov et al., 2003; Ozawa et al., 2005; Ohta et al., 2006; Wallace and Beavan, 2006]. Enough data now exists that we can meaningfully contrast the creep events with ordinary earthquakes.

[4] The ratio of slip to rupture length is often parameterized with the static stress drop $\Delta\sigma_s$. In order to make a convenient comparison between events of varying geometries, we use the approximate relationship

$$\Delta\sigma_s = \mu D/L \quad (1)$$

where μ is the shear modulus, D is the average slip and L is the square root of the rupture area. Equation 1 is

appropriate for studying relative values of the average properties over the slip surface. It should not be strictly interpreted as stress for a particular geometry, however, it captures the dependence of static stress changes on two key physical variables.

[5] Figure 1 shows a comparison of the static stress drops of ordinary earthquakes and the longer duration slow events. For earthquakes, we use representative datasets to illustrate typical behavior with the usual range of stress drops of 0.1–10 MPa.¹ For the slow events, stress drops were calculated from equation 1 using the determinations of rupture area and average slip in the papers cited in the caption of Figure 1. The slow events include episodic slab creep and transient San Andreas fault creep.

[6] Nine out of the ten well-recorded slow events have static stress drops <0.2 MPa while the mean of the earthquake static stress drops is 2 MPa. The only slow event in the range of regular earthquakes is the 1998 Guerrero event that appeared to slip 1.4 m with $L = 57$ km, based on a single-station observation. A multiple station observation of another slow event in the same region in 2001–2002 yielded an average of 10 cm slip with $L = 275$ km [Kostoglodov et al., 2003].

[7] Earthquakes that generate seismic waves, but rupture more slowly than normal, are called tsunami earthquakes because of their ability to perturb the water column through large seafloor offsets despite relatively low levels of high frequency radiation [Polet and Kanamori, 2000]. The tsunami earthquakes cluster at stress drops near those of ordinary earthquake stress drops, and generally at the low end of the range. The special case of shallow very low frequency earthquakes (VLF) in accretionary wedges seems to be the only group of earthquakes with stress drops substantially below those of the creep events [Ito and Obara, 2006].

[8] The data from the ordinary earthquakes is primarily seismic, while that for the slow slips is primarily geodetic. However, entirely geodetic inversions of moderate seismic events agree with the typical range of earthquake average slips reflected in Figure 1 [Hurst et al., 2000]. In summary, earthquakes that rupture a 100 km fault slip an average of a few meters whereas slow creep events with the same rupture length slip an average of several centimeters. This relationship poses an additional constraint for otherwise successful models of creep events [Liu and Rice, 2005].

2. Possible Mechanisms Based on Previous Work

[9] The order of magnitude difference in stress drops of slow events and ordinary earthquakes may clarify the role of inertia or rapid slip in determining the final slip in

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Figure 1. Static stress drop as a function of event duration. Data are compiled from the literature for slow events [Fujii, 1993; Linde *et al.*, 1996; Hirose *et al.*, 1999; Dragert *et al.*, 2001; Lowry *et al.*, 2001; Kostoglodov *et al.*, 2003; Douglas *et al.*, 2005; Ozawa *et al.*, 2005; Ohta *et al.*, 2006; Wallace and Beavan, 2006], tsunami earthquakes [Kanamori and Kikuchi, 1993; Kikuchi and Kanamori, 1995; Swenson and Beck, 1999; Abercrombie *et al.*, 2001; Ammon *et al.*, 2006] and examples of ordinary earthquakes [Abercrombie and Rice, 2005; Venkataraman *et al.*, 2006]. After-slip is not included in this compilation. A stress drop range is shown for creep events with slip ranges reported. Earthquakes at the typical depth of slab creep events (30 km) may have slightly higher average static stress drops, thus enhancing the difference between earthquakes and creep events [Venkataraman and Kanamori, 2004].

ordinary earthquakes. Dynamic models suggest that transient stresses on a rapidly slipping fault can overshoot the final stress drop by as much as $\sim 40\%$ [Madariaga, 1976]. This difference between the dynamic and static stresses is not enough to explain the data in Figure 1. More exotic dynamic weakening mechanisms that incorporate fluid and thermal effects may be able to produce the large stress differences, however, all of the currently proposed mechanisms require either a slip weakening distance that is on the order of 10's of centimeters or continual slip weakening [Brodsky and Kanamori, 2001; Andrews, 2002; Rice, 2006]. Stress drops being nearly constant across a wide range of magnitudes suggest that the final static stress drop is not controlled by a thermal or fluid dynamic weakening mechanism.

[10] The standard model of rate-state friction predicts velocity-dependent stress drops, especially near the transition from earthquake to creep behavior. The velocity-dependence is due to two separate effects. Both of them are well-illustrated by examining the reduction of the steady-state coefficient of friction for a step increase of velocity. For a change of velocity from V_1 to V_2 ,

$$\Delta\sigma_s = (a - b) \ln \frac{V_1}{V_2} \sigma_n \quad (2)$$

where a and b are laboratory-determined parameters and where σ_n is the normal stress across the fault [Scholz, 1990].

During an earthquake or creep event, V_1 is the pre-event velocity on the fault and V_2 is the velocity during the event. Equation 2 explicitly introduces velocity dependence through the logarithmic term. A second, implicit dependence is embedded in the $a - b$ term. In the rate-state model, earthquakes are a result of a higher absolute value of $a - b$ than creep events. The implicit velocity dependence further increases the difference between earthquakes and creep events in equation 2. Experimental and numerical work support this qualitative analysis by demonstrating that frictional systems with steady elastic forcing and $|a - b| \ll a$ have longer duration, lower amplitude stress oscillations than those with $b \gg a$ [Gu *et al.*, 1984; Scholz, 1990; Baumberger *et al.*, 1999].

[11] Figure 2 shows calculated event durations and stress drops for a slider block model with constant normal stress σ_n and a range of $a - b$. Results are in non-dimensional form with stress drop normalized by σ_n and duration of slip events normalized by recurrence interval, i.e., time between slip events. The stress drops are a strong function of the creep event duration. For reference, the dashed line illustrates duration to the power of $-2/3$. It is clear that the dependence is at least this strong in the creep regime. (The dependence must be different in the earthquake regime as otherwise the stress drops would be greater than the background stress.) In Figure 1, the observed stress drops vary by much

Figure 2. Computed non-dimensional stress drops for a slider-block model with rate-state friction. Dashed line is proportional to duration $^{-2/3}$ (see text). The model follows the standard formulation of similar work where a spring with stiffness k is stretched at a constant velocity V_{sp} and pulls a block that has a normal stress σ_n and frictional coefficient $\mu = \mu_0 + A \ln V/V_0 + B \ln \theta/\theta_0$ where V is the velocity of the block, θ is the state variable and μ_0 , A , B , V_0 and θ_0 are constants [e.g., Gu *et al.*, 1984]. The state variable is evolved with the evolution equation $\dot{\theta} = 1 - \theta V/D_c$ where D_c is a constant. For the results presented here $k = 10^{10}$ Pa/m, $\sigma_n = 10^9$ Pa, $V_{sp} = 10^{-9}$ m/s, $\mu_0 = 0.5$, $V_0 = 10^{-6}$ m/s, $B = 0.065$, $\theta_0 = 5$ s, $D_c = 10$ μ m and $a - b$ varies from -10^{-4} to -10^{-1} . The initial perturbation is a step function in velocity at time 0.

Figure 3. Cartoon illustrating the difference in the stopping threshold and resulting slip distances for the dynamic and quasi-static cases. Earthquakes are dynamic ruptures while creep events are quasi-static.

less than an order of magnitude in the creep regime even though the durations vary by over 3 orders of magnitude. Thus, the rate-state explanation for the relationship between the stress drops and event duration is inconsistent with the data.

[12] The simulations presented have a restricted set of parameters (see Figure 2 caption), but the overall result is general. If the difference between the earthquake and creep event stress drops is a manifestation of rate-state friction, then the trend towards decreasing stress drops should persist throughout the range of creep event durations. Instead, we see a bimodal data set where the earthquakes form one population of stress drops and the quasi-static creep events form another.

[13] Generally slap creep events are deeper than the earthquakes used in Figure 1. The increase of σ_n with depth implies that the stress drops for creep events should also increase. As a result, the difference in stress drops between creep events and earthquakes should be somewhat smaller than the calculated results of the simple model shown in Figure 2.

3. Slip on a Heterogeneous Fault for Earthquakes and Creep Events

[14] We suggest another possible explanation for the stress drop relationship. The stress drop can be controlled by the strength heterogeneity, or roughness of the fault, combined with dynamic overshoot. On a fault with spatially varying strength, once slip initiates at a point, the fault continues to slide until it encounters a strong barrier at that spot (Figure 3). In this scenario, the arrest of slip is locally controlled and is thus consistent with slip pulses. The required level of friction to stop the slip depends on the driving traction. For the case of a dynamic earthquake, the waves can transiently provide extra stress that allows the fault to hop over strong spots. Therefore, the threshold for frictionally stopping slip is higher in the dynamic case than in the quasi-static one. This scenario can be phrased in terms of energy. Slip of the earthquake surface persists until the energy state is in a local minimum. If there is an activation energy available in the form of kinetic energy

in a seismic wave, then the fault can be pushed beyond this local minimum into a lower energy state with a lower final elastic energy at the end of the earthquake.

[15] We will now present a simplified quantitative model for both earthquake and creep event average stress drops on a heterogeneous fault with a prescribed strength distribution. We will assume the earthquake and creep events occur on faults with identical strength distributions and the resulting differences in stress drops are entirely a result of the differences in dynamic behavior. This major assumption is in part justified by the fact that both creep events and earthquakes are observed on the same downgoing slabs in a subduction system. With this assumption, the observed difference in stress drops between slow and ordinary earthquakes can be used to solve for a key parameter of the strength distribution that characterizes the fraction of the fault that is strong.

[16] We model the available stress to drive slip at a point by assuming a stress boundary condition defined by the static stress released σ_h relative to the far-field. For a dynamic rupture, more stress may be available due to transient waves. In general, the available stress to drive slip is $F\sigma_h$ where for dynamic ruptures F is a factor of about 1.2–1.4 based on dynamic models [Madariaga, 1976] and for quasi-static ruptures $F = 1$.

[17] To proceed, we need a statistical model for the distribution of strength on the fault. We use a power law probability distribution for strength σ_T per unit fault length of

$$P(\sigma_T) = C_1 \frac{1}{L} \sigma_T^\alpha \quad (3)$$

where $P(\sigma_T)$ is the probability of exceeding strength σ_T per unit length of fault and C_1 and α are constants. The dependence of equation 3 on the rupture length L means that the fault becomes weaker with distance from the hypocenter. An earthquake nucleates near a strong patch and propagates into a statistically weaker zone. Ultimately, we will use the observations to constrain the parameter α that quantifies the fraction of the fault that is relatively strong.

[18] Any real distribution must have a maximum value of σ_T if $\alpha > 0$ or a minimum value if $\alpha < 0$ in order to ensure that the probability remains finite. The constant C_1 is selected so that the integral of the probability over all stresses is normalized properly to 1 and thus C_1 is a function of the minimum or maximum stress, as appropriate.

[19] The criterion for frictionally stopping slip at a point is $\sigma_T - F\sigma_h \geq 0$. From equation 3, the average distance $\langle D \rangle$ that a point slips before being stopped by a barrier is

$$\langle D \rangle = \frac{1}{P(F\sigma_h)} \propto \frac{L}{F^\alpha \sigma_h^\alpha} \quad (4)$$

The data of Figure 1 imply that for a given rupture length L , dynamic fractures have relative slips that are over an order of magnitude greater than static ones. More specifically, the median values of the earthquake (dynamic) slip and the creep (static) slip in Figure 1 are related by

$$\langle D \rangle_{\text{dynamic}} \approx 20 \langle D \rangle_{\text{static}} \quad (5)$$

Figure 4. Profiles of synthetic fault surfaces in arbitrary distance units with pre-stress distributed following equation 3. The stress is normalized by the minimum value, σ_0 . Note the logarithmic scale on the y-axis.

for a given rupture length L . As discussed above, we assume both the earthquake and creeping zones of a subducting slab have the same roughness, thus the dynamic and static slip occur on faults with the same value of α . Using $F_{\text{dynamic}} = 1.2 - 1.4$ and $F_{\text{static}} = 1$, as discussed above, we combine the model in equation 4 and the observation in equation 5 to find that α is between -9 and -17 . Thus, we have constrained the parameter α of the strength distribution by using the stress drop observations.

[20] The large, negative values of α derived describe a surface that is generally very weak (Figure 4). The surface is more like Figure 4c than 4a. Physically, such values for α describe relatively smooth surfaces with a few isolated strong patches. More specific models for the inertial processes will yield a variety of values of α , but as long as the overshoot is on the order of tens of percent, the resulting distribution will be similarly smooth with strongly negative α . The result will hold as long as the difference between dynamic and creep stress drops is attributable to the intrinsic difference between dynamic and quasi-static failure on a rough surface. Such a smooth surface is consistent with field measurements of actual fault surfaces [Sagy *et al.*, 2007].

4. Summary and Conclusion

[21] In this paper we have distinguished a fundamental aspect of creep events from earthquakes. At the time of submission, this was one of the first uses of the data that separate the phenomena based on something other than slip rate. While this paper was in review, Ide *et al.* [2007] made similar observations, but did not consider stress drops as a useful characterization of the distinction nor did they explore the intrinsic differences between dynamic and quasi-static rupture as part of the mechanism.

[22] We used the seismological convention and referred to the quantity defined by equation 1 as “static stress drop” when in fact it is simply a normalization of the displacement by the rupture length. Our preferred model does not require that this quantity physically represent stress at any point on the fault. However, the interpretation in terms of rate and state friction does imply that equation 1 represents a stress. Perhaps this is another reason that equation 2 does not provide a ready explanation for the observation.

[23] The difference in the ratio of slip to rupture length is a non-trivial result and is not easily matched by existing models of inertial overshoot or rate-state friction near the creep regime. More sophisticated dynamic models incorporating frictional and dynamic effects may predict the behavior, but it will take significant extensions of previous work to accomplish this. For now, the best explanation that we have been able to develop is that the distinction results from the ability of earthquakes to jump over rough patches of a fault due to inertial overshoot. In this model, a relatively smooth fault surface can accommodate both types of stress drops given reasonable inertial overshoots. Undoubtedly future work will produce alternative explanations. These new models will have to respect the main feature of the observation: the dynamic, rapid nature of earthquakes plays an important and quantifiable role in determining their final slip.

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