Effective strength heterogeneity controlling slow slip events on laboratory faults

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Rheological differences on natural faults are believed to be a controlling factor of slow earthquakes. We characterized factors controlling behavior on a PMMA laboratory fault, focusing on effects of normal stress and loading velocity on slip and slip rate evolution building up to gross fault rupture. Faults displayed lower strength heterogeneity at lower normal stresses compared with greater strength and stiffness heterogeneity at high normal stresses. Slowly expanding shear rupture originated from central, weaker and more compliant sections of the fault, propagating outwards at speeds of $C_{\text{slow}}/C_{\text{Rayleigh}} \sim (6-191) \times 10^{-7}$, scaling directly to rates for actual slow earthquakes. Rupture expansion, slip rates and stress drop were dependent on the shearing velocity. Smaller number of larger patches in an area led to regular earthquakes, while larger number of smaller patches led to slow earthquakes. Smoothing of the degree of fault complexity by slow earthquakes prepared the condition for standard breakaway events.
1. Introduction

For more than fifty years seismologists have reported the occurrence of slow earthquakes [Beroza and Ide, 2011]. Slow earthquakes are characterized by relatively long-period emanating seismic waves in relation to the short-period seismic waves that make up traditional earthquake signals [Ide et al., 2007]. Denser earthquake-monitoring networks, which include borehole seismometer arrays and continuously recording global positioning system (GPS) networks, have allowed for more accurate and frequent observations of slow fault slip at a range of depths [Kawasaki et al., 1995; Kostoglodov et al., 2003; Igarashi et al., 2003; Rogers and Dragert, 2003].

A variety of slow earthquake phenomena have been observed, such as, deep episodic non-volcanic tremor [Obara, 2002], low-frequency earthquakes (LFEs) [Shelly et al., 2007], very low-frequency earthquakes (VLFs) [Ito et al., 2007], episodic tremor and slip (ETS) [Rogers and Dragert, 2003], and long-term/short-term slow slip events (SSEs) [Hirose and Obara, 2010; Fukuda et al., 2014]. Some of these phenomena may be linked. For example, tremor and LFEs may be the noisy expression from a small fraction of the fault surface undergoing large-scale slow slip episodes [Bostock et al., 2015] but our mechanical understanding of their interaction is limited. In the field, physical mechanisms causing the differences between the various types of slow earthquake behavior (and their interactions) are not well understood, but seem to be characterized by shear slip along interplate boundaries.

Slow fault slip at shallow depth (above the seismogenic zone, i.e. < 10 km) are still not well understood and reside up-dip from large earthquake ruptures. Less focus has been
paid to shallow events but they are believed to occur along the plate interface and should therefore be controlled by friction properties and conditions characterizing the fault zone [Ikari et al., 2013]. Slow earthquakes at greater depths (below the seismogenic zone) are also thought to be attributed to frictional heterogeneity and asperity interactions [e.g., Ando et al., 2010; Nakata et al., 2011; Zigone et al., 2015]. Obara [2011], among others [e.g., Saffer and Wallace, 2015], gives an intricate picture of the variety of faulting conditions that may interact to cause slow earthquakes. It is thought that slow earthquakes occur on rocks exhibiting appropriate hydrothermal conditions [Okazaki and Katayama, 2015] but to date, there has been no unifying mechanism to describe aseismic transients (slow slip). One thought is that slow earthquakes are prematurely arrested large earthquakes due to frictional barriers [Ikari et al., 2013; Mele Veedu and Barbot, 2016].

We present laboratory experiments that examine features of slow slip accumulation that appear analogous to slow slip events (SSEs) on natural faults. The investigation focuses on the possible mechanical factors causing systematic changes in the evolution of slow slip. Using a direct shear friction apparatus we constrain the boundary and initial conditions then shear a frictional interface formed between two blocks of polymethyl methacrylate (PMMA). Prior to shearing, we measure the clustering of asperities along the interface that represent strength heterogeneity, a property controlled by the topographic interaction between the two samples. Increased density of asperities, representative of local fault fluctuations, changed with increased far-field normal stress (σ₀). Both an increase in overall strength and fluctuations in strength were observed at higher σ₀. This causes an increased resistance to the expansion of a slow shear frictional rupture. At high levels of
applied normal stress, we observe a slowly expanding rupture we refer to as slow events (SEs). These events were not present when strength heterogeneity decreased (low $\sigma_0$). We believe that the generation of slow earthquakes could be controlled by large variations in the plate’s interface shear strength and stiffness on sections of the fault that are not entirely locked (i.e. exhibit some level slow slip beforehand).

2. Experimental procedure

The tests were carried out on a direct shear friction apparatus. Additional details of the experimental facilities, material properties are found in Selvadurai and Glaser [2015a] and surface preparation techniques are in Supporting Information S1. The apparatus (Figure 1(a)) allowed us to monitor transition of sliding (from slow to rapid) along a pre-existing frictional fault. The top slider block had dimensions 400 mm x 80 mm x 12 mm thick and the base plate was 610 mm x 300 mm x 50 mm thick. The sliding material was PMMA (shear modulus $\mu = 2200$ MPa and Poisson’s ratio $\nu = 0.32$) and has been often used as an analog for faulting rock due to its similar constitutive responses and lack of plastic flow at lower applied stresses [e.g., Spetzler, 1978; Petit and Barquins, 1988; Dieterich and Kilgore, 1994; Rubinstein et al., 2004; McLaskey and Glaser, 2011; Svetlizky and Fineberg, 2014; Subramanian and Singh, 2015]. The validity if using PMMA as a ductile rock analog is discussed in Selvadurai and Glaser [2015a]. A pressure sensitive film (20 micron pixel size) was used to map the size, location and normal stress supported by asperities along the interface before each test was performed [Selvadurai and Glaser, 2015a]. Seven noncontact eddy current displacement sensors (NC1-NC7) were used to measure differential slip along-strike ($x$-direction) of the fault.
During an experiment asperities along the interface are measured using the pressure sensitive film [Selvadurai and Glaser, 2015a]. The fault was carefully indexed to a datum location then the pressure film was developed by pressing it in between the slider block into the base plate. Normal loading was accomplished by pressing two hydraulic jacks on the massive, rigid loading platen; inducing a far-field normal stress $\sigma_0$. Normal stress was maintained on the fault for $t_{\text{hold}} = 900$ s allowing film time to develop. The fault was unloaded, and the film was removed, digitized and the signal processed (see Figures 1(b) and (c) for examples of the measured asperity distribution). The slider block was then re-indexed back the datum position and reloaded, without the pressure film, to the same $\sigma_0$. The fault was held for $t_{\text{hold}} = 900$ s then the rigid loading platen was driven at a constant far-field velocity $V_{LP}$ that simulated tectonic loading. Shear force $F_s$ built up and slip $\delta$ (relative motion of the slider block and base plate) variably accumulated along the fault until a gross fault rupture. Apparatus stiffness was calculated to be 0.76 N/$\mu$m from the sample averaged slip versus the shear force dropped during gross fault rupture [McLaskey and Kilgore, 2013]. The compliant nature of the apparatus coupled with low driving velocities allowed us to study the slow premonitory phase in more detail [Heslot et al., 1994].

Slip measurement from the noncontact sensors (NC1-NC7) are shown in Figure 2. These measurements represent local slip which varied non-uniformly in the $x$-direction as slow slip accumulated. Linear interpolation was used to expand the local measurements to a 2D spatiotemporal grid for both slip $\delta(x,t)$ and slip rate $\dot{\delta}(x,t)$ (Figure 3), discretized at 10 mm increments and 1 second time intervals. A total of twenty two run-ups to gross
rupture of the fault were studied at two levels of far-field normal stress: Low ($\sigma_0 = 0.8$ MPa for $F_n = 4.0$ kN) and High ($\sigma_0 = 0.4$ MPa for $F_n = 2.0$ kN). The range of loading rates $V_{LP}$ was $1 \mu m/s$, $3 \mu m/s$ and $7 \mu m/s$. After each fault rupture, the slider was unloaded and re-indexed back to the datum location before the subsequent test.

3. Results

3.1. Asperity distribution at low and high normal stress

The key focus of this study is the manner in which slow slip accumulated along the fault before gross rupture occurred. Our results showed that slow slip accumulated non-uniformly along the fault. We note that differential slip has been observed in the some laboratory friction studies [e.g., Ohnaka and Shen, 1999; McLaskey and Kilgore, 2013; McLaskey et al., 2015], but, we will see, the rupture propagation speeds observed in our study are much slower and may be due to the unique choice of material and apparatus configuration. To further understand the factors controlling slow slip along the fault, we investigated the asperity distributions at different normal load levels. Asperities were formed due to topographic surface variations of the slider block at various length scales (see Supporting Information S2) – local clustering of asperities are the result of local variations in topography [e.g., Archard, 1961; Persson, 1999]. Figures 1(b) and (c) show the asperity distributions at the low ($\sigma_0 = 0.4$ MPa) and high ($\sigma_0 = 0.8$ MPa) levels of applied normal stress, respectively. The spatial distributions are shown for asperities whose individual areas $A_i$ were greater than $4 \times 10^{-3} \ mm^2$. Spatial histograms were segmented into 4 mm increments along-strike ($x$-axis) and 0.5 mm increments across the $y$-axis and showed the non-uniform distribution of the real contact area ($A_r$). Supporting
Information S2 discusses the topographic length scale variations along the slider block sample that can account for variations in asperity distribution (Figures 1(b) and (c)) and real contact area (Figure 1(d)) at various length scales ($\lambda$). Asperities on natural faults also appear at a variety of length scales having been reconstructed using coseismic slip distributions [e.g., Mai and Beroza, 2002; Lavallée and Archuleta, 2005].

For low normal stress ($\sigma_0 = 0.4$ MPa), the total real contact area $A_r = 16.44$ mm$^2$ ($\sim$ 0.53 % of the nominal contact area) was composed of $N = 836$ large asperities. Higher normal stress ($\sigma_0 = 0.8$ MPa) increased the total real contact area to $A_r = 99.26$ mm$^2$ ($\sim$ 2.07 % of the nominal contact area) and was composed of $N = 5040$ large asperities. From Mindlin [1949], we can estimate the elastic shear stiffness $\kappa = \left(8\mu \cdot a/(2 - \nu)\right)$ for a single circular contact of radius $a$ under the no-slip condition. Assuming that $a$ is the equivalent radius for an asperity with area $A_r$, we estimate the equivalent fault shear stiffness as $\kappa = 25.0$ N/$\mu$m at $\sigma_0 = 0.4$ MPa and $\kappa = 61.6$ N/$\mu$m at $\sigma_0 = 0.8$MPa. Figure 1(d) compares the two distributions of real contact area along the $x$-axis. Overall levels of real contact increased with $\sigma_0$ and appear to cluster at length scales $\lambda \sim 50$ mm, explained by coming together of a surface with a topographic “waviness” described in Supporting Information S2. For both high and low normal stress there is a significant amount of contacts at the leading edge (LE) of the fault. Figure 1(e) examines the gradient in real contact area along the $x$-axis at the two levels of normal stress $\sigma_0$. There is more variability along the fault for high normal stress. The results shown here were representative of asperity distributions throughout the initial asperity distributions for the 22 tests at both normal load levels.
3.2. Non-uniform slow slip accumulation

Figure 2 shows slow slip evolution patterns observed along the interface during the suite of experiments performed at low $\sigma_0 = 0.4$ MPa (Figure 2(a)) and $\sigma_0 = 0.8$ MPa (Figure 2(b)). (Both the results in Figures 2 and 3 are shown for an identical loading rate of $V_{LP} = 3 \mu m/s$.) For each normal stress level $\sigma_0$, the friction coefficient $\mu (= F_s/F_n)$, local slip ($\delta$) and local slip rate ($\dot{\delta}$) are shown for all noncontact sensors (NC1-NC7) in descending order. The methodology for slip rate calculations are detailed in Supporting Information S3.

Figure 3 illustrates an example of spatio-temporal slip and slip rate evolution along the fault throughout the loading phase. Time was discretized at one-second intervals and the fault discretized at 10 mm intervals along-strike. The values of slip and slip rate are taken from individual noncontact sensors (Figure 2) and used to interpolate and visualize the spatio-temporal evolution during the loading phase. This was done for low (Figure 3(a)) and high (Figure 3(b)) levels of normal stress $\sigma_0$ allowing us to visualize two distinct patterns in slow slip evolution – arrested slow slip events and sudden breakaway events leading to rupture.

3.3. Variations in slow slip

During all tests, slow slip accumulation initiated from the center of the specimen where the density of asperity contacts were lower (see along-strike histogram in Figures 1(b) and (c)), shown in the spatio-temporal plots of Figure 3. Slow slip evolved in two distinct manners and depended on the far-field normal stress $\sigma_0$. The two types slow slip evolution
patterns are described as: (1) a breakaway event (BE only in Figures 2(a) and 3(a)) and
(2) a slow event followed by a breakaway event (SE + BE in Figures 2(b) and 3(b)).

Slip evolution pattern (1) (BE only) is observed only when the fault was loaded to a
lower applied normal stress ($\sigma_0 = 0.4$ MPa). Nucleation was characterized by an outwardly
growing shear rupture originating from the center of the fault. The slow shear rupture
propagated to both LE and TE edges (i.e. along-strike) at speeds $C_{BE} \sim 5.1$ mm/s
at $V_{LP} = 1 \mu$m/s to $C_{BE} \sim 60.94$ mm/s at $V_{LP} = 10 \mu$m/s. Within the rupture, slip
gradually accelerated and, upon reaching the ends of the fault, gross fault rupture ensued
characterized by an average shear force drop $\Delta F_S \sim 109.5$ N.

Slip evolution pattern (2) (SE+BE) is only observed during tests performed at higher
applied normal stress ($\sigma_0 = 0.8$ MPa). During the SE portion of slip evolution pattern
(2), the shear rupture nucleated from the center of the fault, propagating outwards at
rates between $C_{slow} \sim 0.8$ mm/s at $V_{LP} = 1 \mu$m/s to $C_{slow} \sim 26.10$ mm/s at $V_{LP} = 10$
$\mu$m/s. Upon reaching the ends of the fault, the interface decelerated releasing a partial
shear force drop $\Delta F_{slow} = 12.4$ N (see inset of Figure 2(b) and arrows on slip and slip rate
plots). After the partial stress drop, the fault began reloading itself and was followed by
an ensuing BE similar to that in slip evolution pattern (1). Nucleation of the consequent
BE originated from the central section of the fault and expanded at rates of $C_{BE} \sim 3.2$
mm/s at $V_{LP} = 1 \mu$m/s to $C_{BE} \sim 31.7$ mm/s at $V_{LP} = 10 \mu$m/s. The BE events in
slow slip evolution pattern (2) grew outwardly at slower rates than those observed during
slip evolution pattern (1). All expanding shear ruptures were dependent on the far-field
loading velocity \( (V_{LP}) \) – higher loading velocities resulted in faster expansion of both the SE and BE portion of evolution pattern \((2)\).

### 3.4. Rupture growth during slow events (SE)

Spatio-temporal slip rate plots in Figure 3 are used to examine rupture growth rate \( C_{\text{slow}} \) and duration \( t_{\text{slow}} \) for SEs at various loading rates \( V_{LP} \). The dashed white lines in the slip rate evolution (bottom panels in Figure 3) indicate the region along the \( x \)-direction of the fault where slip rate increased beyond a lower threshold. The threshold for approximating the nucleation region was set to \( \dot{\delta}/V_{LP} \sim 0.02 \). The propagation rate \( C_{\text{slow}} \) was calculated as the slope from the nucleation point to the LE. Results from calculating \( C_{\text{slow}} \) for various loading rates are shown in Supporting Information S4.

Table S1 in Supporting Information S4 shows the propagation rate \( C_{\text{slow}} \) of SEs, duration \( t_{\text{slow}} \) and propagation rate normalized by the Rayleigh wave speed of the material \( (C_{\text{slow}}/C_{R}) \) for each loading rate \( V_{LP} \). The Rayleigh wave velocity for PMMA was \( C_{R} = 1297 \text{ m/s} \). The SEs propagated at speeds of \( C_{\text{slow}}/C_{R} \sim (6-191) \times 10^{-7} \); these speeds are much lower than the slow rupture fronts \( (C_{\text{slow}}/C_{R} \sim 0.05) \) observed during breakaway events in other laboratory direct shear friction experiments [e.g., Rubinstein et al., 2004].

### 3.5. Stress drop (\( \Delta \sigma \)) during slow events (SEs)

It is possible to estimate the amount of stress drop across the expanding slow slip region of the SE in terms of linear elastic fracture mechanics [e.g., Andrews, 1976; Ida, 1972; Svetlizky and Fineberg, 2014]. Stress drop \( \Delta \sigma \) can be determined from the ratio between the slip rate and rupture propagation velocity, which is then scaled by the shear modulus (i.e. \( \Delta \sigma = \mu \cdot (\dot{\delta}/C_{\text{slow}}) \)). We provide an estimate of the stress drop using the
rupture propagation calculations $C_{\text{slow}}$ (calculated before) and estimates of the maximum slip rate during the SE. These values are reported in Table S1. We found that stress drop varied between $\Delta \sigma = 0.25, 0.52,$ and $0.97$ MPa for experiments at loading rates of $V_{LP} = 1, 3$ and $7 \mu m/s$. The stress drop increased as loading rate increased. These may be overestimates of $\Delta \sigma$ since the maximum slip rate during the entire SE was used. The stress drop during these tests may have varied spatially due to fluctuations in the fault shear strength (Figure 1) defined by the fault prestress. With respect to the applied normal stress $\sigma_0$, shear stress drop ($\Delta \sigma$) appears relatively large; however, it is only a fraction (1 to 4%) of the normal stress supported on the asperities, estimated using the pressure sensitive film [$\sigma_{\text{asperity}} \sim 23$ to $32$ MPa from Selvadurai and Glaser, 2015b].

4. Discussion

4.1. Slip evolution pattern and its relation to fault strength

While the underlying mechanism remains elusive, slow earthquakes are believed to observe in situ frictional conditions that promote a transitional state between the stick-slip and steady aseismic sliding regimes [Ikari et al., 2013; Mele Veedu and Barbot, 2016; Leeman et al., 2016]. Part of this transitional state is believed to occur due to elevated pore pressures causing low effective stress or variations in frictional properties that promote slow transient slip [Segall et al., 2010; Liu and Rice, 2005, 2009; Okazaki and Katayama, 2015]. Numerical simulations of slowly propagating slip fronts into regions of frictional heterogeneity [Ando et al., 2010; Nakata et al., 2011; Zigone et al., 2015] have shown similarities to natural observations [e.g., Rogers and Dragert, 2003; Ghosh et al., 2010] but realistic distributions of this heterogeneity are not well known. Saffer and Wallace
[2015] examined frictional properties of a few well known subducting plates (Pacific Plate, Philippine Sea Plate and Cocos Plate) and found that the subduction of the rough seafloor may promote both compositional and geometrical heterogeneity – presenting physical conditions that are likely met on many shallow megathrust faults. In this study we found similarities with SSEs in nature and the SEs described when the fault was loaded to higher applied normal stresses. We specifically observe rupture growth rates that scale directly (via the Rayleigh wave speed) to the rock in nature where slow earthquakes propagate at rates of 1 - 10 km/day ($C_{\text{slow}}/C_R \sim (34-340) \times 10^{-7}$)[Bartlow et al., 2011; Fukuda et al., 2014; Kostoglodov et al., 2003; Obara, 2011; Wech and Bartlow, 2014]. We find that changes in slow slip are a function of locally varying frictional properties, i.e. increased fault strength and evolving complexity. Similar conditions are believed to occur in natural settings [Saffer and Wallace, 2015]. The topography of our faults bodies caused heterogeneous variations in the manner in which asperities form and interact. While pore fluids are not present in our experiments, the number of asperities ($A_r$) and local clustering of asperities became more pronounced when the fault was confined to higher far-field normal stresses $\sigma_0$. This may produce analogous complexity to that observed in nature [Schmittbuhl et al., 2006; Candela et al., 2011; Brodsky et al., 2016]. Combining elastic dislocation with friction theory [Scholz, 2002] gives us insight of how local variations in effective normal stress will proportionally change the local variation in effective fault stiffness ($\kappa \propto \sigma_n$). Variations in local stiffness, controlled by the scale dependent interaction of asperities, appear to control the differences the evolution of slip in this laboratory study. On our laboratory fault, heterogeneous stiffness seems to directly
affect the faults ability to resist shear rupture expansion. Dieterich [1978] described this heterogeneity in terms of effective strength: a property that limits the dimensions of slip and, therefore, affects the complexity of the fault being tested. It is possible that the material used, surface preparation techniques and apparatus imposed boundary conditions allowed for dimensions of slip that mimic more closely a natural settings. The direct scaling of the expanding slow rupture rates may be a realization of up-scaling relationships that require more extensive investigations.

Changes in nominal normal stress $\sigma_0$ affect the near-fault (effective) normal stress along the fault. These stress fields have been found to control shear slip dynamics along frictional faults in the laboratory [e.g., Ohnaka and Shen, 1999; Ben-David et al., 2010; McLaskey and Kilgore, 2013; Leeman et al., 2016]. Shear rupture in these studies have been modeled as an expanding shear crack with a slip-weakening zone controlled by the stress field, which can be determined to establish the transition of sliding from stable to unstable [e.g., Andrews, 1976; Chen and Knopoff, 1986]. Initiation of the shear rupture occurs at the weakest location along the fault [Ohnaka, 1992]. In our experiments, nucleation of slow shear rupture originates near the sparsely populated (Figure 1(d)) center of the fault for both types of slip evolution patterns. This follows standard theory.

Similarities compared to field observations, in the manner that all our experimental ruptures initiate (SEs and BEs) may indicate they share similar mechanisms and are controlled by evolving frictional processes. Recent studies suggests that shallow slow earthquakes may indeed nucleate in a similar manner to regular earthquakes but their expansion is prematurely arrested [Ikari et al., 2013] or may follow a complicated frictional
response that cycles between slow and fast earthquakes [Mele Veedu and Barbot, 2016]. In our experiments the exact cause of rupture arrest is not well known but Saffer and Wallace [2015] note that high complexity persists in shallow subduction zones due to geometric complexity and heterogeneous lithology in natural shallow settings and rheological differences from these factors may be responsible for slow earthquakes at these depths.

4.2. Similarities of laboratory to field studies

Slow earthquakes and non-volcanic tremor migrate at velocities varying between \( \sim 1 \) km/day to \( \sim 10 \) km/day along-strike [Kostoglodov et al., 2003; Obara, 2002; Bartlow et al., 2011; Fukuda et al., 2014; Wech and Bartlow, 2014]. Global positioning system (GPS) networks, have been used to map the spatio-temporal progression of a cohesive zone of a very slowly propagating shear rupture believed to be a slow earthquake. Assuming a Rayleigh wave speed of rock about 3450 m/s, we calculate expansion of a shallow slow earthquake in nature ranges between 34 to 340 \( \times 10^{-7} \cdot C_R \), results comparable to those observed in this study (Table S1). It is possible that the PMMA at these stress levels and experimental boundary conditions create a degree of strength heterogeneity that can provide slow transients that propagate at speed directly consistent to those on faults in the Earth.

Slow earthquakes may have common mechanisms, i.e. an outwardly expanding shear rupture from weak and compliant zones with a breakdown region at the crack-tip [Bartlow et al., 2011; Ikari et al., 2013]. For both low and high normal stress states we observe: (i) the level of premonitory sliding before gross fault rupture is comparable (which is expected if the characteristic slip distance \( D_c \) is independent of the applied normal stress
We also know that the stiffness of the gross sample (from the geometry) and the apparatus remains unchanged between tests – only the interfacial stiffness ($\kappa$) is changed when $\sigma_0$ is increased. Increased shear stiffness is proportional to the distributions of asperities along the fault [Berthoude and Baumberger, 1998]. Complexity, we interpret as large gradients in real contact along-strike, grew with an increase in applied normal stress $\sigma_0$. We believe this could be a factor that allows a fault to hold back rupture, such that expansion speeds and slip rates were similar to natural settings for this configuration and material.

While not all slow earthquakes experience a sudden deceleration, the 2013-2014 shallow slow earthquake (see Figure 4(b)) near the Boso peninsula, Japan [Fukuda et al., 2014], had remarkably similar moment and moment rates to the SEs observed in our experiments Figure 4(a). Fukuda et al. [2014] showed that the Boso SSE exhibited two distinct phases: Phase I and Phase II. The slow event associated with $V_{LP} = 3 \mu$m/s in Figure 4(a) was decomposed into P1 and P2 for visual purposes. The phases P1 and P2 appear visually similar to the Phase I and Phase II described by Fukuda et al. [2014] as shown in Figure 4(b). They cite that the transition between phases Phase 1 and Phase 2 occur due to the sudden acceleration of an expanding rupture from $\sim 1$ km/d or less (Phase 1) to $\sim 10$ km/d (Phase 2) which was also accompanied with an increase in seismicity (blue bars in Figure 4(b)). The acceleration of an expanding rupture is observed in standard earthquake nucleation theory, again promotes the idea that slow and regular earthquakes share similar mechanisms and this further promotes the idea that slow earthquakes are the results of spatial heterogeneity in friction along the fault which can prematurely arrested rupture. In

[Dieterich, 1979]) and (ii) nucleation initiates at similar locations for both SEs and BEs.
Figure 3(b) a black dash-dotted line was added to the slip rate plot to show that $C_{\text{slow}}$ may actually be characterized using multiple propagation velocities (as described by Fukuda et al. [2014]) but more study with better resolution in slip sensor and acoustic emission measurements is necessary. The inversion of GPS data performed by Fukuda et al. [2014] also showed that the faults suddenly underwent and rapid deceleration similar to what was observe experimentally here (Figure 4(a)) – a possible indication of rheologic changes along the fault.

If we are assuming slow and regular earthquakes have similar mechanisms, why do the laboratory slow events (SEs) not result in a gross fault rupture? An explanation may be the geometric length scales of the velocity-weakening (VW) region is surrounded by a velocity-strengthening (VS) zone. Noda and Hori [2014] found that if the critical nucleation size was larger than the VW region where the slow earthquake originated, there may exist fluctuations in the slip velocity (increases followed by a sudden drop) in the later stages of the interseismic cycle. Here, the mechanism is the same for slow and regular earthquakes but rupture is arrested prematurely for the slow. While the extent of our laboratory fault may be limiting certain aspects of our understanding of slow earthquakes, the sudden deceleration of the fault may be better explained as if the slow rupture moved into velocity-strengthening region on the fault – on our fault this was the free edge boundary condition.

From standard frictional stability theory, an increase in effective normal stress $\sigma_0$ should decrease critical nucleation size. One would similarly expect slow slip evolution patterns at both levels of normal stress, each leading to gross fault rupture [Latour et al., 2013].
However, if the increasing $\sigma_0$ varied the rate-and state-dependent parameters A and B [see Noda and Hori, 2014], this could allow for a change in slow slip evolution patterns seen here. It is possible that the sharp gradient in real contact area near the leading edge constitutes a rheological change in frictional properties, imposed here by the experimental facilities, that mimics the movement of shear rupture from a VW to VS regime. More laboratory studies should be conducted since our understanding of asperity distributions and its relationship to frictional properties (in terms or the rate and state variables) are not well understood. A better understanding of asperity distribution, in terms of the the contrasting local rheology, could help better understand the complicated dynamic evolution of slip and conditions leading to slow earthquakes.

5. Conclusion

A fault was sheared in a direct shear configuration in the laboratory. Slow slip was observed to accumulate non-uniformly along the fault before gross fault rupture occurred. For some configurations slow slip stalled before becoming gross rupture. Pressure sensitive film was used to visualize changes in asperity distributions caused by the increasing $\sigma_0$. Distributions of asperities along the shearing interface were a direct indication of the initial bulk stress level along the fault that primarily control frictional processes. During the shearing experiments, two levels of applied normal stress $\sigma_0$ caused distinct (and repeatable) dynamic response slip evolution patterns (1) breakaway events and (2) slow earthquakes leading to breakaway events. Only when higher normal load $\sigma_0$ was applied did we observe slow events – characterized by the increased local slip rates within an expanding shear rupture that culminated in partial stress drop and slip deceleration. Slow
earthquakes nucleated from the central (weaker) section and proceeded to grow towards of the (stiffer) leading edge fault end. Growth occurred at propagation speeds much lower than anything seen in typical laboratory studies $C_{\text{slow}}/C_R \sim (6-191) \times 10^{-7}$ similar to rates observed for shallow slow earthquakes. Upon reaching the end, a partial stress drop was observed, accompanied by the deceleration of the fault.

Breakaway events, which followed slow events, nucleated from the same location but culminated in gross fault rupture. Breakaway events propagated faster than slow earthquakes suggesting that the partial stress drop had weakened the relatively locked leading edge section of the fault sufficiently to allow for the subsequent gross fault rupture. We believe that shallow slow earthquakes may have a similar mechanism to large earthquakes. Slow earthquake may be shear ruptures that nucleate on a velocity-weakening section of the fault; they become suddenly arrested as the shear rupture moves into a large velocity-strengthening section of the fault. SSEs present in the upper lithosphere may be due to the mechanical differences in local fault strength and, while further investigation is necessary, the findings presented here show promise that the study of these controlling features are attainable in the laboratory.

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Figure 1. (a) The direct shear friction apparatus used to study the onset of dynamic sliding along a pre-existing frictional interface. (b) Pressure sensitive film measurements showing the asperity distributions when the interface was loaded to $\sigma_0 = 0.4$ MPa. The real contact area $A_{\text{real}}$ is shown in the spatial histograms along the $x$- and $y$-axis. Locations of the noncontact slip sensors (NC1-NC7) are shown in relation to the interface. (c) The same as (b) with the fault loaded to $\sigma_0 = 0.8$ MPa. (d) Comparison of the real contact area $A_r$ along the fault for the different levels of $\sigma_0$. (e) Comparison in the gradient in real contact area $\nabla A_r$ along the $x$-axis for two levels of $\sigma_0$. 
Figure 2. Experimental results during the preslip phase of the loading cycle. Both interfaces were subjected to a loading rate $V_{LP} = 3 \mu m/s$, but the left-hand side (a) and right-hand side (b) were subjected to normal stresses of $\sigma_0 \sim 0.4$ MPa and 0.8 MPa, respectively. The panels in (a) and (b) show the tribological coefficient of friction along the fault $\mu(t)$, slip, slip rate for individual noncontact sensors in decending order. The thick dashed-dot line is the average slip and slip rate in the respective panel. The arrows in (b) indicate the partial stress drop associated with the termination of the slow event (SE).
Figure 3. The interpolated spatiotemporal slip distribution (top panels) and slip rate (bottom panels) distributions for the loading conditions seen in Figures 2(a) and (b). Two types of slow slip evolution patterns were observed: slow slip evolution pattern (1) (BE) and slow slip evolution pattern (2) (SE+BE). SEs were observed only in tests loaded to higher normal stress $\sigma_0$. Estimates for the nucleation point are shown as stars and the rate at which the slow event grew is indicated by the dashed lines whose slope is $C_{\text{slow}}$. Total slow slip occurs over time $t_{\text{cycle}}$, SE over time $t_{\text{slow}}$ and BE over time $t_b$. 

\begin{itemize}
  \item a) $\sigma_0 = 0.4 \text{ MPa}, V_{\rho} = 3 \text{ \mu m/s}$
  \item a) $\sigma_0 = 0.8 \text{ MPa}, V_{\rho} = 3 \text{ \mu m/s}$
\end{itemize}
Figure 4. (a) Laboratory slow events (SE) from two independent tests at $\sigma_0 = 0.8$ MPa and $V_{LP} = 3 \mu m/s$. P1 and P2 are phases determined (approximately) by the change in average slip rate. (b†) Results of the moment release (left) and moment rate (right) for the SSE near the Boso Peninsula, central Japan, from Dec. 2013 to Jan. 2014 [from Fukuda et al., 2014]. Seismicity per day, shown in blue, increased during Phase II sliding.