Laboratory slow slip events in natural geological materials

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SUMMARY

Slow slip events (SSEs) and other forms of slow and transient fault slip are becoming increasingly recognized as important, due to their influence on seismicity and potential to provide information on fault zone properties. The majority of our current knowledge of slow fault slip has been obtained from geodetic and seismologic field measurements, but laboratory data of slow slip are comparatively scarce. Here, I present the results of laboratory friction experiments conducted on 11 natural samples from major fault zones around the world, all obtained by scientific drilling. The experiments are conducted water saturated, at room temperature and 10 MPa effective normal stress, representative of in-situ conditions on the shallow portions of fault zones from which the samples were recovered and where slow slip is known to occur. A key component of these experiments is shearing at realistically slow driving rates of cm yr\(^{-1}\), accurately simulating tectonic driving rates. In most samples, these cm yr\(^{-1}\) driving rates produce laboratory SSEs, which are instances of accelerating slip accompanied by a stress drop. The peak slip velocities and stress drops measured in these laboratory SSEs are comparable with those of natural SSEs measured or estimated from geodetic data. A strong correlation is observed between reduced pre-SSE velocity and higher peak slip velocity for the entire laboratory SSE data set. In contrast to the velocity data, significant scatter is observed in the percentage stress drop measurements. The source of this scatter can be attributed to samples with a significant expandable clay component, which tend to exhibit larger stress drops. Results of velocity-stepping tests at cm yr\(^{-1}\) rates show a tendency for velocity-weakening friction not observed at higher sliding velocities, and that the materials with lower values of the rate-dependent friction parameter \(a-b\) tend to produce faster SSEs. Critical stiffness analyses within the framework of rate-and-state friction laws show that most of the SSEs observed in this study do not satisfy the condition for slip instability. The SSEs are more consistent with accelerating stable slip, although the stiffness condition allowing such behaviour is not always satisfied. Considering the laboratory SSEs to be accelerating stable slip, I present a conceptual model for their nucleation. Key elements of the model are a healing-dominated departure from steady-state causing partial locking and velocity decrease, followed by a transition to a velocity-dominated phase representing the actual slip event. The model is consistent with observations from geodetic measurements and the experimental observations in this study. In general, characteristics of SSE-producing fault portions such as the ability to strengthen and store elastic strain energy released as stress drops may be expected to enhance coseismic slip from remotely nucleating earthquakes, an effect which may be quite limited but should be investigated further.

Key words: Creep and deformation; Fault zone rheology; Geomechanics; Ocean drilling; Rheology and friction of fault zones; Dynamics and mechanics of faulting.

1 INTRODUCTION

Decades since the first observations suggesting the occurrence of slow slip events (SSEs) on plate-boundary faults (e.g. Nason 1969; Goulty & Gilman 1978; Sacks et al. 1978, 1981), recently improved geodetic and seismologic coverage has lead to a rapid increase in observations of these phenomena over the past \(\sim20\) yr (e.g. Schwartz & Rokosky 2007; Saffer & Wallace 2015; Bürgmann 2018). One key observation is that discrete slow fault slip occurs over a wide range of timescales: from slow slip events (SSEs) having
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2 GEOLOGICAL SETTING OF TESTED FAULT ZONES

2.1 Tohoku, Japan Trench

The Japan Trench is formed by the subduction of the Pacific Plate beneath the North American/Okhotsk Plate at a rate of ∼8.3 cm yr⁻¹ (DeMets et al. 1994). This margin hosted the largely unexpected 2011 Mw 9 Tohoku-Oki earthquake and ensuing tsunami. Based on seismic and tsunami waveform inversions, geodetic data, and repeated bathymetry surveys, this event produced ∼50–80 m of coseismic slip that breached the seafloor (Fig. 1a, Fujiiwara et al. 2011; Ito et al. 2011; Sun et al. 2017a). Although the Japan Trench is best known for the Tohoku earthquake, the region has frequently experienced large earthquakes over the last ∼100 yr, mostly of Mw ≥ 7 and occasionally Mw ∼ 8 (e.g. Kawasaki et al. 2001). In addition to large and great earthquakes, the Tohoku area exhibits a full range of slow slip, including low frequency earthquakes (Fukao et al. 2016). These studies have provided great insight into the role of slow slip in reducing earthquake hazard on adjacent fault patches through stress transfer (Mazzotti et al. 2013) and SSEs, two of which occurred immediately preceding the Tohoku earthquake (Ito et al. 2013).

One year after the Tohoku earthquake, the Japan Trench Fast Drilling Project (JFAST) was conducted as Integrated Ocean Drilling Program (IODP) Expedition 343 (Chester et al. 2013a). JFAST drilling was conducted ∼7 km from the deformation front, within the region of highest coseismic slip during the Tohoku earthquake. Core sampling concentrated within ∼650–840 m below seafloor (mbsf) at IODP hole C0019E recovered a relatively structureless siliceous mudstone hanging wall, and a highly localized zone of scaly clay with intense deformation fabric at 821.5–822.5 mbsf. This scaly clay zone is interpreted to be the plate-boundary fault zone, although due to incomplete core recovery it may not necessarily contain the principal slip zone of the Tohoku earthquake (Chester et al. 2013b). The sheared clays are inferred to be derived from pelagic clays observed at the base of an older borehole seaward of the Japan Trench, at Site 436 (Kirkpatrick et al. 2015). The composition of the clays that make up the fault zone is especially rich in smectite, which comprises up to ∼80 per cent of the bulk sediment (Kameda et al. 2015).

2.2 Nankai Trough, Japan

Southeast of the Japan Trench, the Nankai Trough is formed by the subduction of the Philippine Sea Plate beneath the Amurian microplate at a rate of ∼4–6.5 cm yr⁻¹ (Miyazaki & Heki 2001; Loveless & Meade 2010) (Fig. 1b). The region is known for a history of destructive earthquakes that may extend over 1300 yr (Ando 1975). Two recent Mw > 8 earthquakes occurred in 1944 and 1946 (Tanioka & Satake 2001; Cummins et al. 2002; Ichinose et al. 2003; Kikuchi et al. 2003) making this region a main focus of several earthquake studies, including scientific drilling expeditions. Deep Sea Drilling Program (DSDP), Ocean Drilling Program (ODP), and
IODP expeditions have drilled three northwest-southeast trending borehole transects over 300 km along strike aligned with Cape Ashizuri, Cape Muroto, and through the Kumano Basin. We focus on samples recovered from two particular sites; the first is Site 1174, which penetrated the plate-boundary décollement near the Nankai Trough deformation front during ODP Leg 190 (Moore et al. 2001). This site is located updip of the 1944 Tonankai earthquake rupture area (Fig. 1b). The second is Site C0004, which penetrated a major out-of-sequence splay fault known as the “megasplay” during IODP Expedition 316 (Kinoshita et al. 2009). Splay faulting has been implicated as a potential host of coseismic slip in the Nankai Trough, based on seismic reflection surveys (Park et al. 2002; Moore et al. 2007), geodetic and tsunami data (Sagiya & Thatcher 1999; Cummins & Kaneda 2000; Baba et al. 2006), numerical models of rupture branching (Kame et al. 2003), and indications of frictional heating from vitrinite reflectance measurements on core samples (Sakaguchi et al. 2011). Shallow very low frequency earthquakes (VLFE) have been observed within the Nankai accretionary prism (Obara & Ito 2005; Ito & Obara 2006a, 2006b), likely on splay faults (Obara & Ito 2005; Ito & Obara 2006a, 2006b) or the plate boundary décollement (Sugioka et al. 2012). Recently, Araki et al. (2017) documented fluid pressure transients in borehole observatories indicating a shallow SSE propagating updip which may have reached the trench.
2.3 Cascadia

The Cascadia subduction zone is formed by the eastward subduction of the Juan de Fuca Plate beneath the North American Plate, and extends over 1000 km from offshore northern California to Vancouver Island (Fig. 1c). The potential for earthquakes of $M_w \geq 8$ in Cascadia has been the subject of debate; early geodetic studies and a lack of recorded large earthquakes over the past $\sim 180$ yr suggested aseismic subduction of the Juan de Fuca Plate (e.g. Ando & Balazs 1979). However, palaeoseismology studies based on shallow sediment stratigraphy provide evidence that large earthquakes have occurred over the past $\sim 8000$ yr with a recurrence interval of $\sim 500–600$ yr (Atwater 1987; Adams 1990; Kelsey et al. 2002; Goldfinger et al. 2003; Witter et al. 2003), with the latest large earthquake having occurred in $\sim 1700$ (Satake et al. 1996; Jacoby et al. 1997). More recent geodetic measurements indicate plate locking and slip deficit sufficient for future great earthquakes (e.g. Hyndman & Wang 1995; Flück et al. 1997; McCaffrey et al. 2000; Wang et al. 2003; Wang & Tréhu 2016).

The Cascadia subduction zone has become well-known for exhibiting diverse types of slow fault slip, particularly SSEs and tectonic tremor (Dragert et al. 2001; Miller et al. 2002; Rogers & Dragert 2003; Szeliga et al. 2008; Wech & Creager 2011). These observations are based on measurements from onland GPS and seismometer networks, and therefore locate these events to the plate interface deeper than 25 km. The Cascadia margin has been drilled during several expeditions; of particular importance to this study is Site 891 drilled during ODP Leg 146, which penetrated the frontal thrust in the prism toe at the southern Cascadia margin offshore central Oregon (Westbrook et al. 1994).

2.4 Costa Rica

At the Middle America Trench (MAT) offshore Costa Rica, the Cocos Plate subducts beneath the Caribbean Plate at a rate of $\sim 8.8$ cm yr$^{-1}$ (DeMets et al. 2010) (Fig. 1d). We focus here on the Nicoya Peninsula area, which was the location of ODP drilling during Leg 170 (Kimura et al. 1997) and Leg 205 (Morris et al. 2003). The character of the subducting plate in this region is diverse; offshore northern Nicoya Peninsula the crust originates from the East Pacific Rise, offshore southern Nicoya from the Cocos-Nazca spreading centre. A key difference is that the EPR crust is bathymetrically smooth compared to the CNS crust, on which several trench-perpendicular ridges are underthrust at the Lesser Antilles margin. However, the low seismic activity in the vicinity of these ridges has led to their classification as ‘aseismic’ (McCann & Sykes 1984; Bouysse & Westercamp 1990). The Lesser Antilles margin overall exhibits low seismicity, with no earthquakes of $M_w \geq 8$ since 1900 and only one subduction thrust earthquake of $M_w > 6$ since 1973 (Dorel 1981; Stein et al. 1982; Hayes et al. 2014). Although it was suggested that the low seismicity rate is related to the low plate convergence rate and/or aseismic creep, Hayes et al. (2014) suggest that enough elastic strain has accumulated to produce an earthquake of $M_w \geq 8$. Slow slip and tectonic tremor in the Lesser Antilles has not yet been reported from the locally installed GPS and seismic networks.

2.5 Barbados

The Barbados accretionary complex is located in the Lesser Antilles subduction zone, where the diffuse boundary between the North American and South American plates is subducting beneath the Caribbean Plate at a rate of $\sim 2$ cm yr$^{-1}$ (Minster & Jordan 1978; DeMets et al. 2000, 2010) (Fig. 1e). Seismic reflection surveys revealed a wide accretionary complex with several mud volcanoes, which suggest fluid overpressuring (Westbrook & Smith 1983; Westbrook et al. 1988). Exploring the hydrogeology of accretionary prisms provided the motivation for drilling a transect of several boreholes near the deformation front during ODP Legs 110 (Mascle et al. 1988) and 156 (Shipley et al. 1995).

The ODP boreholes are located on a flank of the Tiburon Rise, one of three trench-perpendicular ridges being underthrust at the Lesser Antilles margin. Low seismic activity in the vicinity of these ridges has lead to their classification as ‘aseismic’ (McCann & Sykes 1984; Bouysse & Westercamp 1990). The Lesser Antilles margin overall exhibits low seismicity, with no earthquakes of $M_w \geq 8$ since 1900 and only one subduction thrust earthquake of $M_w > 6$ since 1973 (Dorel 1981; Stein et al. 1982; Hayes et al. 2014). Although it was suggested that the low seismicity rate is related to the low plate convergence rate and/or aseismic creep, Hayes et al. (2014) suggest that enough elastic strain has accumulated to produce an earthquake of $M_w \geq 8$. Slow slip and tectonic tremor in the Lesser Antilles has not yet been reported from the locally installed GPS and seismic networks.

2.6 Hikurangi, New Zealand

The Hikurangi subduction zone is located off the east coast of New Zealand’s North Island (Fig. 1f). At this margin, the Pacific Plate subducts westward beneath the North Island at a rate of 2–6 cm yr$^{-1}$, increasing from south to north (Wallace et al. 2004). Although this margin has hosted megathrust earthquakes with moment magnitude $M_w$ of up to 7.1, none this large have been recorded over the past century (Doser & Webb 2003). However, geodetic locking estimations indicate that the plate interface has the potential to rupture in earthquakes of $M_w$ 8 or larger (Wallace 2009).

The Hikurangi subduction zone hosts frequent SSEs, which have been detected geodetically since 2002 by a continuous GPS network (e.g. Douglas et al. 2005; Wallace et al. 2009, 2012, 2016, 2017; Wallace & Beavan 2010). The SSEs are clearly spatially segmented; north of around 40–41°S the SSEs are shallower (<15 km depth), shorter in duration (1–3 weeks), and recur more frequently (every 1–2 yr), whereas to the south the SSEs are deeper (~35–60 km depth), longer (>1 yr), and less frequent (recurrence of ~5 yr). Furthermore, it has been found that the upper depth limit of SSE rupture areas coincides with the depth extent of geodetic plate locking along the margin (Wallace et al. 2009, 2012; Wallace & Beavan 2010). Noteworthy facets of the Hikurangi SSEs include mutual triggering relationships with ordinary earthquakes (Koulali et al. 2017; Wallace et al. 2017, 2018), and the observation of accompanying microseismicity rather than tremor (Delahaye et al. 2009).

Corning of the subduction inputs, the upper plate, and a frontal thrust was recently completed during IODP Expedition 375 (Saffer et al. 2018). In particular, nearly 1200 m of input material was...
drilled which revealed the major lithologies to be a hemipelagic mud/tubidite trench wedge facies, pelagic chalk, and coarse volcanioclastics. A similar chalk unit was recovered during previous scientific drilling of the incoming sedimentary sequence ~400 km east of the Hikurangi Trough during ODP Leg 181, at Site 1124 (Carter et al. 1999).

2.7 Alpine fault, New Zealand

Southwest from the Hikurangi Trough, the Pacific–Australian Plate boundary continues onland to become the Alpine Fault (Fig. 1f). The fault is right-lateral with a displacement rate of 23–27 mm yr$^{-1}$ over the past ~50 000 yr, based on radiocarbon dating and offset of geological markers (Cooper & Norris 1994; Norris & Cooper 2000; Sutherland et al. 2006). Palaeoearthquakes studies estimate the recurrence of major earthquakes ($M_w > 7.0$) to be 260–400 yr over the past 8000 yr (Sutherland & Norris 1995; Bull 1996; Berrymen et al. 2012), with none occurring in the past ~300 yr (Cooper & Norris 1990; Sutherland et al. 2007). Geodetic measurements indicate that the central segment of the Alpine Fault is fully locked down to depths of 5–8 km and partially locked to ~18 km, with no creep occurring at the surface (Beavan et al. 1999, 2007; Norris & Cooper 2000; Wallace et al. 2007). Furthermore, based on earthquake hypocentres the Alpine Fault seismogenic zone is estimated to range from ~12 km up to near the Earth’s surface (Leitner et al. 2001). Thus, previous studies collectively indicate that large magnitude earthquake, likely surface-breaking, should be expected in the near future.

Evidence that the Alpine Fault is nearing the end of its earthquake cycle has motivated the Deep Fault Drilling Project (DFDP) within the framework of the International Continental Scientific Drilling Project (ICDP). During the initial phase of the project, two shallow boreholes located 80 m apart were drilled through the Alpine Fault: DFDP-1A drilled to 96 m depth, and DFDP-1B drilled to 151 m depth. Both boreholes successfully penetrated principal slip zones (PSZs) consisting of clayey gouge, representing the fault zone proper (Sutherland et al. 2012; Townend et al. 2013). In DFDP-1B, two PSZs were identified (Toy et al. 2015). The main (upper) PSZ, which was tested in this study, is ~20 cm thick and located within green cataclasites at a depth of ~128 m. The second PSZ is located at ~144 m depth; it is not clear which PSZ was most recently active.

2.8 San Andreas fault, California

The San Andreas Fault runs along western California, and is known for its central creeping section between San Juan Bautista in the north and Cholame in the south (Fig. 1g). The average displacement rate on the San Andreas Fault is ~25–32 mm yr$^{-1}$ from GPS measurements (Titus et al. 2005, 2006; Ryder & Bürgmann 2008). The fault is locked and accumulating elastic strain outside of the creeping section (Irwin & Barnes 1975), and even within the creeping section some studies suggest slip deficit accumulation that could be released as a $M_w$ ~6–7 earthquake (Ryder & Bürgmann 2008; Johnson 2013; Maurer & Johnson 2014; Jolivet et al. 2015). The San Andreas Fault is one of the first fault zones on which slow slip was identified, from strainmeter measurements near the southern (Goulty & Gilman 1978) and northern (Nason 1969; Gladwin et al. 1994; Linde et al. 1996) ends of the creeping zone. These are shallow events occurring at <500 m depth.

The San Andreas Fault Observatory at Depth (SAFOD) penetrates the San Andreas Fault at ~2.7 km depth with a northeast-dipping borehole near Parkfield, California (Zoback et al. 2011). The drill site is located at the southern boundary of the SAF central creeping section, approximately 25 km northwest of the fault patch that hosted 7 repeating earthquakes near Parkfield dating back to 1857, the most recent being the $M_w$ 6.0 event in 2004 (Bakun & McEvilly 1984). In the direct vicinity, several clusters of repeating $M_w = 2.0$ earthquakes have been observed (Nadeau & Johnson 1998; Zoback et al. 2011).

Borehole geophysical logs, cuttings, and core obtained during SAFOD identified two actively creeping fault strands, termed the Central Deforming Zone (CDZ) and the Southwest Deforming Zone (SDZ). The CDZ is considered the main strand of the active SAF accommodating plate motion, whereas the SDZ marks the southwest edge of the modern SAF damage zone (Zoback et al. 2011; Moore 2014; Carpenter et al. 2015a). A key finding from the SAFOD core is the large amount of magnesium-smectite (saponite) within the CDZ (e.g. Solum et al. 2006; Schleicher et al. 2012), implicated in controlling the slip behaviour of the creeping SAF (Carpenter et al. 2011, 2012, 2015a; Holdsworth et al. 2011; Lockner et al. 2011; Hadizadeh et al. 2012; Ikari et al. 2016a).

2.9 Woodlark Basin, Papua New Guinea

The Woodlark Basin is formed by north-south extension between Papua New Guinea to the west and the Solomon Islands to the east (Fig. 1h). The basin lies in a tectonically complex region between the Pacific and Australian plates, where the extension defines the boundary between the northern edge of the Australian Plate and the Woodlark microplate. Extension in the Woodlark basin is transitional from continental rifting at a rate of 3.6 cm yr$^{-1}$ at the western end, to seafloor spreading at a rate of 6.7 cm yr$^{-1}$ at the eastern end (Taylor et al. 1995; Tregoning et al. 1998; Wallace et al. 2014). In order to investigate the processes associated with rifting and continental breakup, ODP Leg 180 drilled several boreholes near the Moresby seamount, on the western (continental rifting) side of the Woodlark basin (Taylor et al. 2000; Taylor & Huchon 2002). At one site in particular (Site 1117), fault gouge exposed at the seafloor by dip-slip motion was successfully recovered. In this region, earthquake hypocentres interpreted to be faults (Japan Trench, Nankai Trough,
Table 1. Experiment and sample details.

| Location                  | Sample Type          | Expedition          | Sample ID | Depth of Recovery (m) | Plate Tectonic Rate (cm yr\(^{-1}\)) | Experimental Driving Rate (cm yr\(^{-1}\)) | Sample Composition (wt %) (abundances ≥ 5 percent) | References                                      |
|---------------------------|----------------------|---------------------|-----------|-----------------------|----------------------------------------|---------------------------------------------|-------------------------------------------------|------------------------------------------------|
| Japan Trench Subduction Decollement | IODP Exp. 343 Japan Trench Fast Drilling Project (JFAST) | C0019E 17R-1 | 823 | 8.3 | 8.4 | Smectite 81, Illite 8, Quartz 7 | Chester et al. (2013a); DeMets et al. (1994) XRD: Kameda et al. (2015) |
| Nankai Trough Subduction Decollement | ODP Leg 190    | 1174B 73R-1 | 833 | 6.3-6.8 | 6.3 | Quartz 39, Smectite 20, Illite 12, Plagioclase 10, Calcite 10 | Miyazaki & Heki (2001); Moore et al. (2001) XRD: Steurer & Underwood (2005) |
| Nankai Trough Splay Fault | IODP Expedition 316 Nankai Trough Seismogenic Zone Experiment (NaTroSEIZE) | C0004D 29R-3 | 277 | 4.0-6.8 | 6.3 | Quartz 21, Plagioclase 18, Smectite 15, illite 13, Chlorite 6 | Miyazaki & Heki (2001); Kinoshita et al. (2009) |
| Cascadia Subduction Frontal Thrust | ODP Leg 146  | 891B 43X | 333 | 4.2 | 5.3 | Plagioclase 25, Quartz 22, Illite 12, Mixed-layer clay 8, K-feldspar 8, Muscovite 5, Chlorite 5 | Westbrook et al. (1994) |
| Costa Rica Subduction Decollement | ODP Leg 170  | 1040C 21R-6 | 359 | 8.2-8.8 | 8.7 | Smectite 46, Illite 12, Plagioclase 9, Mixed-layer clays 8, Zeolite 8 | Kimura et al. (1997) |
| Costa Rica Input Sediment | ODP Leg 205  | 1253A 3R-2 | 387 | 8.2-8.8 | 8.7 | Calcite 66, Amorphous Silica 20 | Morris et al. (2003) |
| Barbados Subduction Decollement | ODP Leg 156  | 948C 11X-4 | 513 | 2.0-3.7 | 5.3 | 29 Illite, Kaolinite 12, Smectite 11, Mixed-layer clays 9, Quartz 7, Plagioclase 7, Glauconite 7 | DeMets et al. (1994); Shipley et al. (1995) |
| Hikurangi Input Sediment | ODP Leg 181  | 1124C 20X-5, 21X-5, 22X-5 (1b59,1.19.DS) | 195-205 | 2.0-6.0 | 5.3 | Calcite 43, Plagioclase 13, Quartz 9, Mixed-layer clays 8 | Carter et al. (1999); Wallace et al. (2004) |
| Alpine Fault Principal Slip Zone | Deep Fault Drilling Project (DFDP)-1b | 127 | 2.6 | 5.3 | Quartz 20, Plagioclase 18, Illite 10, Smectite 8, Calcite 8, Chlorite 7, K-feldspar 6, Muscovite 5, Clinopyroxene 5 | Toy et al. (2015) |
| San Andreas Fault Central Deforming Zone | San Andreas Fault Observatory at Depth (SAFOD) | G43 | ~2700 | 2.5 | 5.3 | Smectite 39, Chlorite 17, Illite 10, Chrysotile 6, Mixed-layer clay 5 | Titus et al. (2005); Zoback et al. (2011) |
| Woodlark Basin Detachment Normal Fault | ODP Leg 180  | 1117A 1R-2 | 9 | 1.0 (3.7) | 5.3 | Smectite 12, Pyrite 12, Kaolinite 11, Talc 10, Serpentine 10, Amphibole 8, Chlorite 7, Orthopyroxene 5 | Taylor et al. (2000) |
Cascadia, Costa Rica décollement and Barbados, and two other samples (Costa Rica pelagic chalk, Hikurangi) were recovered from the input sediments on the incoming plate. Incoming sediments at subduction zones are a valuable resource for investigating the mechanical behaviour of the plate boundary because they constitute the eventual shallow megathrust fault (Underwood 2007; Hüpers et al. 2017; Ikari et al. 2018).

The composition of the samples tested here ranges widely, as measured by X-ray diffraction (XRD) (Table 1). XRD for most samples was performed with a Philips X'Pert Pro multipurpose diffractometer following the double-identification method of Vogt et al. (2002), which employs a preliminary mineral identification with the software MacDiff followed by a full-pattern identification with the QUAX mineral database. XRD data for the Japan Trench fault zone samples was reported in Kameda et al. (2015); for the Nankai Trough décollement sample in Moore et al. (2001) and Steurer & Underwood (2005); and for the Nankai Trough megasplay sample in Kinoshita et al. (2009).

### 3.2 Experimental procedure

For these experiments, the samples were tested as powdered gouge with a maximum grain size of 125 μm. The powders were mixed with 3.5 percent NaCl brine to form a stiff paste, which was then cold-pressed into a sample cell that holds a cylindrical volume of ~25 mm height and 25 mm diameter. For the shear experiments, the cell was loaded into a single-direct shear device within which the normal stress is applied to the top face of the sample, aligned parallel to the cylinder axis (Fig. 2). The samples were allowed to consolidate and drain for at least 18 hr (in some cases several days), until the sample height (~15–20 mm) reaches a steady value. This is done because the apparatus in not equipped with a direct pore pressure measurement system, therefore by allowing full consolidation we may assume that the sample is drained before shearing and that the applied normal stress is equal to the effective normal stress (σ_u). All experiments were conducted at room temperature and 10 MPa effective normal stress, which were chosen for two main reasons: (1) to approximate the in-situ conditions of the samples, most of which were recovered from depths of <1 km (Table 1) and (2) due to force limitations of the apparatus. The apparatus stiffness was determined by loading and unloading a steel blank having the same dimensions as the samples, at a slow rate that matches the cm yr^{-1} shearing velocities in this study. The unloading stiffness under 10 MPa normal load, which provides a direct comparison to stress drops, is 7–11 MPa mm^{-1} depending on the magnitude of the shear stress.

Following consolidation, the apparatus displaces the bottom half of the sample cell relative to the top half inducing planar shear perpendicular to the cylinder axis. As the bottom half of the sample is displaced away, the top half of the sample shears against the lower sample cell. Therefore, although the sample-sample contact reduces with displacement, the full area of the top sample half continuously supports load and experiences deformation throughout the experiment. No systematic changes in friction are observed as a function of displacement, indicating that effects of shearing against the lower sample cell are negligible (see Ikari et al 2015a for further details).

During shear, the shear strength (τ) is continuously recorded, from which an apparent coefficient of sliding friction μ is calculated as the ratio τ/σ_u (Fig. 2). All samples were initially sheared with a constant driving velocity (load point velocity V_p) of 10 μm s^{-1} for up to ~4–5 mm, by which a steady-state shear strength is reached. The slip velocity was then decreased to 5.3–8.7 cm yr^{-1} (1.7–2.8 mm s^{-1}) for ~2 mm to simulate naturally slow plate tectonic driving rates (Table 1). At these plate rates, laboratory SSE tend to occur from which important parameters that may be directly compared with natural SSEs are measured: stress drop, which is directly measured from the shear stress record; peak slip velocity during SSE stress drop and initial slip velocity prior to the stress drop; and unloading stiffness, calculated as the maximum slope of the shear stress-sample displacement curve during the stress drop. SSE duration and event slip are also measured, however due to the small size of the samples these values are not directly comparable to natural SSEs without an appropriate scaling factor (e.g. Rabinowitz et al. 2018).

For most of these experiments, the shear device utilized two horizontal displacement sensors (Fig. 2); one is mounted at the horizontal load cell which monitors the driving velocity, that is the load point velocity V_p. The second is mounted directly at the shear cell, which measures the offset of the cell plates and therefore the true sample displacement rate V. This dual-measurement system allows us to determine transient changes in slip rate which would occur during frictional slip events. Furthermore, by taking the difference between the sample and driving displacements it can be assessed whether a slip deficit or slip excess is occurring. In some earlier experiments where the load point sensor was not yet installed, slip deficit or excess can still be evaluated by removing the driving rate trend from the sample displacement record. Sample slip velocity measurements were calculated from a running average slope of at least 11 data points from the displacement time series. Due to the importance in sample slip velocity for evaluating SSEs and comparing them to natural geodetic observations, as well as the tendency for velocity perturbations to be small, estimations of error in the slip velocity measurements were made by measuring the standard deviation of the noise level in the displacement sensors when they are stationary. The standard deviation in both sample slip displacement and load point displacement ranges between 0.10 and 0.65 μm; which translates to a maximum uncertainty in slip velocity of ±2.2 cm yr^{-1} or 37 per cent in the most conservative case. Peak slip velocities for all SSEs in this study are well above this threshold.

Following shearing at the plate rate, the slip velocity was then increased by a factor of 3 (but sometimes up to a factor of ~11) to measure the rate- and state-dependence of friction (RSF) using established inverse modelling techniques (e.g. Reinen & Weeks 1993; Saffer & Marone 2003; Ikari et al. 2009, Fig. 2b, Table 3). The frictional response to a velocity step is described as:

\[
\mu = \mu_o + a \ln \left( \frac{V}{V_o} \right) + b_1 \ln \left( \frac{V_o \theta_i}{d_{c1}} \right) + b_2 \ln \left( \frac{V_o \theta_i}{d_{c2}} \right),
\]

(1)

\[
\frac{d\theta_i}{dt} = 1 - \frac{V_o \theta_i}{d_{c1}}, \quad i = 1, 2,
\]

(2)

where the parameters a, b_1 and b_2 are dimensionless constants, θ_1 and θ_2 are state variables having units of time, and d_{c1} and d_{c2} are critical slip distances over which frictional state evolves to a new steady value (e.g. Dieterich 1979, 1981; Marone 1998a; Scholz 2002). In most cases the data is adequately described using one state variable, eliminating the last term in eq. (1). In the cases where the data is better described using the two state variables, we report the parameter b as b = b_1 + b_2. A linear slip-dependent trend in friction is commonly superimposed on the data, which is removed during the modelling process (e.g. Blanpied et al. 1998).
Eq. (2) describes the time-dependence of friction via the evolution of the state variable $\theta$, known as the ‘Dieterich’, or ‘aging’ law. Other forms of eq. (2) exist, the most prominent being a version in which the state variable evolution depends on slip (i.e. ‘slip law’, Ruina 1983). The choice between the aging and slip laws is a topic of debate; the appropriateness of each law highly depends on its application (e.g. Sleep 2005, 2006, 2012), with the slip law being supported by some recent studies (e.g. Bhattacharya et al. 2017).

For this study, I employ the Dieterich aging law because it captures the property that friction can change as a function of time and not only slip, which previous studies have verified by analysing ‘slide-hold-slide’ frictional healing experiments (e.g. Beeler et al. 1994; Ikari et al. 2016b) and by observing time-dependent increase in real area of contact at grain asperities during static loading (e.g. Dieterich & Kilgore 1994, 1996b; Goldsby et al. 2004). The aging law has been proven useful for simulating various types of slow fault slip (e.g. Perfettini & Ampuero 2008; Helmbestetter & Shaw 2009), with some studies finding that the aging law better describes natural phenomena (Rubin 2008; Kanu & Johnson 2011). I leave open the possibility that the slip law can provide an alternative explanation to the results presented here, but a detailed investigation comparing the two evolution laws is not a specific goal of this work.

Regardless of which evolution law is used, at steady state $V = d_s/\theta$, so that $d\theta/dt = 0$, and eqs (1) and (2) simplify to:

$$a - b = \frac{\Delta \mu_s}{\Delta \ln V},$$

(3)

where $\mu_s$ is a steady-state value of friction. In addition to $a-b$ values determined by inverse modelling of friction changes due to velocity increases or ‘upsteps’, a second set of $a-b$ values was determined by measuring the steady-state change in friction following the decrease in slip velocity from the run-in value of $10 \mu m s^{-1}$ to the cm yr$^{-1}$ slip velocities, or ‘downsteps’ (Ikari & Kopf 2017). These were calculated using eq. (3) because modelling of these downsteps proved difficult, likely due to the large velocity difference ($\sim 3.5$ orders of magnitude) and low apparatus stiffness (e.g. Bhattacharya et al. 2015).

The parameter $a-b$ is of great importance for fault slip behaviour because it determines the possibility of unstable frictional slip, which results in stick-slip behaviour in the laboratory and earthquakes in nature (e.g. Dieterich 1986; Tullis 1988; Dieterich & Kilgore 1996a; Marone 1998a, Scholz 2002). A positive value of $a-b$ is known as velocity-strengthening behaviour, a condition that prohibits the nucleation of unstable slip by producing a negative stress drop that damps a potential event. A negative value of $a-b$ indicates velocity-weakening frictional behaviour, which is a requirement for frictional instability. Velocity-weakening does not necessarily guarantee unstable slip, which also requires specific elastic conditions related to the fault surroundings, which is the apparatus in the laboratory and wall rock in nature (Cook 1981; Scholz 1998). Within this framework, various types of slow and transient fault slip events are thought to arise from frictional behaviour near the stability transition or ‘conditional stability’, where ‘self-sustained oscillatory slip’ may occur (e.g. Dieterich 1986; Dieterich & Kilgore 1996a; Scholz 1998).
4 FRICTIONAL STRENGTH AND VELOCITY DEPENDENCE

The steady-state coefficient of sliding friction measured during the 10 $\mu$m s$^{-1}$ portion of these experiments establishes a baseline friction level that can be compared with previous work under similar conditions. These friction values range from 0.12 (San Andreas Fault) to 0.60 (Costa Rica inputs chalk) and exhibit a clear inverse dependence on phyllosilicate mineral content determined by XRD measurements (Fig. 3a, Table 1), a pattern well-established by several previous laboratory friction studies, especially under water-saturated conditions (Byerlee 1978; Lupini et al. 1981; Shimamoto & Logan 1981; Logan & Rauenzahn 1987; Morrow et al. 1992, 2000; Brown et al. 2003; Saffer & Marone 2003; Niemeijer & Spiers 2005; Ikari et al. 2007, 2011a,b; Crawford et al. 2008; Carpenter et al. 2009; Smith & Faulkner 2010; Tembe et al. 2010; Giorgetti et al. 2015). Some notable observations which match previous results under similar testing conditions include: the extremely low friction of samples from the Tohoku region of the Japan Trench (Remitti et al. 2015; Ikari et al. 2015a,b), the San Andreas Fault CDZ (Carpenter et al. 2011, 2012, 2015a; Lockner et al. 2011; Coble et al. 2014; Ikari et al. 2016a), and the Barbados décollement (Kopf & Brown 2003); the moderately high strength of the Alpine Fault principal slip zone (Boulton et al. 2014; Ikari et al. 2014, 2015c; Niemeijer et al. 2016); the large strength contrast between the weak clay-rich and stronger carbonate-rich samples from Costa Rica (Ikari et al. 2013; Kurzawski et al. 2016); and the larger strength of the Nankai Trough megasplay at Site C0004 compared to the décollement at Site 1174 (Ikari & Saffer 2011).

The steady-state sliding friction coefficients measured at plate-rate velocities of 0.0017–0.0028 $\mu$m s$^{-1}$ (5.3–8.7 cm yr$^{-1}$) exhibit the same dependence on phyllosilicate content as the values obtained at 10 $\mu$m s$^{-1}$, but the absolute values show significant differences (Fig. 3a). This difference allows the calculation of $a-b$ values from the downstep in velocity from 10 $\mu$m s$^{-1}$ to the plate rates using eq. (3), in addition to the $a-b$ values determined by inverse modelling of the 3- to 11-fold velocity upstep from the plate rates. There is significant scatter, however, it can be seen that for both sets of $a-b$ values, velocity-weakening occurs for gouges of all strength but that velocity-strengthening is only observed for materials with at least $\mu a-b$ to $-0.003$ for the modelled velocity upsteps. Since the magnitude of the velocity change is accounted for in the calculation of $a-b$ (eq. 3), the large negative $a-b$ values suggests some fundamental change in frictional behaviour caused by either the large velocity changes or by shearing at extremely slow rates. Upsteps with the same magnitude velocity change as the downsteps (3.6–3.8 orders of magnitude) would be needed to determine if the asymmetry in $a-b$ values depends on the sign of the large velocity change.

5 SITE-SPECIFIC OBSERVATIONS OF LABORATORY SSE

5.1 Japan Trench

For this study, the same sample tested previously at 7 MPa effective normal stress by Ikari et al. (2015a) was tested at 10 MPa (Fig. 4). This experiment produced six SSEs having stress drops ranging from 60 to 130 kPa, peak slip velocities of 14–29 cm yr$^{-1}$ (approximately two to four times the driving rate), and unloading stiffnesses of 3–9 MPa mm$^{-1}$ (Table 2). The peak slip velocity and a slip excess (or slip deficit recovery) clearly occur during the stress drop. It can also be seen that there is a background friction level, from which loading occurs before the stress drop and returns to after the stress drop. Before the first two SSEs, there is a non-negligible amount of displacement that occurs between the loading and the stress drop, however for the last four events this is not the case. Both the background friction level at the post-stress drop strength and the slip between loadup and stress drop are features that are also observed in several other samples in this study.

Ito et al. (2013) reported two SSEs prior to the 2011 Tohoku earthquake, one in 2008 and one in 2011 that was ongoing at the time of the Tohoku event. The data presented here clearly confirm that the Tohoku plate boundary material has the ability to generate SSEs. However, the conditions of these experiments represent much shallower depth than the location of the Tohoku SSEs, which occurred at a depth of $\sim 10–15$ km. This difference in depth and therefore stress conditions could explain why the peak slip velocities measured in these experiments only reach 5–10 percent of the average slip velocity of the 2008 and 2011 Tohoku SSEs, which is estimated to be $\sim 0.1$ $\mu$m s$^{-1}$.

5.2 Nankai trough

The sample from the Nankai Trough megasplay fault zone exhibits several SSEs, which are smaller and occur more frequently compared to the Japan Trench sample (Fig. 5). 9 events occur over $\sim 1.5$ mm displacement with stress drops ranging from 15 to 61 kPa, peak slip velocities of 12–22 cm yr$^{-1}$ (compared to a 6.3 cm yr$^{-1}$ driving rate), and unloading stiffnesses of $\sim 3-6$ MPa mm$^{-1}$. In some cases, stress drops are observed that appear to be SSEs, however a peak in sample slip velocity could not be distinguished from the noise level. This can be seen in Figs 5(b)–(e), where two such perturbations are shown in detail. The first stress drop is clearly associated with an increase in slip rate and an episode of slip excess relative to the background driving, the second is not. Therefore, only the first of these two stress drops is considered to be an SSE. In contrast to the megasplay sample, the sample from the Nankai Trough décollement at Site 1174 exhibits no SSEs (Fig. 6). Although some very small stress drops can be seen which are accompanied by some small velocity increases, the velocity perturbations are in phase with the stress perturbations. The peak slip velocity therefore does not coincide with the stress drop, so these perturbations are not considered to be SSEs.

The observation that the megasplay fault sample tested here produces SSEs suggests that megasplay might be expected to host SSE slip, whereas the lack of SSEs observed in the Nankai Trough décollement sample suggests that slow slip may not be able to reach the trench. It should be pointed out, however, that the décollement sample tested here is from a different transect, lying $\sim 200$ km southwest of the observatories which detected the recent SSE (Araki et al. 2017).

5.3 Cascadia

The sample from the southern Cascadia frontal thrust zone offshore Oregon exhibits 4 SSEs with stress drops of $\sim 50–80$ kPa, peak slip velocities of 14–26 cm yr$^{-1}$, and unloading stiffnesses of $\sim 4–7$ MPa mm$^{-1}$. Other than the first event, a notable characteristic of the Cascadia laboratory SSEs is the large amount of steady-state slip (up to 0.5 mm) that occurs between the loading phase and the stress...
drop (Fig. 7). The differential displacement record shows no slip deficit accumulation during the pre-event slipping phase, consistent with steady shearing before the stress drop.

Little is known about the possibility of slow slip at shallow depths in Cascadia, under conditions similar to those explored in these experiments. A recent study using borehole seismic and geodetic data offshore Vancouver Island show that teleseismic waves do not trigger shallow SSEs or tremor, whereas in other subduction zones such activity does result in triggered slow activity (McGuire et al. 2018). On the other hand, Wech & Creager (2011) show some sparse tremor activity seaward, including some events near the Site 891 borehole offshore Oregon. Adopting the assumption of Wech & Creager (2011) that tremor accompanies SSEs in this region, the SSEs observed in the Site 891 sample is consistent with shallow slow slip in the southern Cascadia margin. The lack of triggered slow slip to the north may suggest along-strike variations in mechanical properties at Cascadia, of which there is previously documented evidence (e.g. Brudzinski & Allen 2007; Han et al. 2017).

5.4 Costa Rica

The clay-rich décollement sample from the Costa Rica subduction zone exhibits two SSEs (along with a third stress perturbation for which an increase in slip velocity could not be resolved, and a fourth that could not be measured due to a machine error, Fig. 8). The first SSE has a much larger stress drop (254 kPa) than the second SSE (52 kPa), but the peak slip velocities are similar (18.4 mm\(^{-1}\) for the first SSE, 15.7 mm\(^{-1}\) for the second SSE). The detrended displacement record clearly shows a slip deficit accumulation during the loading phase, and slip excess during the stress drop. The unloading stiffness is 5.6 MPa mm\(^{-1}\) for the first SSE and 2.0 MPa mm\(^{-1}\) for the second SSE; an interesting observation is that the first SSE has an unloading stiffness that is larger than the loading stiffness, however for the second SSE this asymmetry is reversed (Fig. 8c).

The chalk sample recovered from the input sediment seaward of the trench at Costa Rica shows strikingly different behaviour compared to the clay-rich décollement sample. At the plate tectonic driving rate of 8.7 cm yr\(^{-1}\), several closely spaced, repetitive slip events are observed, which strongly resemble ordinary stick-slip events (e.g. Brace & Byerlee 1966; Byerlee & Brace 1968; Savage & Marone 2007, Fig. 9). Unlike the SSEs in the Costa Rica décollement and other samples in this study, the peaks and troughs in frictional strength are relatively uniform and there is no tendency toward steady-state slip at any point. The stress drops are \(\sim 2.5 \text{ MPa}\) which represents 40 per cent of the shear strength, the largest value in this study (Table 2). The stress drops clearly correlate with increments of slip. The inter-event slip rate appears to be negative, with indications of a precursory acceleration phase prior to the events. More detailed analysis of an individual event shows that the peak slip velocity reaches 10 \(\mu\text{ms}^{-1}\), orders of magnitude faster than any other slip event in this study.

The shallow subduction zone offshore Costa Rica exhibits both SSEs and VLFEs (e.g. Outerbridge et al. 2010; Walter et al. 2013; Dixon et al. 2014). Due to the erosional nature of this subduction zone, the heterogeneously distributed clay and chalk on the incoming plate are expected to heavily influence the shallow megathrust behaviour. The data from these two lithologies is consistent with the activity at Costa Rica: the clay-rich hemipelagic material produces SSEs in these experiments, whereas the chalk produces much faster stick-slip-like events. The clayey sediments are more widespread on the incoming plate; therefore the observations from these experiments are consistent with the hemipelagic clay producing the broad SSEs, and smaller chalk asperities producing the VLFEs observed at Costa Rica. The average slip velocity during the 2007 Nicoya Peninsula SSE (Outerbridge et al. 2010) is estimated to be \(\sim 24 \text{ cm yr}^{-1}\) (~2 cm slip over 30 d, Jiang et al. 2012; Walter et al. 2013), a value similar to the peak slip velocities of 16-18 cm yr\(^{-1}\) measured in the sample from the clay-rich décollement. Slip velocity was not estimated for the Costa Rica VLFEs, but the 10 \(\mu\text{ms}^{-1}\) measured for the slip events in the chalk sample approach the lower end of the range 0.05–2 mm s\(^{-1}\) estimated for VLFEs in the Nankai Trough (Ito & Obara 2006b), albeit for a different material.

5.5 Barbados

The sample from the Barbados subduction décollement produced three SSEs (excluding two superimposed on velocity step tests, which were not evaluated). The SSEs occurred at a driving rate...
Figure 4. Experimental data for a sample from the Tohoku region of the Japan Trench. (a) Coefficient of friction as a function of displacement. (b) Closeup of SSEs generated during plate-rate shearing. (c) Closeup of shear stress during an individual SSE, showing the measurement of the stress drop $\Delta \tau$ and unloading stiffness $K_s$. (d) Sample slip velocity $V$ and driving (load point) velocity $V_{lp}$. (e) Difference between sample displacement and load point displacement and (f) unloading stiffness (measured as $-d\tau/dx$, where $x$ is the sample slip displacement) for the SSE shown in (c). Shading indicates the SSE stress drop. Note that this figure is a full example of the laboratory SSE measurements, for brevity not all measurements may be shown in subsequent Figs 5–14.

of 5.3 cm yr$^{-1}$, two of which had exceptionally large stress drops (725 and 880 kPa), with peak slip rates of 35.1 and 42.2 cm yr$^{-1}$ and unloading stiffnesses of 6.9 and 8.5 MPa mm$^{-1}$, respectively (Fig. 10). The stress drops of these two SSEs represent $\sim$30 per cent of the shear strength, among the higher values in this study (Table 2). A notable feature of these two SSEs is their broad shape in the shear stress-displacement record, exhibiting 180–190 $\mu$m of event slip (i.e. slip during the stress drop), which is larger than that of any sample except the San Andreas Fault. A smaller SSE occurred between the two large events which exhibited a stress drop of 93 kPa, and lower peak slip velocity (12.0 cm yr$^{-1}$) and unloading stiffness (3.8 MPa mm$^{-1}$) compared to the two larger stress drop events.

The lack of reported slow slip activity near Barbados hampers comparison between laboratory and natural observations. However, the observation that SSEs occur in the Barbados décollement sample does suggest that this material has the ability to store some amount of elastic strain energy, which supports the assertion of Hayes et al. (2014) that enough geodetically-measured strain deficit has accumulated with the potential for a $M_w \sim 8$ earthquake.

5.6 Hikurangi, New Zealand

Three experiments were conducted on the Hikurangi sediment input sample at 10 MPa effective normal stress; these experiments were previously included in studies by Rabinowitz et al. (2018) and Ikari et al. (2019). Two of these experiments exhibited a total of five SSEs, with one experiment exhibiting no SSEs. The Hikurangi SSE stress drops are generally small and range from 23 to 46 kPa, peak slip velocities range from 11.6 to 19.3 cm yr$^{-1}$, and unloading stiffnesses are 2.5–13.7 MPa mm$^{-1}$ (Fig. 11). Little slip deficit is observed during the loading phase, whereas slip excess can clearly be seen during the stress drop. The form of the stress drop is variable,
Table 2. Parameters measured from laboratory SSE.

| Experiment | Location                  | Stress drop (kPa) | Percentage stress drop (%) | Slip (μm) | Duration (s) | Pre-SSE velocity (nms⁻¹) | Pre-SSE velocity (cm yr⁻¹) | Peak velocity (nms⁻¹) | Peak velocity (cm yr⁻¹) | $V_p/V_s$ | $K_s$ (MPa mm⁻¹) | $K$ (MPa mm⁻¹) | $K/K_s$ |
|------------|----------------------------|-------------------|-----------------------------|-----------|--------------|---------------------------|---------------------------|---------------------|------------------------|-----------|----------------|----------------|---------|
| B651       | Japan Trench              | 121               | 4.6                         | 39.0      | 11161        | 2.6                       | 8.2                       | 7.1                 | 22.4                   | 2.7       | 8.7            | 8.7            | 1.0     |
|            |                            | 127               | 4.8                         | 36.9      | 8688         | 2.6                       | 8.0                       | 8.5                 | 26.8                   | 3.3       | 5.2            | 8.7            | 1.7     |
|            |                            | 79                | 3.1                         | 34.3      | 7353         | 1.9                       | 5.9                       | 7.7                 | 24.3                   | 4.2       | 2.9            | 8.9            | 3.1     |
|            |                            | 118               | 4.6                         | 40.6      | 8499         | 2.6                       | 8.2                       | 8.8                 | 27.7                   | 3.4       | 3.7            | 8.9            | 2.4     |
|            |                            | 63                | 2.5                         | 58.3      | 19892        | 2.6                       | 8.2                       | 4.5                 | 14.2                   | 1.7       | 3.0            | 8.8            | 2.9     |
|            |                            | 131               | 5.1                         | 31.6      | 6818         | 2.2                       | 7.1                       | 9.1                 | 28.6                   | 4.0       | 4.7            | 8.9            | 1.9     |
| B518       | Nankai Trough Megasplay   | 33                | 0.8                         | 8.2       | 3820         | 1.1                       | 3.4                       | 3.7                 | 11.7                   | 3.4       | 5.1            | 10.7           | 2.1     |
|            |                            | 36                | 0.9                         | 11.0      | 4520         | 1.8                       | 5.7                       | 4.7                 | 14.9                   | 2.6       | 4.3            | 10.8           | 2.5     |
|            |                            | 15                | 0.4                         | 5.8       | 2350         | 2.0                       | 6.4                       | 3.7                 | 11.6                   | 1.8       | 3.5            | 10.7           | 3.0     |
|            |                            | 26                | 0.6                         | 19.9      | 8520         | 1.7                       | 5.2                       | 3.7                 | 11.6                   | 2.2       | 2.9            | 10.5           | 3.7     |
|            |                            | 33                | 0.8                         | 13.4      | 4120         | 1.4                       | 4.4                       | 5.6                 | 17.5                   | 4.0       | 3.0            | 10.7           | 3.6     |
|            |                            | 51                | 1.3                         | 18.9      | 8580         | 1.1                       | 3.5                       | 4.7                 | 14.7                   | 4.2       | 4.1            | 10.7           | 2.6     |
|            |                            | 26                | 0.7                         | 11.4      | 5240         | 2.5                       | 7.8                       | 5.0                 | 15.9                   | 2.0       | 3.8            | 10.5           | 2.8     |
|            |                            | 42                | 1.1                         | 12.3      | 4060         | 1.4                       | 4.4                       | 4.3                 | 13.7                   | 3.1       | 4.8            | 10.5           | 2.2     |
|            |                            | 61                | 1.5                         | 21.2      | 6900         | 2.0                       | 6.4                       | 7.1                 | 22.4                   | 3.5       | 5.6            | 10.5           | 1.9     |
| B541       | Nankai Trough Decollement | 0                | 0                           | 0         | 0            | 2.0                       | 2.0                       | 6.3                 | 6.3                    | 1         | 0              | -              | -       |
| B565       | Cascadia                  | 81                | 1.7                         | 13.7      | 6210         | 0.4                       | 1.3                       | 5.5                 | 17.4                   | 13.4      | 7.1            | 10.2           | 1.4     |
|            |                            | 67                | 1.4                         | 13.4      | 4200         | 0.9                       | 2.8                       | 5.6                 | 17.6                   | 6.2       | 5.2            | 10.2           | 2.0     |
|            |                            | 78                | 1.6                         | 20.4      | 4610         | 1.4                       | 4.4                       | 8.3                 | 26.3                   | 6.0       | 4.6            | 10.2           | 2.2     |
|            |                            | 49                | 1.0                         | 16.4      | 6500         | 1.6                       | 5.0                       | 4.5                 | 14.2                   | 2.9       | 4.0            | 10.2           | 2.5     |
| B517       | Costa Rica Decollement    | 254               | 11.6                        | 97.1      | 24990        | 2.8                       | 8.9                       | 5.8                 | 18.4                   | 2.1       | 5.6            | 7.8            | 1.4     |
|            |                            | 52                | 2.7                         | 52.3      | 13410        | 2.2                       | 7.1                       | 5.0                 | 15.7                   | 2.2       | 2.0            | 7.5            | 3.8     |
| B561       | Costa Rica Inputs (chalk) | 2466              | 40.3                        | 12.8      | 1            | 0                         | 0                         | 0                   | 10386.0                 | -         | 225.9          | 9.6            | 0.0     |
| B575       | Barbados                  | 725               | 29.1                        | 180.2     | 48520        | 1.5                       | 4.6                       | 11.1                | 35.1                   | 7.6       | 6.9            | 8.9            | 1.3     |
|            |                            | 93                | 5.1                         | 48.0      | 21210        | 1.0                       | 3.0                       | 3.8                 | 12.0                   | 4.0       | 3.8            | 7.4            | 1.9     |
|            |                            | 880               | 32.4                        | 188.6     | 39800        | 0.9                       | 3.0                       | 13.4                | 42.2                   | 14.2      | 8.5            | 8.8            | 1.0     |
| B628       | Hikurangi Inputs          | 23                | 0.5                         | 16.5      | 4943         | 0.9                       | 2.8                       | 4.7                 | 14.9                   | 5.3       | 2.5            | 8.2            | 3.2     |
|            |                            | 23                | 0.5                         | 13.5      | 5050         | 1.5                       | 4.8                       | 4.5                 | 14.2                   | 2.9       | 2.5            | 8.9            | 3.6     |
| B675       | Hikurangi Inputs          | 0                 | 0                           | 0         | 0            | 1.7                       | 1.7                       | 5.3                 | 5.3                    | 1         | 0              | -              | -       |
| B678       | Hikurangi Inputs          | 36                | 0.8                         | 5.8       | 3976         | 1.7                       | 5.3                       | 5.9                 | 18.7                   | 3.5       | 13.7           | 8.9            | 0.7     |
|            |                            | 30                | 0.7                         | 11.2      | 2378         | 1.5                       | 4.7                       | 6.1                 | 19.3                   | 4.1       | 3.1            | 8.9            | 2.9     |
|            |                            | 45                | 1.0                         | 11.8      | 4558         | 1.4                       | 4.3                       | 3.7                 | 11.6                   | 2.7       | 8.0            | 8.9            | 1.1     |
| B560       | Alpine Fault (DFDP-1B)    | 128               | 2.6                         | 20.1      | 3130         | 0.7                       | 2.3                       | 12.7                | 40.2                   | 17.7      | 7.6            | 10.2           | 1.3     |
| Experiment | Location               | Stress drop (kPa) | Percentage stress drop | Slip (μm) | Duration (s) | Pre-SSE velocity (nm s\(^{-1}\)) | Pre-SSE velocity (cm yr\(^{-1}\)) | Peak velocity (nm s\(^{-1}\)) | Peak velocity (cm yr\(^{-1}\)) | \(V_p/V_s\) | \(K_s\) (MPa mm\(^{-1}\)) | \(K\) (MPa mm\(^{-1}\)) | \(K/K_s\) |
|------------|------------------------|-------------------|------------------------|-----------|-------------|----------------------------------|----------------------------------|-------------------------------|-------------------------------|------------|------------------------|--------------------|----------|
| pre-SSE    |                        | 138               | 2.8                    | 18.9      | 3580        | 0.8                              | 2.5                              | 12.1                          | 38.2                          | 15.2       | 10.4                   | 10.2               | 1.0      |
|            |                        | 155               | 3.2                    | 20.7      | 2920        | 0.5                              | 1.4                              | 17.0                          | 53.7                          | 37.3       | 11.3                   | 10.2               | 0.9      |
|            |                        | 155               | 3.1                    | 23.2      | 3190        | 0.5                              | 1.5                              | 14.7                          | 46.4                          | 30.0       | 8.9                    | 10.2               | 1.1      |
|            |                        | 155               | 3.2                    | 23.9      | 2560        | 0.6                              | 2.0                              | 21.9                          | 69.2                          | 35.0       | 8.5                    | 10.2               | 1.2      |
|            |                        | 155               | 3.2                    | 23.9      | 2560        | 0.6                              | 2.0                              | 21.9                          | 69.2                          | 35.0       | 8.5                    | 10.2               | 1.2      |
|            |                        | 155               | 3.2                    | 23.9      | 2560        | 0.6                              | 2.0                              | 21.9                          | 69.2                          | 35.0       | 8.5                    | 10.2               | 1.2      |
| B713 Alpine Fault (DFDP-1B) | 230               | 4.6                    | 31.6      | 3200        | 0.4                              | 1.4                              | 22.2                          | 70.0                          | 49.9       | 9.3                    | 10.2               | 1.1      |
|            |                        | 245               | 4.9                    | 28.8      | 2700        | 0.4                              | 1.3                              | 19.8                          | 62.5                          | 48.6       | 10.7                   | 10.2               | 1.0      |
|            |                        | 244               | 4.9                    | 31.3      | 3100        | 0.7                              | 2.1                              | 20.9                          | 66.0                          | 31.1       | 9.8                    | 10.2               | 1.0      |
|            |                        | 255               | 5.1                    | 35.7      | 2200        | 0.4                              | 1.4                              | 28.1                          | 88.6                          | 64.1       | 8.3                    | 10.2               | 1.2      |
|            |                        | 244               | 4.9                    | 31.9      | 3000        | 0.5                              | 1.6                              | 23.2                          | 73.2                          | 46.2       | 8.8                    | 10.2               | 1.2      |
| B571 San Andreas Fault CDZ | 187               | 14.0                  | 62.4      | 19760       | 1.1                              | 3.3                              | 6.2                           | 19.5                          | 5.9        | 4.4                    | 7.1                | 1.6      |
| B594 San Andreas Fault CDZ | 217               | 15.3                  | 877.0     | 98440       | 8.1                              | 25.5                             | 15.2                          | 47.9                          | 1.9        | 1.5                    | 7.2                | 4.9      |
|            |                        | 364               | 27.5                  | 953.0      | 109300      | 7.2                              | 22.6                             | 20.8                          | 65.5                          | 2.9        | 2.3                    | 7.0                | 3.1      |
|            |                        | 364               | 27.5                  | 953.0      | 109300      | 7.2                              | 22.6                             | 20.8                          | 65.5                          | 2.9        | 2.3                    | 7.0                | 3.1      |
|            |                        | 401               | 32.4                  | 788.0      | 81500       | 6.7                              | 21.3                             | 14.7                          | 46.5                          | 2.2        | 2.0                    | 6.8                | 3.5      |
| B570 Woodlark Basin | 16                | 0.9                    | 7.8       | 3710        | 1.5                              | 4.7                              | 2.8                           | 8.9                           | 1.9        | 2.5                    | 7.4                | 3.0      |
|            |                        | 47                | 2.7                    | 27.3       | 12710       | 1.6                              | 4.9                              | 4.0                           | 12.5                          | 2.5        | 3.4                    | 7.4                | 2.2      |
|            |                        | 32                | 1.9                    | 17.7       | 7410        | 1.5                              | 4.9                              | 2.7                           | 8.7                           | 1.8        | 3.7                    | 7.4                | 2.0      |
|            |                        | 30                | 1.7                    | 17.6       | 7900        | 1.6                              | 4.9                              | 4.1                           | 12.8                          | 2.6        | 4.5                    | 7.4                | 1.7      |
Table 3. RSF parameters from inverse modelling of velocity step data.

| Experiment Location | V_0 (μms⁻¹) | V (μms⁻¹) | a | b | D_0 | D_1 | D_2 | \text{\textit{a}} \text{-} \text{\textit{b}} | \text{\textit{std a}} | \text{\textit{std b}} | \text{\textit{std D_0}} | \text{\textit{std D_1}} | \text{\textit{std D_2}} |
|---------------------|-------------|-----------|---|---|-----|-----|-----|-------------|---------|---------|-----------|-----------|-----------|
| B651 Japan Trench   | 0.0027      | 0.0087    | 0.0021 | 0.0015 | 0.0017 | 0.0017 | 0.0017 | 0.0008 | 0.0003 | 0.0002 | 0.0004 | 0.0003 | 0.0004 |
| B518 Nankai Trough Megasplay | 0.0020 | 0.0050 | 0.0020 | 0.0010 | 0.0012 | 0.0012 | 0.0012 | 0.0010 | 0.0009 | 0.0007 | 0.0009 | 0.0008 | 0.0011 |
| B611 Cascadia       | 0.0020      | 0.0050    | 0.0020 | 0.0010 | 0.0012 | 0.0012 | 0.0012 | 0.0010 | 0.0009 | 0.0007 | 0.0009 | 0.0008 | 0.0011 |
| B517 Costa Rica Decollement | 0.0020   | 0.0050    | 0.0020 | 0.0010 | 0.0012 | 0.0012 | 0.0012 | 0.0010 | 0.0009 | 0.0007 | 0.0009 | 0.0008 | 0.0011 |
| B541 Nankai Trough Decollement | 0.0020 | 0.0050 | 0.0020 | 0.0010 | 0.0012 | 0.0012 | 0.0012 | 0.0010 | 0.0009 | 0.0007 | 0.0009 | 0.0008 | 0.0011 |
| B597 Barbados        | 0.0020      | 0.0050    | 0.0020 | 0.0010 | 0.0012 | 0.0012 | 0.0012 | 0.0010 | 0.0009 | 0.0007 | 0.0009 | 0.0008 | 0.0011 |
| B570 Woodlark Basin  | 0.0020      | 0.0050    | 0.0020 | 0.0010 | 0.0012 | 0.0012 | 0.0012 | 0.0010 | 0.0009 | 0.0007 | 0.0009 | 0.0008 | 0.0011 |

In some cases there is little to no slip between the loading and the stress drop (Fig. 11B), whereas in other cases significant pre-SSE slip (up to 190 μm) occurs before the stress drop (cf. Fig. 6 in Rabinowitz et al. 2018).

The peak slip velocities of the laboratory SSEs are comparable to the lower range of southern Hikurangi SSE slip rates of 9–110 cm yr⁻¹, but are significantly slower than the northern Hikurangi SSEs, which exhibit peak slip velocities of at least 73 cm yr⁻¹ (Wallace & Beavan 2010; Wallace et al. 2012). As reported by Ikari et al. (2019), the form of the slip history curve of the laboratory SSEs closely resembles the displacement history during SSEs from cGPS measurements, and the general comparison between laboratory and natural Hikurangi SSEs for the stress drop and peak slip velocity data at 10 MPa effective stress also applies to tests conducted at 1, 5 and 15 MPa.

5.7 Alpine Fault, New Zealand

Two experiments were conducted on a sample of the principal slip zone of the Alpine Fault recovered in borehole DFDP-1B. Both experiments showed repetitive, uniform stick-slip-like events (Fig. 12), which are similar to but clearly slower and less energetic than the events seen in the Costa Rica chalk. Representative sequences of 5 consecutive events were analysed from each experiment, which showed differences in stress drop (128–155 kPa and 230–255 kPa) but similarities in peak slip velocity (≈40–70 cm yr⁻¹ and ≈60–90 cm yr⁻¹) and unloading stiffness (≈8–11 MPa mm⁻¹ for both experiments). Unlike SSEs in the other samples in this study (except the Costa Rica chalk), the interevent slip rate approaches zero. There is an observable phase of increasing slip rate prior to the stress drop, as also observed for the Costa Rica chalk sample. The unloading stiffness in this case can be seen to be much lower than the stiffness during the loading phase prior to the stress drop.

On the Alpine Fault, the shallow portion is locked and slow slip and tremor are only observed deeper than 20–25 km (Wech et al. 2012; Chamberlain et al. 2014); thus far there has been no indication of shallow slow slip on the Alpine Fault. The repetitive slip events observed in the laboratory tests on the DFDP-1B sample indicate the potential for shallow slip instability which is consistent with the near-surface locking, therefore slip events would be expected to occur as small magnitude earthquakes (e.g. Leitner et al. 2001), not SSEs or VLFEs.

5.8 San Andreas Fault

Two experiments on a sample from the San Andreas Fault Central Deforming Zone produced four total SSEs. One of these SSEs occurred at a driving rate of 5.3 cm yr⁻¹, but the other three were generated at a driving rate of 25 cm yr⁻¹ (10 times the long-term slip rate, e.g. Titus et al. 2005). The San Andreas Fault SSEs exhibit stress drops of ≈190–400 kPa, peak slip velocities of 20–66 cm yr⁻¹, and unloading stiffnesses of 1.5–4.4 MPa mm⁻¹. The one SSE generated at 5.3 cm yr⁻¹ has the smallest stress drop, lowest peak slip velocity and the highest unloading stiffness (Fig. 13). The San Andreas Fault SSEs bear several similarities to SSEs produced in the Barbados sample, including the broad form of the stress curve, large event slip, large percentage stress drop, and large slip deficit accumulation during loading. An interesting feature of the SSE generated at 5.3 cm yr⁻¹ is a distinct two-phase stress drop, with an initially small, steep drop before a more gradual drop. The initial drop is associated with the peak slip velocity, with the velocity during the...
Figure 5. Experimental data for a sample from the Nankai Trough megasplay. (a) Coefficient of friction as a function of displacement. (b) Closeup of SSEs generated during plate-rate shearing. (c) Closeup of two strength perturbations as candidate SSEs. (d) Sample slip velocity and, (e) sample displacement with the load point displacement trend removed for the two friction perturbations (shading). Note that only the first perturbation shows a slip velocity increase and excess slip above the driving rates, meaning that the first perturbation is considered an SSE and the second is not.

gradual drop being smaller but still elevated above the driving rate. This feature also occurs in the SSEs at the faster driving rate, but is not as prominent.

Based on data presented by Gladwin et al. (1994), the slip rate of northern San Andreas Fault creep events near Hollister can be roughly estimated to range from $\sim 3$ cm yr$^{-1}$ (slightly higher than the long-term slip rate) to 73 cm yr$^{-1}$ (nearly 30 times the long-term slip rate). Similar estimates from the creepmeter data of Goulty & Gilman (1978) for the southern creep events near Parkfield are approximately 10–38 cm yr$^{-1}$, or 4–15 times the long-term slip rate. A curiosity of the SSEs observed in the SAFOD gouge in this study is that the SSEs occurred at driving rates that are 2–10 times faster than the plate rate of $\sim 2.5$ cm yr$^{-1}$ on the San Andreas Fault. However, all four laboratory SSEs exhibit peak slip velocities that are $\sim 2–6$ times the driving rate, consistent with the lower range of the creepmeter observations.

5.9 Woodlark Basin, Papua New Guinea

The sample from the Woodlark Basin produced four regularly-spaced SSEs. The first SSE is notably smaller than the other three in terms of stress drop (16 kPa) and event slip and duration. The other three SSEs are remarkably similar (Fig. 14). These three SSEs exhibit stress drops of 30–47 kPa, peak slip velocities of $\sim 9-13$ cm yr$^{-1}$ and unloading stiffnesses of 3.4–4.5 MPa mm$^{-1}$. Unlike most of the SSEs in this study, large slip deviations (i.e. slip deficit and/or excess) are not observed. In addition to their periodicity (roughly every $\sim 0.5$ mm) a noticeable feature of these three Woodlark SSEs is the $\sim 110–150$ μm of pre-event slip between the loading phase and the stress drop, which is similar to the behaviour of the Cascadia and some of the Hikurangi SSEs. Shallow slow slip activity on the Woodlark Basin detachment normal fault has not been documented, but the laboratory observations presented here suggest that SSEs are possible in this region.

6 MECHANISMS OF SLOW SLIP BEHAVIOUR

6.1 Slow slip in the context of critical stiffness theory

Since observational studies of plate-boundary fault motion have generally shown that slow slip tends to occur near the edges of geodetically locked zones, a common assumption is that slow slip arises at the transition from stable to unstable frictional behaviour
The RSF parameters from velocity steps (Table 3) can be used to determine the frictional properties of natural fault materials that exhibit slow slip, and examine any potential relationship between the RSF and SSE parameters. The RSF formulation (eqs 1–3) provides a quantitative means for determining whether stable or unstable slip is expected, assuming the system can be approximated by a simple spring-slider model (e.g. Rice & Ruina 1983; Gu et al. 1984). The data presented here show that changes in shear stress are approximately proportional to the deviation between load point velocity and sample slip velocity (Figs 4, 5, 7, 8, 10–14, Table 2). In some cases the proportionality is not always constant, however in most cases they are sufficient for the system to be considered an approximate, if not ideal, spring-slider system.

The criterion for slip instability, defined as self-acceleration to slip speed only limited by inertia, is:

$$K < K_c = \frac{(b-a) \sigma_\gamma}{d_c},$$

where $K$ is the stiffness of the fault surroundings, which must be lower than the critical stiffness value $K_c$ (Dieterich 1981, 1986; Rice & Ruina 1983; Gu et al. 1984; Dieterich & Kilgore 1996a; Scholz 1998; Marone 1998a). For simplicity, I assume that $d_c$ is $d_{c1}$ from eqs (1) and (2) in the case that two state variables are employed. It can be seen from eq. (4) that if $b-a < 0$, indicating a velocity-strengthening material, it is impossible to satisfy the condition $K < K_c$, and therefore instability should not occur. If this condition is satisfied, audible stick-slip events occur in laboratory experiments which are analogous to ordinary earthquakes (Brace & Byerlee 1966; Cook 1981). Thus, the concept of a critical stiffness for unstable slip was used in many early studies to describe the seismic cycle (e.g. Cao & Aki 1986; Dieterich 1986, 1992, 1994; Okubo 1989; Rice 1993; Dieterich & Kilgore 1996a).

The critical stiffness criterion described by eq. (4) is useful for predicting slip instabilities that are analogous to ordinary earthquakes, but it is not certain whether various types of slow slip should also be considered instabilities. For example, SSEs do not emit seismic signals and have limited velocity (usually a few times the plate rate) and therefore may not be true slip instabilities, even though they may also nucleate and accelerate. Therefore, I do not make an a priori assumption as to whether the experimentally observed slip events should be classified as instabilities or not. I consider a general model of slip event nucleation that sequentially includes: a perturbation in the initial slip velocity, accelerating slip on a growing fault patch, and finally runaway slip acceleration (instability) after a critical patch size has been attained (e.g. Dieterich 1986, 1992; Roy & Marone 1996; Perfettini & Ampuero 2008). The initial acceleration phase for an event that later becomes fully unstable is the slowly slipping earthquake nucleation phase (Iio 1992, 1995; Ellsworth & Beroza 1995). If it does not become a full instability it can be considered a slow slip event, assuming that the nucleation process is independent of the final slip velocity that is attained. The criterion for the accelerating slip phase before instability is:

$$K < K_b = \frac{b \sigma_\gamma}{d_c} + \frac{T}{V_s},$$

where $T$ is an external shear stressing rate $dr/dt$ and $V_s$ is the initial slip velocity; this second term on the right-hand side is commonly neglected (e.g. Dieterich 1992). As noted by Dieterich (1992), since the parameter $b$ rather than $b-a$ appears in eq. (5), it is possible for velocity-strengthening materials to exhibit accelerating slip, a property that has been successfully used in numerical simulations of slow slip and earthquake afterslip (Perfettini & Ampuero 2008; Kanu & Johnson 2011).
Figure 7. Experimental data for a sample from the southern Cascadia frontal thrust. (a) Coefficient of friction as a function of displacement. (b) Closeup of SSEs generated during plate-rate shearing. (c) Shear stress, (d) sample slip velocity, and (e) difference between sample displacement and load point displacement for the SSE indicated in (b). Shading indicates the SSE stress drop. Note the significant steady-state slip between the loadup and stress drop in (c).

6.2 Critical stiffness evaluation

Values of $b-a$ and $b$ from velocity upsteps from 0.0017 to 0.0028 $\mu$m s$^{-1}$ are compared with the size of SSEs generated in the same material. Here, the SSE size is quantified as the ratio of the peak slip velocity $V_p$ to the initial slip velocity $V_i$ measured prior to the stress drop (Table 2). The velocity increase was chosen to represent SSE size rather than stress drop for two reasons: (1) stress drops without a distinguishable velocity increase were sometimes observed and (2) the velocity increase in the laboratory SSEs can be directly related to natural faults by comparison with geodetic measurements. Fig. 15 shows that the largest SSEs tend to occur in materials which exhibit large values of $b-a$, which is generally consistent with critical stiffness theory. However, there are several instances of SSEs occurring in samples that exhibit velocity-strengthening behaviour (i.e. negative $b-a$) and $b-a$ values that are essentially zero, which suggest the inability to produce a stress drop and therefore are inconsistent with the appearance of SSEs in these samples.

Comparing the SSE velocity increase with the parameter $b$ shows a similar relationship as observed with $b-a$ despite significant scatter, showing generally higher $V_p/V_i$ for samples with larger $b$ (Fig. 15). A key aspect of the critical stiffness criterion for accelerating stable slip using $K_s$ is that nearly all samples have the capability for stress drop, and therefore stably accelerating slip can occur even in velocity-strengthening materials. The exception is the sample from the Woodlark Basin, which exhibits a negative value of $b$ and still produced SSEs.

Using eqs (4) and (5), the critical stiffness values for instability $K_s$ and for accelerating slip $K_b$ can be calculated. Incorporating the measured apparatus stiffness $K$ is a useful indicator for the type of slip expected, i.e. unstable for $K/K_s < 1$ and stable accelerating for $K/K_b > 1$ and $K/K_s < 1$. Stable creep, that is lack of accelerating slip, is expected if $K/K_s > 1$. Note that the essence of these stiffness criteria is that the stiffness of the surroundings must be lower than the ability of the sample to unload stress. Therefore, the directly measured unloading stiffness during the SSE stress drop is also a critical stiffness value, designated here $K_s$. Thus, there are three permutations of critical stiffness from these experiments which may be compared.

Comparison of critical stiffness values shows that in most cases (Japan Trench, Costa Rica décollement, Barbados, San Andreas Fault, Woodlark Basin) $K_s$ values are much larger than $K_s$ and $K_b$, some of which are vanishingly small (Fig. 16). For the Nankai
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**Figure 8.** Experimental data for a sample from the Costa Rica décollement. (a) Coefficient of friction as a function of displacement. (b) Closeup of shear stress for both SSEs generated during plate-rate shearing. (c) Sample slip velocity, (d) sample displacement with the load point displacement trend removed and (e) unloading stiffness for the two SSEs. Shading indicates the SSE stress drops.

Trough megasplay, Cascadia and Hikurangi, $K_b$ more closely matches the SSE unloading stiffnesses. $K/K_s$ values range from 0.7 to 4.9, with the exception of an exceptionally small value of 0.04 observed for the large events in the Costa Rica chalk. In contrast, many of the $K/K_c$ values have exceedingly large values ranging up to $\sim 160$. These large values are clearly due to the small values of $K_c$ calculated from the velocity step data. Similar to the $K_c$ data, $K/K_c$ is only consistent with $K/K_b$ for the Nankai Trough megasplay and Hikurangi samples. In the cases of the Costa Rica décollement, San Andreas Fault, and Woodlark Basin, $K/K_c$ could not be calculated due to velocity-strengthening behaviour, also the case for $K/K_b$ for the Woodlark Basin due to negative $b$ values.

A small number of $K/K_s$ values are $< 1$, observed for the Japan Trench, Costa Rica chalk, Hikurangi and Alpine Fault samples (Fig. 16). The Hikurangi sample also exhibits a $K/K_s$ value $< 1$. The Costa Rica décollement, Barbados, San Andreas Fault, and Woodlark Basin did not exhibit any stiffness ratios $< 1$, regardless of the method used in the calculation. Repetitive stick-slip-like events generated in the Costa Rica chalk sample exhibited a $K/K_c$ value of 0.04, and those for the Alpine Fault sample which exhibited $K/K_c$ of 0.9–1.3. Note that the stick-slip-like events in the Costa Rica chalk and the Alpine Fault samples did not allow the extraction of RSF parameters to facilitate comparison.

This analysis implies that the SSEs observed in these laboratory experiments, with the possible exception of the Costa Rica chalk and Alpine Fault events, are not frictional instabilities but instances of accelerating stable slip. Using $K_c$ from eq. (4) is therefore not an appropriate predictor of SSE behaviour, whereas $K_b$ and $K_s$ appear to be more consistent with the observed SSE behaviour. However, using $K_b$ and $K_c$ only predict accelerating slip in a few cases, which is not consistent with the majority of the laboratory SSE data set. One possibility is that the measured ratio $b\sigma_n'/d\epsilon$ may underestimate the stiffness of the sample. However, direct measurement of the peak unloading stiffness during velocity step tests showed values similar to $K_c$, therefore this possibility can be ruled out.

Another possibility is that SSEs can be generated within a range of $K/K_b$ or $K/K_s$ values from positive to negative, in a continuum that does not exhibit a sharp transition or divergent behaviour at the critical value. For the data in this study, this suggests that accelerating stable slip can occur for $K/K_b$ values up to $\sim 5$. $K/K_s$ values are significantly more scattered, but separate into two populations: $K/K_b < 3$ (Nankai Trough, Cascadia, Hikurangi), and $K/K_s > 19$ (Japan Trench, Costa Rica decollement, Barbados and San Andreas Faults).
Figure 9. Experimental data for a sample of pelagic chalk from the Costa Rica input sediment. (a) Coefficient of friction as a function of displacement. (b) Closeup of shear stress, and (c) sample displacement as a function of time for the slip events generated during plate-rate shearing. (d) Closeup of shear stress and sample displacement as a function of time for a single slip event, showing that the slip velocity is $\sim 10 \, \mu \text{m s}^{-1}$.

Figure 10. Experimental data for a sample from the Barbados décollement. (a) Coefficient of friction as a function of displacement. Closeup of (b) shear stress, (c) sample slip velocity and (d) difference between sample displacement and load point displacement for the SSE indicated in (a) generated during plate-rate shearing. Shading indicates the SSE stress drop. Note the exceptionally large percentage stress drop.
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Figure 11. Experimental data for a sample from the Hikurangi input sediment. (a) Coefficient of friction as a function of displacement. Closeup of (b) shear stress, (c) sample slip velocity and (d) sample displacement with the load point displacement trend removed for the SSE indicated in (a) generated during plate-rate shearing. Shading indicates the SSE stress drop.

Fault). Slow stick slip occurring in combination with stiffness ratios slightly greater than 1 were also observed by Scuderi et al. (2016) and Leeman et al. (2016), albeit for a different material (quartz silt) and driving velocity (10 μm s\(^{-1}\)). For the purposes of discussing the current data, I will consider the slip events observed in these laboratory experiments to be SSEs originating from accelerating stable slip after a perturbation but without reaching true instability. Exactly why these events occur for stiffness ratios \(>1\) should be examined in future research. An exception is the Costa Rica chalk sample, which may represent LFE or VLFE and emit inaudible low-frequency signals although this must be verified with further measurements.

Although the critical stiffness formulation presents difficulties in predicting whether SSEs will occur, laboratory SSE behaviour collectively does follow a pattern consistent with expectations from critical stiffness theory. Ikari & Kopf (2017) showed that larger peak slip velocities correlated with higher percentage stress drops for laboratory SSEs, a characteristic also observed for stick-slip instabilities in granite (Lockner & Okubo 1983; Ohnaka et al. 1987; Beeler et al. 2012). Fig. 17 shows the important observation that the size of an SSE correlates with increasing stiffness, and therefore decreasing \(K/K_s\). Therefore, the systematics of SSEs, as accelerating stable slip events, are similar to those expected for full slip instabilities.

6.3 Conceptual model of SSE generation

Dieterich (1992) demonstrated that the critical stiffness for stably accelerating slip following a perturbation is \(\sigma c \dot{d} \theta \) (eq. 5). A specific requirement here is that the instantaneous slip velocity \(V \gg \dot{d}/\theta\), so that the system is far away from steady-state. This makes the term \(V\dot{d}/\theta\) large so that the 1 in eq. (2) can be neglected. Since the time-dependence of friction arises from the 1 in the aging law, the approximation neglecting the 1 has been referred to as the ‘no healing’ condition (Rubin & Ampuero 2005). However, in both natural SSEs and the laboratory SSEs observed here, the peak slip velocities are often small and close to the plate tectonic driving rates. In eq. (2), it can be seen that there is another way to deviate from steady state, and that is with a very large \(\theta\) (and/or small \(d_c\), which I do not consider here for simplicity). Although in this case \(V\dot{d}/\theta >> 1\), it cannot be referred to as a no-healing condition because large \(\theta\) indicates that the healing process is dominant.

A key question is how the nucleation of slip acceleration to a higher velocity can be initiated in the first place, including the nature of the initial perturbation. In nature, such a perturbation could be driven externally, for example from movement on a neighboring fault or fault patch. However, the experiments here show that SSEs occur spontaneously with no apparent external influence. Previous numerical simulations of the nucleation process have shown that the condition \(V\dot{d}/\theta >> 1\) is necessary for initially stably accelerating slip (Dieterich 1992; Rubin & Ampuero 2005; Ampuero & Rubin 2008; Perfettini & Ampuero 2008; Rubin 2008), however these studies also noted that it is not always straightforward when the condition \(V\dot{d}/\theta >> 1\) is valid because \(V\) and \(\theta\) are inversely related. Therefore, \(V\) and \(\theta\) could drive the system away from steady state in combination. I suggest that the initiation of stable accelerating slip that results in SSEs could arise from advanced healing, that is rapid time-dependent frictional strengthening due to advanced evolution of \(\theta\). This process can be summarized as follows: (1) at initially steady-state sliding at the plate rate, state evolution during sliding begins to outpace the slip rate, increasing \(\theta\) so that additional healing during sliding occurs, (2) the additional healing forces \(V\) to
Figure 12. Experimental data for a sample from the Alpine Fault DFDP-1B principal slip zone. (a) Coefficient of friction as a function of displacement. (b) Closeup of the slip events generated during plate-rate shearing. Closeup of (c) shear stress, (d) sample slip velocity, (e) difference between sample displacement and load point displacement and (f) unloading stiffness for a sequence of five slip events indicated in (b). Shading indicates the SSE stress drop. Note the large peak slip velocities and low interevent velocity.

decrease to a small but non-zero value, which allows $\theta$ to become large and drives the system away from steady state, (3) failure of the strengthened material is reached, resulting in an increase in $V$, which functions as the initial slip rate perturbation, and (4) the perturbation in $V$ causes $\theta$ to decrease, driving $V$ higher so that $V >> \theta/d_c$ and slip may accelerate according to eq. (5). This is illustrated semi-schematically in Fig. 18; data from the first SSE in the Costa Rica décollement sample (Fig. 8) is used as an example for the shear stress and slip velocity. In this figure, the quantity $\theta$ is shown schematically, since away from steady state $\theta \neq d_c/V$ and hence is not well constrained.

One key feature of this model is a transition from a healing-dominated departure from steady-state to a velocity-dominated departure from steady-state. This transition necessitates a point at which $V$ and $\theta$ are balanced and steady-state is reached. This near-steady-state condition is likely short-lived, however a temporary steady-state condition explains the observation of a measurable amount of steady-state slip after the loading phase and before the stress drop of some of the SSEs, one example being the SSEs in the Cascadia sample.

Another important aspect of this model is that the initial perturbation is in the slip velocity, not necessarily in the shear stress as is usually employed in numerical models (e.g. Roy & Marone 1996; Perfettini & Ampuero 2008), although the velocity and shear stress are of course closely related. For most of the SSEs in these experiments the pre-event slip velocity $V_i$ is below the driving rate, ranging from 1.3 to 8.9 cm yr$^{-1}$ for driving rates of 5.3–8.7 cm yr$^{-1}$. Exceptions are the San Andreas Fault, which exhibited some SSEs at a higher driving rate (25 cm yr$^{-1}$) and the Costa Rica chalk events.
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Figure 13. Experimental data for a sample from the San Andreas Fault CDZ. (a) Coefficient of friction as a function of displacement. Closeup of (b) shear stress, (c) sample slip velocity and (d) difference between sample displacement and load point displacement for the SSE indicated in (a).

which exhibited $V_i \sim 0$. All but four of the SSEs discussed in this study have initial slip velocities that drop below the uncertainty in slip velocity defined by noise in the displacement sensors. Normalizing the initial slip velocity and peak slip velocity by the driving (load point) velocity for all slip events in this study shows a clear inverse relationship between $V_i/V_{lp}$ and $V_p/V_{lp}$ (Fig. 19). This demonstrates that the farther the velocity decreases below the driving rate, the larger the eventual peak slip velocity becomes, but that the velocity does not need to approach 0. For natural faults this means that full geodetic locking is not a requirement for SSE generation, but larger events result from stronger locking. This is consistent with the observation that ordinary earthquakes result from true geodetic locking, and also with geodetic measurements during the interval between repeating SSEs, which document a decrease in velocity that does not necessarily reach full locking before the slip reversal signaling the event (McCaffrey 2009; Vergnolle et al. 2010; Wallace & Beavan 2010; Jiang et al. 2012).

6.4 Role of frictional healing

The origin of the decrease in initial velocity or partial locking could be advanced healing which leads to large values of $\theta$. I suggest that such healing is facilitated by the extremely low slip velocities in these experiments, which allow the state to advance further than in standard friction experiments conducted in the $\mu$m s$^{-1}$ to mm s$^{-1}$ range. An advanced healing mechanism at cm yr$^{-1}$ driving rates is consistent with the RSF data in this study, including the large strength increase upon the velocity downsteps and the tendency for velocity-weakening friction (Fig. 3, Ikari & Kopf 2017). This mechanism is also consistent with a study by Kanu & Johnson (2011), who conducted numerical simulations of a creep event on the Hayward fault and concluded that evolution in $\theta$ from the aging law is necessary to simulate the event. The time- (state-) dependent evolution of friction known as frictional healing can be isolated and measured in slide-hold-slide friction experiments. In these tests, sliding at a constant rate is interrupted by periods of 0 driving rate (the 'hold') of various durations; the resulting excess peak in strength above the steady-state strength depends on the duration of the hold (Dieterich 1972). Most laboratory studies conduct SHS tests for hold times up to $\sim$3000 s, which show that frictional healing depends linearly on the logarithm of the hold time, and that at room temperature the rate of healing varies depending on the gouge composition, background driving rate, and normal stress (e.g. Beeler et al. 1994; Marone 1998b; Niemeijer & Spiers 2006; Ikari et al. 2012; Tesei et al. 2012; Carpenter et al. 2016). However, healing tests up to 3000 s are insufficient to explain the observations at low driving velocities near the plate rate. For example, assuming a grain length of 100 $\mu$m, an asperity contact being displaced at 1 $\mu$m s$^{-1}$ would exist for 100 s; whereas at 1.7 mm s$^{-1}$ (5.3 cm yr$^{-1}$) this contact would last for $\sim$60 000 s.

Previous studies using the same San Andreas Fault sample tested here indicated very small healing rates at typical laboratory hold times of up to 3000 s (Carpenter et al. 2011, 2012, 2015a), which would suggest that a healing mechanism leading to partial locking and a decrease in slip rate would not be effective. This would therefore present difficulty explaining the observation of SSEs in this sample and also the repeating creep events on the San Andreas Fault. Using this same San Andreas Fault sample, Ikari et al. (2016a) demonstrated that for hold times up to 10$^6$ s, there is a distinct increase in healing at very large hold times (>3000 s). In this case,
Figure 14. Experimental data for a sample from the Woodlark Basin detachment normal fault. (a) Coefficient of friction as a function of displacement. Closeup of (b) shear stress, (c) sample slip velocity, (d) unloading stiffness and (e) difference between sample displacement and load point displacement for the three SSEs shown in (b).

the healing rate is better fit with a power law rather than a log-linear function. This suggests that extremely slow driving rates of cm yr⁻¹ are necessary to allow enough time for advanced healing to occur, which facilitates the partial locking leading to accelerating stable slip. Whether the other samples in this study also exhibit increased healing at very large hold times will require further testing. Power-law healing is also observed in calcite samples (Chen et al. 2015; Carpenter et al. 2015b), which could explain why the Costa Rica chalk sample exhibited full locking and very fast slip events.

On natural faults, mechanisms other than healing may also contribute to an initial decrease in slip velocity prior to SSEs. One possibility is geometric irregularities, such as subducting topographic highs or specific fault surface roughness. Another is heterogeneous or patchy strength, which could arise from spatially varying lithology or compartmentalized pore fluid pressure. These mechanisms do not contradict the effect of advanced healing proposed here; rather their effects likely enhance each other. For example, faults which have the ability to heal rapidly would be more sensitive to small perturbations that may be introduced by sliding against such geometric irregularities.

6.5 Effect of mineralogic composition

An intriguing aspect of the San Andreas Fault sample which exhibits advanced healing that could allow the generation of SSEs, is its notably high smectite content. This has led to the conclusion that smectite controls the mechanical behaviour of the central San Andreas Fault to a large degree (e.g. Carpenter et al. 2011, 2012, 2015a; Holdsworth et al. 2011; Lockner et al. 2011; Hadizadeh et al. 2012; Schleicher et al. 2012; Moore 2014; Ikari et al. 2016a). Three other notable fault zones which exhibit SSEs in this study have been previously implied to be controlled or heavily influenced by smectite: the Japan Trench (Kameda et al. 2015; Ikari et al. 2015a,b), Barbados (Deng & Underwood 2001; Kopf & Brown 2003), and Costa Rica (Spinelli & Underwood 2004) subduction zones. Smectite, and other expandable clays, are notable for having the unique property of strongly attracting water molecules to their surfaces and interlayers, heavily affecting their mechanical behaviour (e.g. Byerlee 1978; Lupini et al. 1981; Bird 1984; Ikari et al. 2007; Behnse & Faulkner 2013). For the samples used in this study, the Japan Trench sample is composed of 81 per cent smectite, the Costa Rica décollement 46 per cent smectite and 8 per cent mixed-layer clays, the San Andreas Fault sample 39 per cent smectite and 5 per cent.
mixed-layer clays, and the Barbados sample 11 per cent smectite and 9 per cent mixed-layer clays. Mixed-layer clays are typically alternating layers of smectite with illite (or another clay) and therefore included as expandable clays (Srodon 1999). All other samples in this study have an expandable clay component of 3–25 per cent. Note that it is these four samples which produced SSEs despite exhibiting very large $K/K_b$ values, which should lead to an expectation of stable non-accelerating slip rather than SSEs.

Strong evidence for an effect of expandable clays can be seen in the SSE stress drop data. Stress drop is one measure of slip event size and can be expected to correlate positively with peak slip velocity (e.g. Lockner & Okubo 1983; Ohnaka et al. 1987; Beeler et al. 2012). Therefore, larger percentage stress drops should also correlate with lower initial slip velocity, $V_i$. However, these relationships are obscured by significant scatter (Fig. 20). Closer inspection of the data shows that the scatter is caused by the samples with a significant expandable clay component: the Japan Trench, Costa Rica décollement, Barbados and the San Andreas Fault. After separating these samples out, the remaining percentage stress drop data show the expected relationships with $V_p/V_o$ and $V_p/V_i$ (Fig. 20). The scatter can thus be attributed to the extraordinarily large stress drops produced by the four expandable clay-bearing samples. It is not clear why the samples bearing expandable clay produce larger stress drops, but may be related to an advanced healing mechanism observed in the San Andreas Fault sample (Ikari et al. 2016a). Considering this compositional effect, a general correlation can be observed between both low overall clay and low expandable clay with $V_p/V_o$ (Fig. 21). However, the correlation between expandable clay and measured SSE parameters is not a strictly direct one and should be considered a general tendency; for example the Nankai Trough megasplay sample has a smectite content of 25 per cent but does not exhibit especially large stress drops.

### 6.6 Implications for general fault slip behaviour

Due to their potential effect on coseismic fault slip, it is of interest to discuss the implications of these laboratory SSEs in the context of the complete spectrum of fault slip behaviour. A caveat to these analyses is that these experiments were performed at 10 MPa effective normal stress and at room temperature, which represents the shallowest portions of major faults (~1–2 km). Although slow slip can propagate to such shallow levels (e.g. Wallace et al. 2016; Araki et al. 2017), care should be taken in comparing results from these experiments to natural fault slip.
testing conditions with natural SSEs that originate near the updip limit of geodetic locking and/or seismogenesis.

A key finding of this work and of Ikari & Kopf (2017) is that naturally slow, plate-rate shearing can result in velocity-weakening friction not observed under more commonly used shearing rates. Velocity-weakening friction suggests the potential for unstable slip, however the critical stiffness analysis performed here indicates that the majority of the SSEs observed in these experiments are accelerating stable slip events and not frictional slip instabilities. This nominally suggests that velocity-weakening friction on shallow, SSE-producing fault portions may not significantly enhance the risk of seismic slip.

On the other hand, the occurrence of SSEs does indicate the ability to store and release elastic strain energy, which could be a factor contributing to enhanced coseismic slip. There are two observations from the laboratory SSEs that may be especially relevant to coseismic slip propagation. The first is that the SSEs demonstrate the ability for velocity-strengthening faults to exhibit a stress drop, even the Woodlark Basin sample which is velocity-strengthening and exhibits a negative b. This implies that RSF does not always completely describe a materials’ ability to release stored energy. Secondly, during SSEs the frictional strength increases from a baseline level, before a stress drop of equal magnitude back to the baseline level. This implies that while extra strengthening may allow additional energy to be stored on SSE faults, the occurrence of the SSEs does not necessarily relieve stress below the background level. Therefore, based on the observations in this study, SSE-producing fault portions may be expected to enhance coseismic slip from a propagating earthquake rather than resisting it. However, the overall small stress drops observed in this study, the stably-sliding nature of the SSEs, and the small overall strength increase at plate-rate shearing suggests that this effect is likely to be small. As noted by Faulkner et al. (2011), weak clay-rich faults, which describes most of the samples in this study, are expected to provide little resistance against propagating earthquake slip despite velocity-strengthening frictional behaviour at intermediate slip rates. The effect of elastic strain energy accumulation in weak, SSE-generating materials on coseismic slip propagation merits further study.

As a final note, many of the experiments presented here were performed on samples from well-studied fault zones with records of recent large earthquakes (e.g. Nankai Trough, Japan Trench and Costa Rica). However, the complete data set suggests that SSEs can
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Figure 17. Ratio of SSE slip velocity increase $V_p/V_i$ as a function of (a) measured SSE unloading stiffness $K_s$ and (b) unloading stiffness ratio $K/K_s$.

Figure 18. Illustration of the conceptual model for SSE generation using Dieterich’s aging law. Shear stress and slip velocity data from the large SSE observed for the Costa Rica décollement sample shown as an example for reference. Note that the slip velocity curve follows the experimental data, while the curve for the state variable $\theta$ is shown as a schematic and is not meant to represent actual values.

Figure 19. Peak slip velocity normalized by the driving velocity $V_p/V_{ip}$ as a function of the initial pre-SSE velocity normalized by the driving velocity $V_i/V_{ip}$. The point 1:1 indicates no velocity perturbation and therefore no SSEs, which is the case for the Nankai Trough décollement and one experiment using the Hikurangi input sediment.

Figure 20. Variation of peak slip velocity with initial shear stress for each sample. The Costa Rica décollement and a sample of chalk from Barbados show evidence of SSEs, while the other samples do not.

7 CONCLUSIONS

This paper presents the results of laboratory friction experiments conducted on 11 natural samples from major fault zones around the world, conducted at cm yr$^{-1}$ shearing rates and simulating in situ conditions on the shallow (upper few km) region of major fault zones. The cm yr$^{-1}$ driving rates produce spontaneously occurring laboratory SSEs, which exhibit peak slip velocities and stress drops comparable to those of natural SSEs estimated from geodetic data. Generally, the stress drops range from 10s to several hundreds of kPa, or $>1$–32 per cent stress drop. Slip velocity tends to decrease below the driving rate prior to the SSE stress drops, then accelerates to peak slip velocities that generally range from $\sim9$ to $\sim90$ cm yr$^{-1}$, for driving rates of 5.3–8.7 cm yr$^{-1}$. An exception is a sample of
Figure 20. Percentage stress drop as a function of peak slip velocity normalized by the driving velocity $V_p/V_l$ (A, B, C) and as a function of the initial pre-SSE velocity normalized by the driving velocity $V_i/V_l$ (D, E, F). Panels (a) and (d) show the complete population of laboratory SSEs, (b) and (e) show samples from the Japan Trench, Costa Rica décollement, Barbados and San Andreas Fault which have an expandable clay content of 20–81 per cent and (c) and (f) show all other samples which have an expandable clay content of 3–25 per cent. Note that the Costa Rica chalk is only shown in panels (a) and (d) due to a very large value of $V_p/V_l = 3765$.

Figure 21. Ratio of SSE slip velocity increase $V_p/V_i$ as a function of (a) total phyllosilicate content and (b) expandable clay content.
pelagic chalk from the incoming plate of the Costa Rica subduction zone, which produced stick-slip like events with relatively rapid (10 μm s⁻¹) slip velocities and large (40 per cent) stress drops which are likely analogous to VLFEs rather than SSEs. Despite the overall similarity of the laboratory SSEs, some characteristics are observed which are specific to certain samples that, in many cases, are consistent with geodetic observations in the regions from which they were sampled.

The laboratory SSE data set shows a strong correlation between pre-SSE velocity and peak slip velocity, where a larger decrease in slip velocity before the event results in a faster event. The size of the SSEs, quantified by the slip velocity increase, correlates with higher values of unloading stiffness during the SSE stress drop, suggesting that SSEs behave similarly to true frictional instabilities. Results of velocity-stepping tests at cm yr⁻¹ rates show a tendency for velocity-weakening friction not observed at higher sliding velocities, and that the materials with lower values of the rate-dependent friction parameter a–b and larger values of b tend to produce faster SSEs. Using RSF measurements on samples that produce SSEs, critical stiffness analyses show that the condition for slip instability is not satisfied. Rather, the SSEs should be considered accelerating stable slip events, based on the observation that the stiffness condition for this type of slip behaviour is satisfied more often.

Based on Dieterich’s aging law, I develop a conceptual model for the nucleation of SSEs. The model utilizes a competition between the state variable θ and the instantaneous slip velocity V, which drives the system away from steady state in two phases: an initial ‘healing-dominated’ phase where advanced healing mechanism causes the onset of partial locking and velocity decrease, followed by failure of the strengthened material and the ensuing transition to a ‘velocity-dominated’ phase of accelerating (stable) slip. This model explains several aspects of the laboratory SSEs and is also consistent with geodetic measurements of natural SSEs, most notably the relationship between low pre-SSE velocity and high peak slip velocity.

The role of sample composition is explored, with the data showing that the frictional strength and velocity-dependence of friction depend on the phyllosilicate mineral content, consistent with previous studies. Faster SSEs tend to occur in low-phyllosilicate materials, but the presence of expandable clays (typically smectite) can result in exceptionally larger stress drops.

Applying the behaviour observed in SSE-producing samples to natural fault zones, the appearance of SSEs indicates the ability to strengthen above the steady-state plate-rate level, thereby storing elastic strain energy later released as the SSE stress drop. The data show that this can occur even in gouges that are velocity-strengthening and exhibit stably accelerating slip rather than slip instability. This can be interpreted to enhance coseismic slip in the specific case of an earthquake propagating onto the SSE-producing fault portion. However, several lines of evidence such as the fact that the criteria for slip instability is rarely satisfied and the small overall stress drops indicate that the enhancement of coseismic slip may be relatively small, although this effect warrants further study.

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