1 **The potential for supershear earthquakes in damaged fault zones**

2 **- Theory and observations**

3 Yihe Huang^{1,2*}, Jean-Paul Ampuero², Don V. Helmberger²

4 ¹Department of Geophysics, Stanford University, Stanford, CA 94305, USA;

5 ²Seismological Laboratory, Division of Geological and Planetary Sciences, California
6 Institute of Technology, 1200 E. California Blvd., Pasadena, CA 91125, USA.

7 *Correspondence to: yihe@stanford.edu

8

9

10

11

12

13

14

15

16

17

18

19

20

21

22

23

Abstract

The potential for strong ground shaking in large earthquakes partly depends on how fast the earthquake rupture propagates. It is observed that strike-slip earthquakes usually propagate at speeds slower than the Rayleigh wave speed (v_R) but occasionally jump to speeds faster than the S wave speed (v_s), or supershear speeds. Supershear earthquakes can be more catastrophic and cause unusually large ground motions at long distances. Here we use both fully dynamic rupture simulations and high-resolution seismic observations to show that supershear earthquakes can be induced by damaged fault zones, the low-velocity layers of damaged rocks that typically exist around major faults and serve as waveguides for high-frequency energy. In contrast to supershear ruptures in homogeneous media, supershear ruptures in damaged fault zones can occur under relatively low fault stress and propagate stably at speeds within the range usually considered as unstable.

Keywords: Earthquake rupture; Supershear; Damaged fault zone

1. Introduction

Several large strike-slip earthquakes have been observed to rupture unusually fast and propagate at speeds faster than the S wave speed (v_s), or supershear speeds, while most others stay at speeds slower than the Rayleigh wave speed (v_R). In Fig. 1, we summarize the estimates of rupture speeds for the large strike-slip supershear earthquakes, including the 1979 M_w 6.5 Imperial Valley earthquake [Archuleta, 1984; Spudich and Cranswick, 1984], the 1999 M_w 7.6 Izmit earthquake [Bouchon et al., 2001; Bouchon, et al., 2002;

47 Sekiguchi and Iwata, 2002] and subsequent M_w 7.1 Düzce earthquake [Bouchon et al.,
 48 2001; Birgören et al., 2004; Bouin et al., 2004; Konca et al., 2010], the 2001 M_w 7.8
 49 Kokoxili earthquake [Bouchon and Vallée, 2003; Antolik et al., 2004; Robinson et al.,
 50 2006; Vallée et al., 2008; Walker and Shearer, 2009], the 2002 M_w 7.9 Denali earthquake
 51 [Dunham and Archuleta, 2004; Ellsworth et al., 2004; Asano et al., 2005; Walker and
 52 Shearer, 2009], the 2010 M_w 6.9 Qinghai earthquake [Wang and Mori, 2012] and the
 53 2013 M_w 7.5 Alaska earthquake [Yue et al., 2013]. Note that we only consider crustal
 54 strike-slip earthquakes here. The 2013 Sea of Okhotsk earthquake, a M_w 6.7 earthquake
 55 at 642 km deep, was also reported to propagate at a supershear speed [Zhan et al., 2014].
 56 To distinguish among the observed supershear rupture speeds, we normalize the rupture
 57 speeds by the local v_s as stated in the data sources or the average crustal v_s (3.5 km/s)
 58 when the local v_s is not specified.

59 Fig. 1: Rupture speeds observed in large earthquakes (circles and diamonds), laboratory
 60

experiments (squares) and numerical models (triangles), normalized by S wave speeds. The circles denote the estimation of the average supershear speed for a portion of the fault, and diamonds with bars denote the estimations of the range of supershear speeds with mean values, from different observations of supershear earthquakes. The red strip indicates the forbidden speed range, and the grey patch indicates the unstable supershear range (SS) in homogeneous media. The two open squares display the limiting speeds of supershear earthquakes in laboratory experiments. The open triangle and grey triangles show the asymptotic speeds of supershear ruptures in homogeneous media and in damaged fault zones, respectively.

Both classical theories of dynamic fractures [Freund, 1979; Burridge et al., 1979; Broberg, 1999] and laboratory experiments [Rosakis et al., 1999] conducted in homogeneous materials find supershear ruptures can propagate stably at speeds between $\sqrt{2}v_s$ and v_p , referred to as stable supershear. In contrast, the speed range $v_s < v < \sqrt{2}v_s$ (gray patch in Fig. 1) is unstable, and ruptures in this speed range would either accelerate rapidly to stable supershear or decelerate to sub-Rayleigh speed. The speed range $v_R < v < v_s$ (red strip in Fig. 1) is forbidden for steady-state in-plane shear cracks because the energy flux into the crack tip becomes negative [Broberg, 1999]. Earthquake ruptures can propagate through this forbidden zone but only transiently [Bizzarri and Das, 2012]. However, the observed supershear ruptures show that both unstable and stable supershear speeds can occur in nature. Among them, the 2001 M_w 7.8 Kokoxili and the 2002 M_w 7.9 Denali earthquakes likely propagated at stable supershear speeds, while the 1979 M_w 6.5 Imperial Valley and the 1999 M_w 7.1 Düzce earthquakes ruptured more slowly at unstable supershear speeds. Depending on methods used to investigate supershear ruptures, the estimations may be either local over a limited rupture segment or

an average over the entire rupture length. Thus, it is possible that the observed unstable supershear speeds result from rupturing partly at sub-Rayleigh speeds and partly at stable supershear speeds. However, especially for the Düzce earthquake, different methods lead to consistent results, and an unstable supershear rupture is also required by the seismic records at a near-fault station [Bouchon et al., 2001; Bouin et al., 2004].

Fig. 2: Rupture speed variations along Big Bear fault and the correlation with fault zone structure. a, Rupture speeds of several of Mw 3.3-4.3 aftershocks in the 2003 Big Bear sequence are shown in colors. The star indicates the mainshock and colored circles are aftershocks for which rupture speeds have been estimated. Arrows indicate the rupture propagation directions, and their lengths are proportional to the rupture lengths. Grey circles are epicenters of relocated seismicity. b, Earthquakes propagating through the fault zone can have faster speeds than earthquakes outside the fault zone if the supershear transition is induced by the dynamic stress transfer carried by head waves.

Besides large strike-slip earthquakes, smaller ones (M_w 3 to 5) can also be studied in great detail if dense seismic array data is available, and some were inferred to propagate at unexpectedly fast speeds, probably equal or faster than the local S wave speeds [Luo et

al., 2010; Tan and Helmberger, 2010]. In particular, the observed rupture speeds of several M_w 3.3-4.3 aftershocks (strike-slip) in the 2003 Big Bear, CA sequence suggest that the earthquakes closer to the main fault were more likely to propagate at faster speeds [Tan and Helmberger, 2010] (Fig. 2a). The aftershock closest to the mainshock hypocenter (hereafter named event A) has a dominant strike-slip mechanism on a nearly vertical fault plane and shows strong rupture directivity. Assuming a 1D Haskell model and using a nearby smaller event as empirical Green's function (EGF), its rupture speed and rise time are solved by minimizing the sum of misfits between the data and the convolution of the source time function and the EGF over all stations. The minimum misfit is reached when the rupture speed is 3.62 km/s. However, because of the trade-off between rupture speed and rise time, the rupture speed of event A ranges between 3.28 km/s and 4.30 km/s, which falls in the forbidden zone or unstable supershear speed range given $v_s = 3.5$ km/s.

Here, we address the question of how supershear ruptures in nature can propagate stably at speeds within the unstable supershear range predicted by theoretical studies of dynamic fracture on faults embedded in a homogenous medium. The key point considered here is that natural faults are usually surrounded by damaged rocks. The damaged fault zones, characterized by low-velocity structures (as summarized in Table 1 in Huang et al., 2014), can lead to a complex wave field that modulates rupture properties [Harris and Day, 1997; Huang and Ampuero, 2011; Huang et al., 2014]. We first present fully dynamic models of supershear ruptures in damaged fault zones and show that such supershear ruptures can propagate stably in the speed range considered as unstable or forbidden in homogeneous media. We then investigate whether a damaged fault zone could induce the

unusually fast speed of event A in the 2003 Big Bear, CA sequence. We calculate the synthetic velocity spectra of P waves recorded by stations at different azimuths for a small earthquake in a damaged fault zone, and compare them with the velocity spectra of P waves recorded by stations in southern California for a cluster of small earthquakes around event A. The similarity of the azimuth-dependent characteristics of the synthetic and observed spectra suggests the possible existence of a damaged fault zone around event A. We further analyze the properties of fault zone rocks by estimating V_p/V_s ratios of the Big Bear source region and find that the region around event A has an anomalously low V_p/V_s ratio compared to the surrounding region, typical of damaged materials.

2. Fully dynamic models of supershear ruptures in damaged fault zones

The role of damaged fault zones in enhancing supershear rupture behavior was previously studied through numerical simulations by Huang et al. [2014]. Here we systematically investigate the supershear transition as a function of the width and velocity reduction of the fault zone and the background stress. As in our previous study, we simulate inplane dynamic ruptures on faults bisecting damaged fault zones, which are modeled as uniform low-velocity layers. The rupture is nucleated by an over-stressed patch in the middle of the fault and propagates spontaneously shortly after nucleation. The fault is governed by a linear slip-weakening friction law, which allows the fault to slip if the fault stress reaches the static shear strength, and thereafter the fault strength decreases linearly as a function of slip until slip is larger than the critical slip distance D_c . We normalize the fault zone width by the critical nucleation size $L_c = GD_c/\Delta\tau_s$, where G is shear modulus and $\Delta\tau_s$ is the strength drop (the difference between static and dynamic

shear strength). The background stress is quantified non-dimensionally by the relative stress drop, $\Delta\tau_r$, defined as the ratio of stress drop (the difference between initial shear stress and dynamic shear strength) to strength drop. Note that $\Delta\tau_r = 1/(S + 1)$, where S is the strength parameter introduced by Das and Aki [1977]. The dynamic rupture model is solved using a 2D spectral element method, which satisfies numerical convergence as element size decreases [Huang and Ampuero, 2011]. We choose the element size small enough so that the slip-weakening process zone contains significantly more than five grid nodes, and the supershear transition is well resolved.

Fig. 3: Rupture speed normalized by the S wave speed of the host rock as a function of time. For the given stresses ($\Delta\tau_r = 0.25$), rupture speed approaches the Rayleigh wave speed in a homogeneous medium (black), but accelerates to supershear speeds in a fault zone of width $4 L_c$ with different velocity reductions (color). The dashed lines show the P wave speeds in each fault zone.

Fig. 3 shows the evolution of rupture speed in dynamic rupture simulations with various fault zone properties. In all the examples shown, the rupture speed starts in the subshear range, eventually jumps to supershear and then approaches the P-wave speed of the fault zone rock. Huang et al. [2014] found that the supershear transition in damaged fault zones can be triggered by fault zone head waves, which travel faster than the initial rupture front

(Fig. 2b). When the shear stress mediated by head waves overcomes the fault strength, a daughter crack is nucleated ahead of the initial rupture front and rapidly develops into a new leading front that propagates at supershear speed.

The rupture speeds that would be considered as unstable supershear and forbidden in a homogeneous medium made of host rock become stable in the presence of a damaged fault zone. The supershear rupture speeds in fault zones are around the P wave speed of the damaged rock, that is, between $0.9v_s$ and $1.4v_s$ in fault zones of 19-48% velocity reductions for small $\Delta\tau_r$ values (grey triangles in Fig. 1 and dashed lines in Fig. 3), where v_s now denotes the S wave speed of the host rock. Thus, it is possible that observed earthquakes achieved nominally unstable supershear or forbidden speeds (relative to the host rock) due to interactions with their damaged fault zones.

A systematic set of simulations was conducted to establish the robustness of the fault zone supershear mechanism over a wide range of model parameters. Fig. 4 shows how the supershear transition length (the rupture propagation distance at which the supershear transition occurs) depends on background stress and fault zone width, for a fault zone with 30% velocity reduction. Supershear earthquakes are usually thought to occur on fault segments with large $\Delta\tau_r$, i.e., large stress drop and relatively small strength drop [Andrews, 1976; Das and Aki, 1977]. Previous studies in homogeneous media find that 2D in-plane shear cracks under slip-weakening friction can become supershear if $\Delta\tau_r > 0.361$ ($S < 1.77$) [Andrews, 1985]. Such $\Delta\tau_r$ values seem high for mature faults that operate at low background stresses, and localized stress concentrations have been proposed to promote supershear transitions in such cases [Dunham et al., 2003; Liu and Lapusta, 2008]. Because of the additional stress stimulus from head waves, the supershear transition in damaged

190 fault zones can occur at background stresses lower than those required in homogeneous
 191 media ($\Delta\tau_r < 0.361$) for a wide range of velocity reductions (Fig. 3) and fault zone widths
 192 (Fig. 4).

193 **Fig. 4: Contours of supershear transition lengths in damaged fault zones of 30% velocity reduction as**
 194 **a function of $\Delta\tau_r$ and fault zone width.** The transition lengths and fault zone widths are normalized by the
 195 size of the critical nucleation zone, L_c . The gray dashed line on the top left (overlain with the contour of
 196 $100 L_c$) is the boundary above which supershear transition happens further than $100 L_c$. The black dashed
 197 line near the bottom shows the boundary below which supershear transitions are not observed. The crosses
 198 denote the fault zone widths and $\Delta\tau_r$ values for which oscillations of rupture speed occur.

200

201 The supershear transition length increases with the fault zone width when $\Delta\tau_r < 0.361$
 202 (Fig. 4), because head waves need to propagate for a longer time across a wider fault zone
 203 to reach the rupture front. If the fault zone is too wide, the rupture appears to propagate in a

homogeneous medium with the material properties of the fault zone. When $\Delta\tau_r > 0.361$, supershear ruptures can also be induced by a stress peak traveling at S wave speed as in homogeneous media [Andrews, 1976]. Because of the coexistence of supershear transitions caused by the fault zone head wave and the stress peak as in homogeneous media, the supershear transition length has a complicated dependence on the fault zone width in the intermediate range of fault zone widths when $0.361 < \Delta\tau_r < 0.5$. The supershear transition length is independent of the fault zone widths when $\Delta\tau_r \geq 0.5$, because the supershear transition in a homogeneous medium with the material properties of the fault zone is dominant. Given the same $\Delta\tau_r$ values that allow supershear ruptures in both homogeneous media and damaged fault zones, supershear transitions in damaged fault zones occur at much shorter rupture propagation distances, making them more likely to be a dominant mechanism on mature faults.

3. Seismic observations of damaged fault zones in Big Bear

3.1 Kinematic models of small earthquakes in damaged fault zones

Our dynamic rupture simulations suggest that it is possible to estimate the potential of faults to host supershear ruptures based on the properties of their damaged fault zones. The fault zone properties are usually investigated using waveform modeling of fault zone guided waves [Li et al., 1994] and multiple reflections [Li et al., 2007; Yang and Zhu, 2010] recorded by near-fault seismic arrays. However, the required data is available only in some major fault zones. Here, we develop a new method of imaging fault zones based on azimuthal variations of frequency content of seismic waves radiated by small earthquakes in damaged fault zones.

228

229 **Fig. 5: The synthetic P-wave velocity spectra compared with the observed P-wave velocity spectra for**
 230 **a cluster of Mw 2.1-3.1 earthquakes around event A. a,** Model configuration for an earthquake (blue
 231 star) in a damaged fault zone (green) embedded in a host rock (brown). Stations (blue triangles) are located
 232 every 3° in the whole range of azimuths θ from fault normal. Stations at $\theta = 75^\circ$ - 105° and $\theta = 255^\circ$ -
 233 285° are in Zone 1 and stations at $\theta = -30^\circ$ - 30° and $\theta = 150^\circ$ - 210° are in Zone 2. **b,** Synthetic P waves
 234 recorded by stations at $\theta = 0^\circ$ - 180° . **c,** Synthetic 2D P-wave velocity spectra in Zone 1 (red) and Zone 2
 235 (blue), except nodal stations. **d,** Synthetic 3D P-wave velocity spectra in Zone 1 (red) and Zone 2 (blue),
 236 except nodal stations. **e,** The average P-wave velocity spectra observed at stations SLR (red) and RVR
 237 (blue). The locations of the seismic stations are shown in Fig. 6.

238

We demonstrate our concept in a 2D kinematic model with a damaged fault zone (Fig. 5a). The damaged fault zone, embedded in a host rock with $v_p=6$ km/s and $v_s=3.5$ km/s, is 400 m wide and 4 km long. Both P and S wave speeds are 40% lower in the central 2 km of the fault zone and increase linearly to the wave speeds in the host rock at both ends. Without the transition zones, waves would be strongly scattered by the abrupt ends of the fault zone. We set stations over all azimuths at a distance of 15 km from the earthquake, far enough to separate the arrivals of P and S waves. We model the earthquake in the fault zone as a double-couple source with a Gaussian source time function and a central frequency of 10 Hz, near the corner frequencies of Mw 2-3 earthquakes. The recorded P waves show several secondary arrivals caused by waves reflected from the interfaces between the fault zone and the host rock (Fig. 5b). Noticeably the characteristic periods of these secondary P arrivals are shorter at azimuths near fault strike. Given our assumption of a purely elastic and homogeneous host rock, P waves at stations farther away have the same frequency content but weaker amplitudes.

We hereafter denote the zone that spans 15° from fault strike as Zone 1, while the zone that spans 30° from fault normal as Zone 2 (Fig. 5a). We calculate the velocity spectra of P waves by taking a 1 s window starting immediately before the first P arrival. The velocity spectra show a peak between 5 and 10 Hz at all azimuths, associated with the central frequency of the source time function. However, at stations in Zone 1 we see a second spectral peak between 10 and 20 Hz, corresponding to the characteristic frequencies of the secondary P arrivals at azimuths near fault strike. Given the fault zone P wave velocity of 3600 km/s and the fault zone width of 400m, the characteristic frequency of the fault zone P wave is around 10 Hz. The spectral peak between 10 and 20

Hz does not appear at stations in Zone 2 (Fig. 5c). We confirm that these patterns persist after considering the approximate 2D-to-3D correction (a factor of $1/\sqrt{f}$, where f is frequency), the Brune's source model and the attenuation of the medium (Fig. 5d). We choose $Q = 100$ to reproduce the observed high-frequency to low-frequency spectral ratios in Zone 2 (Fig. 5e, see section 3.2). The ratios would be larger for higher Q values. Overall, the 2D-to-3D conversion has insignificant effects on the second spectral peak at stations in Zone 2 and on the difference between spectra of Zone 1 and Zone 2. The azimuth-dependent characteristics of P-wave velocity spectra reflect the amplification effect of damaged fault zones on high-frequency waves and can be used to estimate the extent of damaged fault zones when the regional seismic network has a good azimuthal coverage and includes stations in both Zone 1 and Zone 2.

3.2 P-wave spectra analysis of Big Bear sequence

Although the rupture speed of event A of the 2003 Big Bear sequence is forbidden or unstable supershear in a homogeneous medium, our dynamic rupture models suggest that it may be explained by a stable supershear rupture in a damaged fault zone. Here we seek evidence of this mechanism by applying the P-wave spectra analysis to the source region of the 2003 Big Bear sequence, to investigate whether event A occurred in a damaged fault zone. Prominent damaged fault zones have been found previously around the nearby Landers fault [Li et al., 1994; Li et al., 2007]. We use the data recorded by the Southern California Seismic Network for a cluster of Mw 2.1-3.1 earthquakes in the 2003 Big Bear sequence (Cluster A in Fig. 8a). The 14 earthquakes in the cluster are 4-6.5 km deep and within a distance of 300 m from event A [Hauksson et al., 2012]. These earthquakes close to event A (and to the main fault) are likely to have similar focal mechanisms. In fact,

Tan and Helmberger [2010] show that most of the events close to the main fault have similar strike-slip focal mechanisms. Differences of spectra due to slight variability of focal mechanisms are mitigated by averaging the spectra. The variations of average P-wave spectra at stations in Zone 1 and Zone 2 are mainly due to the effects of wave propagation in fault zones rather than source mechanisms.

We select 17 high-quality broadband stations that are 20-80 km from the cluster (Fig. 6). The earthquakes are considered as double-couple point sources as their rupture lengths are much shorter than the source-receiver distances. After removing the instrument response, we calculate the velocity spectra of P waves by taking a 1 s window starting immediately before the first P arrival on the vertical component and average the P-wave spectra of the 14 earthquakes at each station. We find that all the three stations in Zone 1 (SLR, DGR and DNR) show second peaks at 10-20 Hz in the average P-wave velocity spectra, which are similar to the synthetic spectra and not observed at the other stations (Fig. 5d and Fig. 7). Wave propagation along the paths outside fault zones are not likely to have a significant effect on the second spectral peaks because the three stations span an azimuth of $\sim 25^\circ$. Though hard rock sites can have site-specific ground motion amplifications at 10-20 Hz, there are no obvious differences between the rock sites for stations in Zone 1 and Zone 2 [Tan and Helmberger, 2007]. The similarity between the observed and synthetic P-wave spectra supports the possible existence of a damaged fault zone around event A.

305

306

307

308

Fig. 6: Locations of seismic stations used in the P-wave spectra calculation and V_p/V_s ratio estimation. The blue circle denotes the location of the epicenter of event A.

Fig. 7: P-wave velocity spectra for the cluster around event A recorded at stations in Zone 1 (left) and stations in Zone 2 (right). Blue solid lines show the P-wave velocity spectra for each earthquake in the cluster. Red solid lines show the average P-wave velocity spectra. Station names are indicated above each figure.

3.3 V_p/V_s ratio estimation of Big Bear fault zone

In addition, we also find that the fault zone materials around event A feature an anomalously low V_p/V_s ratio compared to the surrounding region. We determine V_p/V_s ratios using data from the cluster of Mw 2.1-3.1 earthquakes around event A and another reference cluster on the south side of the fault for comparison (Cluster B in Fig. 8a). Both clusters have 14 earthquakes with similar depth ranges and have a size of several hundred meters. Given the cluster size is much smaller than the cluster-station distance, the V_p/V_s

ratio inverted from each pair of events in the cluster can be expressed as follows [Lin and Shearer, 2007]:

$$\frac{V_p}{V_s} = \frac{\Delta l^i / V_s}{\Delta l^i / V_p} = \frac{\Delta t_s^i}{\Delta t_p^i} \quad (1)$$

where Δl^i is the difference in the ray-path distances between each pair of events given a certain station i , and Δt_p^i and Δt_s^i are differential P- and S-wave travel times to station i , respectively.

Fig. 8: Inversions of V_p/V_s ratios for the Big Bear fault zone. (a) Two clusters of M_w 2.1-3.1 earthquakes are shown by circles. The moment magnitude, M_w , rupture length, fl , and rupture speed, V_r , of event A are also shown at the top. The gray dashed arrow displays the approximate rupture propagation direction of the mainshock. (b) The circles show the demeaned differential P- and S-wave travel times of event pairs. The slope of each solid line provides the estimated V_p/V_s ratio, while the black dashed lines have a slope of $\sqrt{3}$.

We filter data at each station using a four-pole Butterworth filter with a frequency band of 0.5-5 Hz, lower than the corner frequency of these earthquakes, and calculate differential P- and S-wave arrival times at each station using waveform cross-correlations. We only use event pairs with a cross-correlation coefficient larger than 0.85 for at least five stations. The measured differential arrival times are demeaned for each event pair to eliminate the differences of origin times of the earthquakes, which yields demeaned differential P- and S-wave travel times. Their ratio is V_p/V_s , as in equation (1), provided V_p/V_s is uniform in the vicinity of the earthquake pair. We can estimate the uncertainty in the calculation of differential travel times using the following equation [Knapp and Carter, 1976]:

$$\sigma = \frac{1}{2\pi} \sqrt{\frac{3(1-c^2)}{c^2 T (f_2^3 - f_1^3)}} \quad (2)$$

where c is the cross-correlation coefficient, $[f_1 f_2]$ is the frequency band and T is the time window. Using $c = 0.85$ and $T = 1$, we find $\sigma = 0.015$ s, which is in the same order as the dispersion of our measurements (Fig. 8b). We then apply a robust linear regression method [Huber, 1973] to the differential P- and S-wave travel times to obtain the V_p/V_s ratios. The standard errors are estimated by randomly resampling the differential times 10000 times. The reference cluster gives a V_p/V_s ratio of 1.689 ± 0.035 , near the median value of V_p/V_s ratios of Southern California [Lin and Shearer, 2009], while the V_p/V_s ratio is 1.571 ± 0.043 for the fault zone around event A, indicating rocks there are likely to be extensively damaged (Fig. 8b). Our results suggest fault zone materials vary along strike even within a distance of ~ 1 km, which may lead to variations of rupture speeds of small

earthquakes. In particular, events in the damaged fault zone can have faster rupture speeds compared to events outside the damaged fault zone (Fig. 2b).

4. Discussion and conclusions

Here we present both dynamic rupture simulations and observational evidence of supershear ruptures induced by head waves in damaged fault zones. We find supershear ruptures can occur at background stresses lower than those required in homogeneous media and propagate at speeds considered as unstable supershear or forbidden in homogeneous media. Our simulations are based on a fault zone model with sharp and planar boundaries, consistent with the abrupt changes of wave speeds in the borehole data of the San Andreas fault zone [Ellsworth and Malin, 2011] and the observation of multiple reflections in high-frequency body waves in the Landers and San Jacinto fault zones [Li et al., 2007; Yang and Zhu, 2010]. While we only considered elastic fault zones here, wave attenuation and off-fault plasticity may postpone the transition of supershear ruptures in damaged fault zones. Earthquake ruptures may generate new damage (coseismic changes of fault zone material properties) that can modulate rupture behavior [Xu et al., 2015], but the fundamental mechanism of supershear transition is likely similar to that described in Section 2. The existence of damaged zones off the fault can also induce supershear ruptures at smaller distances and lower stresses [Ma and Elbanna, 2015].

Though our observation in Big Bear supports the correlation between a damaged fault zone and a small supershear earthquake, the dynamic rupture simulations suggest that such mechanism of supershear transition is also applicable to large earthquakes. Given

the same fault zone and fault stresses, supershear transitions in small and large earthquakes may occur at different rupture propagation distances (Fig. 4), hence making the likelihood of observing supershear earthquakes vary with earthquake magnitude, fault zone properties and fault stresses. Damaged fault zones have also been found in several regions where large supershear earthquakes have ruptured. Adjacent to the supershear rupture region of the 1999 M_w 7.1 Düzce earthquake, trapped wave analysis revealed a 100-m-wide damaged fault zone of 50% velocity reductions [Ben-Zion et al., 2003]. The 1906 M_w 7.9 San Francisco earthquake may have ruptured to the north at an unstable supershear rupture speed of ~ 3.9 km/s ($\sim 1.1 v_s$) along the part of San Andreas fault where the existence of a damaged fault zone was inferred from geodetic measurements [Song et al., 2008; Jolivet et al., 2009]. However, there is still insufficient characterization of fault zone properties in regions where large supershear earthquakes have been observed, and further studies resolving along-strike variations of fault zone properties and rupture speed are warranted.

Supershear ruptures can enhance strong ground motions and induce the extensive rock pulverization observed in major fault zones [Doan and Gary, 2009], hence contributing to the damage of fault zones through seismic cycles. The enhancement of high-frequency radiation in the azimuths near fault strike is a general feature of earthquakes that occur in damaged fault zones, regardless of their rupture speed. Characterization of the extent of damaged fault zones can provide a useful tool for estimating the potential of seismic hazards on active faults. It has been found that several past earthquakes tend to experience faster rupture speeds, occasionally supershear, along their rupture segments located in the most mature parts of the causative faults [Perrin et al., Along-strike change

in fault structural maturity controls location of largest earthquake slip and fast rupture,
manuscript submitted to *Earth Planet. Sci. Lett.*, 2015], which may be a manifestation of
the fault zone supershear mechanism studied here.

References

1. Archuleta, R. J. A faulting model for the 1979 Imperial Valley earthquake. *J. Geophys. Res.* **89**, 4559-4585 (1984).
2. Spudich, P. & Cranswick, E. Direct observation of rupture propagation during the 1979 Imperial Valley earthquake using a short baseline acceleraometer array. *Bull. Seismol. Soc. Am.* **74**, 2083-2114 (1984).
3. Bouchon, M. *et al.* How fast is rupture during an earthquake? New insights from the 1999 Turkey earthquake. *Geophys. Res. Lett.* **28**, 2723-2726 (2001).
4. Bouchon, M. *et al.* Space and time evolution of rupture and faulting during the 1999 Izmit (Turkey) Earthquake. *Bull. Seismol. Soc. Am.* **92**, 256-266 (2002).
5. Sekiguchi, H. & Iwata, T. Rupture process of the 1999 Kocaeli, Turkey, earthquake estimated from strong-motion waveforms. *Bull. Seismol. Soc. Am.* **92**, 300-311 (2002).
6. Birgören, G., Sekiguchi, H. & Irikura, K. Rupture model of the 1999 Düzce, Turkey, earthquake deduced from high and low frequency strong motion data. *Geophys. Res. Lett.* **31**, L05610 (2004).
7. Bouin, M. P., Bouchon, M., Karabulut, H. & Aktar, M. Rupture process of the 1999 November 12 Düzce (Turkey) earthquake deduced from strong motion and Global Positioning System measurements. *Geophys. J. Int.* **159**, 207-211 (2004).

424 8. Konca, A. O., Lepreine, S., Avouac, J.-P. & Helmberger, D. V. Rupture process of the
425 1999 M_w 7.1 Düzce earthquake from Joint analysis of SPOT, GPS, INSAR, Strong-
426 motion, and teleseismic data: a supershear rupture with variable rupture velocity. *Bull.*
427 *Seismol. Soc. Am.* **100**, 267-288 (2010).

428 9. Antolik, M., Abercrombie, R. E. & Ekström G. The 14 November 2001 Kokoxili
429 (Kunlunshan), Tibet, Earthquake: Rupture transfer through a large extensional step-over.
430 *Bull. Seismol. Soc. Am.* **94**, 1173-1194 (2004).

431 10. Bouchon, M. & Vallée, M. Observation of long supershear rupture during the
432 magnitude 8.1 Kunlunshan earthquake. *Science* **301**, 824-826 (2003).

433 11. Robinson, D. P., Brough, C. & Das, S. The M_w 7.8, 2001 Kunlunshan earthquake:
434 extreme rupture speed variability and effect of fault geometry. *J. Geophys. Res.*
435 **111**, B08303 (2006).

436 12. Vallée, M., Landès, M., Shapiro, N. M. & Klinger, Y. The 14 November 2001
437 Kokoxili (Tibet) earthquake: High-frequency seismic radiation originating from the
438 transitions between sub-Rayleigh and supershear rupture velocity regimes. *J. Geophys.*
439 *Res.* **113**, B07305 (2008).

440 13. Walker, K. T. & Shearer, P. M. Illuminating the near-sonic rupture velocities of the
441 intracontinental Kokoxili M_w 7.8 and Denali fault M_w 7.9 strike-slip earthquakes with
442 global P wave back projection imaging. *J. Geophys. Res.* **114**, B02304 (2009).

443 14. Dunham, E. M. & Archuleta, R. J. Evidence for a supershear transient during the
444 2002 Denali fault earthquake. *Bull. Seismol. Soc. Am.* **94**, S256-S268 (2004).

445 15. Ellsworth, W. L. et al. Near-field ground motion of the 2002 Denali fault, Alaska,
446 earthquake recorded at Pump Station 10. *Earthquake Spectra* **20**, 597-615 (2004).

447 16. Asano, K., Iwata, T. & Irikura, K. Estimation of source rupture process and strong
448 ground motion simulation of the 2002 Denali, Alaska, earthquake. *Bull. Seismol. Soc.
449 Am.* **95**, 1701-1715 (2005).

450 17. Wang, D. & Mori, J. The 2010 Qinghai, China, earthquake: a moderate earthquake
451 with supershear rupture. *Bull. Seismol. Soc. Am.* **102**, 301-308 (2012).

452 18. Yue, H. et al. Supershear rupture of the 5 January 2013 Craig, Alaska (M_w 7.5)
453 earthquake. *J. Geophys. Res.* **118**, 5903-5919 (2013).

454 19. Zhan, Z., Helmberger, D. V., Kanamori, H. & Shearer, P. Supershear rupture in a M_w
455 6.7 aftershock of the 2013 Sea of Okhotsk earthquake. *Science* **345**, 204-207 (2014).

456 20. Freund, L. B. The mechanics of dynamic shear crack propagation. *J. Geophys. Res.*
457 **85**, 2199-2209 (1979).

458 21. Burridge, R., Conn, G. & Freund, L. B. The stability of a rapid mode II shear crack
459 with finite cohesive traction. *J. Geophys. Res.* **85**, 2210-2222 (1979).

460 22. Broberg, K. B. Cracks and fracture. Academic Press (1999).

461 23. Rosakis, A. J., Samudrala, O. & Coker, C. Cracks faster than the shear wave speed.
462 *Science* **284**, 1337-1340 (1999).

463 24. Bizzarri, A. & Das, S. Mechanics of 3-D shear cracks between Rayleigh and shear
464 wave rupture speeds. *Earth Planet. Sci. Lett.* **357**, 397-404 (2012).

465 25. Luo, Y. et al. Source mechanism and rupture directivity of the 18 May 2009 M_w 4.6
466 Inglewood, California, earthquake. *Bull. Seismol. Soc. Am.* **100**, 3269-3277 (2010).

467 26. Tan, Y. & Helmberger, D. Rupture directivity characteristics of the 2003 Big Bear
468 sequence. *Bull. Seismol. Soc. Am.* **100**, 1089-1106 (2010).

- 469 27. Harris, R. A. & Day, S. M. Effects of a low-velocity zone on a dynamic rupture. *Bull.*
470 *Seismol. Soc. Am.* **87**, 1267-1280 (1997).
- 471 28. Huang, Y., Ampuero, J.-P. & Helmberger, D. V. Earthquake ruptures modulated by
472 waves in damaged fault zones. *J. Geophys. Res.*, **119**, 3133–3154,
473 doi:10.1002/2013JB010724 (2014).
- 474 29. Andrews, D. J. Rupture velocity of plane strain shear cracks. *J. Geophys. Res.* **81**,
475 5679-5687 (1976).
- 476 30. Das, S. & Aki, K. A numerical study of two-dimensional spontaneous rupture
477 propagation. *Geophys. J. Roy. Astron. Soc.* **50**, 643-668 (1977).
- 478 31. Andrews, D. J. Dynamic plane-strain shear rupture with a slip-weakening friction law
479 calculated by a boundary integral method. *Bull. Seismol. Soc. Am.* **75**, 1-21 (1985).
- 480 32. Dunham, E. M., Favreau, P. & Carlson, J. M. A supershear transition mechanism for
481 cracks. *Science* **299**, 1557-1559 (2003).
- 482 33. Liu, Y. & Lapusta, N. Transition of model II cracks from sub-Rayleigh to intersonic
483 speeds in the presence of favorable heterogeneity. *J. Mech. Phys. Solids* **56**, 25-50 (2008).
- 484 34. Huang, Y. & Ampuero, J.-P. Pulse-like ruptures induced by low-velocity fault zones.
485 *J. Geophys. Res.* **116**, B12307 (2011).
- 486 35. Li, Y.-G., Aki, K., Adams, D., Hasemi, A. & Lee, W. H. K. Seismic guided waves
487 trapped in the fault zone of the Landers, California, earthquake of 1992. *J. Geophys. Res.*
488 **99**, 11705–11722 (1994).
- 489 36. Li, H., Zhu, L. & Yang, H. High-resolution structures of the Landers fault zone
490 inferred from aftershock waveform data. *Geophys. J. Int.* **171**, 867-881 (2007).

491 37. Yang, H. & Zhu, L. Shallow low-velocity zone of the San Jacinto fault from local
492 earthquake waveform modeling. *Geophys. J. Int.* **183**, 421-432 (2010).

493 38. Hauksson, E., Yang, W. & Shearer, P. M. Waveform relocated earthquake catalog for
494 Southern California (1981 to 2011). *Bull. Seismol. Soc. Am.* **102**, 2239-2244 (2012).

495 39. Tan, Y. & Helmberger, D. A new method for determining small earthquake source
496 parameters using short-period P waves. *Bull. Seismol. Soc. Am.* **97**, 1176-1195 (2007).

497 40. Lin, G. & Shearer, P. Estimating local V_p/V_s ratios within similar earthquake
498 clusters. *Bull. Seismol. Soc. Am.* **97**, 379-388 (2007).

499 41. Knapp, C. & Carter, G.C. The generalized correlation method for estimation of time
500 delay. *IEEE T. Acoust. Speech*, **24**, 320-327 (1976).

501 42. Huber, P. J. Robust regression: asymptotics, conjectures and Monte Carlo. *Ann.*
502 *Statist.* **1**, 799-821 (1973).

503 43. Lin, G. & Shearer, P. Evidence for water-filled cracks in earthquake source regions.
504 *Geophys. Res. Lett.* **36**, L17315 (2009).

505 44. Ellsworth, W. L. & Malin, P. E. Deep rock damage in the San Andreas Fault revealed
506 by P- and S- type fault-zone-guided-waves, *Geol. Soc. London Spec. Publ.* **359**, 39-53
507 (2011).

508 45. Xu, S., Ben-Zion, Y., Ampuero, J.-P. & Lyakhovsky, V. Dynamic ruptures on a
509 frictional interface with off-fault brittle damage: Feedback mechanisms and effects on
510 slip and near-fault motion, *Pure Appl. Geophys.* **172**, 1243-1267 (2015).

511 46. Ma, X. & Elbanna, A. E. Effect of off-fault low-velocity elastic inclusions on
512 supershear rupture dynamics, *Geophys. J. Int.* **203**, 664-677 (2015).

47. Ben-Zion, Y. et al. A shallow fault-zone structure illuminated by trapped waves in the Karadere-Düzce branch of the North Anatolian Fault, western Turkey. *Geophys. J. Int.* **152**, 699-717 (2003).
48. Song, S. G., Beroza, G. C. & Segall, P. A unified source model for the 1906 San Francisco earthquake. *Bull. Seismol. Soc. Am.* **98**, 823-831 (2008).
49. Jolivet, R., Bürgmann, R. & Houlié, N. Geodetic exploration of the elastic properties across and within the northern San Andreas Fault zone. *Earth Planet. Sci. Lett.* **288**, 126-131 (2009).
50. Doan, M.-L. & Gary, G. Rock pulverization at high strain rate near the San Andreas fault. *Nature Geosci.* **2**, 709-712 (2009).

Acknowledgements

This work was supported by the NSF (grants EAR-0944288 and EAR-1151926) and SCEC (funded by NSF EAR-0106924 and USGS 02HQAG0008 cooperative agreements). The seismic data is available from Southern California Earthquake Data Center (SCEDC) at www.data.scec.org/. We thank Roland Bürgmann, Greg Beroza, Hiroo Kanamori and Luis Rivera for discussions and comments.

Figure Legends

Fig. 1: Rupture speeds observed in large earthquakes (circles and diamonds), laboratory experiments (squares) and numerical models (triangles), normalized by S wave speeds. The circles denote the estimation of the average supershear speed for a portion of the fault and diamonds with bars denote the estimations of the range of

supershear speeds with mean values, from different observations of supershear earthquakes¹⁻¹⁸. The red strip indicates the forbidden speed range and the grey patch indicates the unstable supershear range (SS) in homogeneous media. The two open squares display the limiting speeds of supershear earthquakes in laboratory experiments. The open triangle and grey triangles show the asymptotic speeds of supershear ruptures in homogeneous media and in damaged fault zones, respectively.

Fig. 2: Rupture speed variations along Big Bear fault and the correlation with fault

zone structure. a, Rupture speeds of several of Mw 3.3-4.3 aftershocks in the 2003 Big Bear sequence are shown in colors. The star indicates the mainshock and colored circles are aftershocks for which rupture speeds have been estimated. Arrows indicate the rupture propagation directions, and their lengths are proportional to the rupture lengths. Grey circles are epicenters of relocated seismicity. **b,** Earthquakes propagating through the fault zone can have faster speeds than earthquakes outside the fault zone if the supershear transition is induced by the dynamic stress transfer carried by head waves.

Fig. 3: Rupture speed normalized by the S wave speed of the host rock as a function

of time. For the given stresses ($\Delta\tau_r = 0.25$), rupture speed approaches the Rayleigh wave speed in a homogeneous medium (black), but accelerates to supershear speeds in a fault zone of width $4 L_c$ with different velocity reductions (color). The dashed lines show the P wave speeds in each fault zone.

Fig. 4: Contours of supershear transition lengths in damaged fault zones of 30%

velocity reduction as a function of $\Delta\tau_r$ and fault zone width. The transition lengths and fault zone widths are normalized by the size of the critical nucleation zone, L_c . The gray dashed line on the top left (overlain with the contour of $100 L_c$) is the boundary

above which supershear transition happens further than $100 L_c$. The black dashed line near the bottom shows the boundary below which supershear transitions are not observed. The crosses denote the fault zone widths and $\Delta\tau_r$ values for which oscillations of rupture speeds occur.

Fig. 5: The synthetic P-wave velocity spectra compared with the observed P-wave velocity spectra for a cluster of Mw 2.1-3.1 earthquakes around event A. a, Model configuration for an earthquake (blue star) in a damaged fault zone (green) embedded in a host rock (brown). Stations (blue triangles) are located every 3° in the whole range of azimuths θ from fault normal. Stations at $\theta = 75^\circ$ - 105° and $\theta = 255^\circ$ - 285° are in Zone 1 and stations at $\theta = -30^\circ$ - 30° and $\theta = 150^\circ$ - 210° are in Zone 2. b, Synthetic P waves recorded by stations at $\theta = 0^\circ$ - 180° . c, Synthetic 2D P-wave velocity spectra in Zone 1 (red) and Zone 2 (blue), except nodal stations. d, Synthetic 3D P-wave velocity spectra in Zone 1 (red) and Zone 2 (blue), except nodal stations. e, The average P-wave velocity spectra observed at stations SLR (red) and RVR (blue). The locations of the seismic stations are shown in Fig. 6.

Fig. 6: Locations of seismic stations used in the P-wave spectra calculation and V_p/V_s ratio estimation. The blue circle denotes the location of the epicenter of event A.

Fig. 7: P-wave velocity spectra for the cluster around event A recorded at stations in Zone 1 (left) and stations in Zone 2 (right). Blue solid lines show the P-wave velocity spectra for each earthquake in the cluster. Red solid lines show the average P-wave velocity spectra. Station names are indicated above each figure.

Fig. 8: Inversions of V_p/V_s ratios for the Big Bear fault zone. (a) Two clusters of Mw 2.1-3.1 earthquakes are shown by circles. The moment magnitude, M_w , rupture length, fl ,

582 and rupture speed, V_r , of event A are also shown at the top. The gray dashed arrow
583 displays the approximate rupture propagation direction of the mainshock. (b) The circles
584 show the differential P- and S-wave travel times of event pairs. The slope of each solid
585 line provides the estimated V_p/V_s ratio, while the black dashed lines have a slope of $\sqrt{3}$.