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Earthquake source properties from pseudotachylite

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Abstract

4 Earthquake-radiated motions contain information that can be interpreted as source displacement
5 and therefore related to stress drop. Except in a few notable cases, these displacements cannot be
6 easily related to the absolute stress level, the fault strength, or attributed to a particular physical
7 mechanism. In contrast paleo-earthquakes recorded by exhumed pseudotachylite have a known
8 dynamic mechanism whose properties constrain the co-seismic fault strength. Pseudotachylite
9 can be used to directly address a discrepancy between seismologically-measured stress drops,
10 which are typically a few MPa, and much larger dynamic stress drops expected from thermal
11 weakening during slip at seismic speeds in crystalline rock [Sibson, 1973; McKenzie and Brune,
12 1969; Lachenbruch, 1980; Mase and Smith, 1986; Rice, 2006], and as have been observed in
13 laboratory experiments at high slip rates [Di Toro et al., 2006a]. This note places
14 pseudotachylite-derived estimates of fault strength and inferred crustal stress within the context
15 and bounds of naturally observed earthquake source parameters: apparent stress, stress drop, and
16 overshoot, including consideration of fault surface roughness, off-fault damage, fracture energy,
17 and the 'strength excess'. The analysis, which assumes stress drop is related to corner frequency
18 by the Madariaga [1976] source model, is restricted to earthquakes of the Gole Larghe fault zone
19 in the Italian Alps where the dynamic shear strength is well-constrained by field and laboratory
20 measurements. We find that radiated energy is similar to or exceeds the shear-generated heat and
21 that the maximum strength excess is ~ 16 MPa. These events have inferred earthquake source
22 parameters that are rare, for instance a few percent of the global earthquake population has stress
23 drops as large, unless: fracture energy is routinely greater than in existing models,
24 pseudotachylite is not representative of the shear strength during the earthquake that generated it,
25 or unless the strength excess is larger than we have allowed.

Introduction

26 Within the earthquake source region a large number of inelastic processes are thought to
27 operate: frictional sliding, rock fracture, dilatancy, melting, devolatilization, thermal expansion
28 of pore fluid, hydrofracture, and creation of new fracture surface energy are among many known
29 and proposed processes [Andrews, 1976; Scholz, 2002; Rice, 2006]. The processes that actually
30 occur depend on mineralogy, ambient temperature and stress conditions, total slip, the degree of
31 shear localization, the amount of shear dilatancy, and fault zone hydraulic properties. Outside the
32 source, the surrounding rock is assumed predominantly elastic and the motions radiated from the
33 source as elastodynamic waves can be related to the spatial time history of displacement within
34 the source. Accounting for attenuation, scattering, and other path effects, information
35 propagating from the source is interpretable at the surface in terms of, for example, source stress
36 drop, moment, radiated energy, and displacement or velocity spectrum, but only on rare
37 occasions to the absolute level of stress [e.g., Spudich, 1992]. For earthquakes that have source
38 mechanisms that are predominately double couple, to date there is little observational or
39 theoretical research that ties surface recorded motions to a particular physical mechanism within
40 the source. So, with the exception of a very few notable claims [e.g., Kanamori *et al.*, 1998],
41 what source processes actually occur for any particular earthquake is anyone's guess.

Field observations and melt shear strength.

42 A well-understood exception are the ancient earthquakes recorded in exhumed
43 pseudotachylites [Sibson, 1975]. Pseudotachylite is thought by most to be the definitive record of
44 an earthquake where dynamic strength was controlled by shear melting [Jeffreys, 1942;
45 McKenzie and Brune, 1972; Sibson, 1975], though there are alternative interpretations [e.g., Pec
46 *et al.*, 2012 and references therein]. In the present study we assume that natural pseudotachylites
47 are generated by coseismic shear heating and take advantage of field and laboratory constraints
48 on the co-seismic properties of the shear zone. Melt layers are viscous and therefore have
49 strengths that are quite strongly slip rate- and thickness-dependent. In addition the viscosity can

50 depend on the characteristics of the flow regime and melt composition [*Spray*, 1993; *Lavallee et*
51 *al.*, 2015]. The field measurements avoid these complexities and produce empirical constraints
52 on the dynamic shear strength during the event [*Sibson*, 1975]. Specifically, field-measured
53 values of the thickness of a pseudotachylite layer, w , are used to estimate the heat necessary to
54 melt a particular volume of rock of a particular composition using the protolith heat capacity.
55 *Sibson* assumed all the shear generated heat remains in the slipping zone and causes melting
56 immediately at the melting temperature T_m of the constituent minerals. Somewhat more recently
57 *Wenk et al.*, [2000] and *Di Toro et al.* [2005] repeated the same type of analysis while also
58 allowing for some of the shear heat to be absorbed in the slipping zone as latent heat of fusion.
59 Accordingly the heat necessary to convert a thickness of rock entirely to melt is

$$60 \quad Q = A\rho w \left[(T_m - T_0)c_p + H \right], \quad (1a)$$

61 where c_p is the heat capacity (energy/mass K), H is the heat of fusion (in energy/mass), A is fault
62 area, ρ is density (mass/volume) and T_0 is the initial slipping zone temperature. The two terms
63 on the righthand side of (1a) are from left to right, the change in thermal energy within the
64 slipping zone and the energy necessary to drive the endothermic melting reaction, the latent heat
65 stored within the melt. This assumes that significant heat does not diffuse away from the fault
66 coseismically, which is reasonable given the low thermal diffusivity of rocks ($\kappa \approx 10^{-6} \text{ m}^2 \text{ s}^{-1}$)
67 and the few second duration, Δt , of earthquake slip [*Lachenbruch*, 1980], which results in a heat
68 penetration distance, $z \approx \sqrt{\kappa \Delta t} < 2\text{-}3 \text{ mm}$. An additional requirement of (1a) is that the slipping
69 zone temperature does not exceed the melting temperature (no superheating) which is expected if
70 the phase change buffers the temperature increase. The displacement-averaged shear strength is

$$71 \quad \tau_m = \frac{Q}{A\Delta d}, \quad (1b)$$

72 where Δd is fault slip as measured in the field using offset markers across the fault [*Sibson*,
73 1975]. Combining (1a) and (1b) the displacement-averaged shear strength during seismic slip
74 that produces a shear melt is

$$75 \quad \hat{\tau}_m = \frac{\rho w}{\Delta d} \left[(T_m - T_0)c_p + H \right]. \quad (1c)$$

76 Note that the heat of fusion is on the order of 10^5 J/kg while the heat capacity is of order 10^3
77 J/kgK for granitic compositions. So long as the temperature difference $T_m - T_0$ is 1000 K or more
78 the change in thermal energy greatly exceeds the heat of fusion and dominates the sum (1c). This
79 is the case for both the natural [Di Toro *et al.*, 2005] and laboratory [Di Toro *et al.*, 2006a]
80 settings of tonalitic pseudotachylite generation that we consider in this study.

81 Thickness displacement ratios, $w/\Delta\delta$ measured by Di Toro *et al.* [2005; 2006a] for
82 pseudotachylite in tonalite within the Gole Larghe fault zone in the southern European Alps
83 exhumed from hypocentral depths of 9 to 11 km and T_0 of 250°C are between 0.01 and 0.004.
84 The associated calculated shear strengths are between 15 and 48 MPa, as depicted in **Figure 1**.
85 This technique to estimate melt shear strength, equation (1c), was confirmed for normal stresses
86 > 20 MPa in experiments simulating coseismic slip on gabbro [Niemeijer *et al.*, 2011]. In the
87 field the approach also requires some independent measure of the ambient temperature prior to
88 the earthquake. Hypocentral temperature ($T_0 \approx 250$ °C) of the Gole Larghe was estimated from
89 deformation microstructures of quartz in cataclasites associated with the pseudotachylites, and by
90 the mineral assemblage of coeval metamorphic alteration by Di Toro and Pennacchioni [2004].

Lab observations of melt shear strength.

91 Meanwhile, advances in experimental design and technique [Tsutsumi and Shimamoto, 1997;
92 Hirose and Shimamoto, 2003, 2005] and related theoretical developments [Nielsen *et al.*, 2008,
93 2010; Di Toro *et al.*, 2006b] allow determination of the shear strength and constitutive response
94 of friction melts of identical composition to the Gole Larghe fault zone field exposures at a few
95 to a few 10's of MPa normal stress [Di Toro *et al.*, 2006a]. Laboratory shear melting
96 experiments by Di Toro *et al.* [2006a] were conducted at normal stresses between 5 and 20 MPa
97 at a sliding velocity of 1.3 m/s for 4 to 8 s on the source tonalite collected from the Adamello
98 batholith in the southern Italian Alps from which the natural pseudotachylites were exhumed. In
99 our study the reported steady-state shear strengths from Di Toro *et al.* [2006a] are assumed to be
100 analogous to their natural equivalents. The melt steady-state shear strength resembles the

101 unmelted strength of granitic faults [Byerlee, 1978] at the lowest normal stresses (**Figure 1**) but
102 is considerably weaker at 10 to 20 MPa normal stress, the highest normal stresses tested. For
103 extrapolation to the conditions of the natural pseudotachylites, the ‘pressure dependence’ of fault
104 strength $d\tau/d\sigma_e$ is the necessary metric; for these faults shear strength increases very weakly with
105 normal stress (0.05 MPa per MPa) and using this extrapolation from *Di Toro et al.* [2006a], the
106 implied natural strength at 9 to 11 km is less than 20 MPa (**Figure 1**).

107 In this study we examine the energy budget of earthquakes that generate shear melts of
108 tonalitic composition. Knowing both the shear generated heat from field observations and the
109 shear strength from laboratory measurements puts constraints on energy partitioning that are
110 lacking for all other earthquakes. Our approach is to use the laboratory and field measurements
111 of co-seismic fault strength along with the known static strength of the granitic host rock as the
112 independent variables and determine the possible range of source parameters for the paleo-
113 earthquakes that generated these melts. Throughout the paper we refer to these prehistoric
114 seismic events as earthquakes for simplicity. Particular goals are to establish whether these
115 events could be consistent with typical earthquake source properties and what seismically
116 observed properties may be diagnostic of melting. We find that earthquakes generating these
117 pseudotachylites have atypical source properties that arise from the very high static frictional
118 strength of granitic rock and the very low strength of shear melts. As in this particular example,
119 and likely in other rocks that have high frictional strengths at low sliding speed and for which
120 shear heating produces a weak melt, the result is a large stress drop and relatively high radiated
121 energy.

Energy during dynamic slip

122 Ignoring gravitational and rotational terms, the total energy of an earthquake E_T can be
123 partitioned between heat Q_{all} and radiated energy E_S ,

$$124 \quad E_T = Q_{all} + E_S. \quad (2)$$

125 Here, as follows from the analysis of *Savage and Wood* [1971] [e.g., *McGarr*, 1999; *Beeler*
126 2006], we have included in Q_{all} both the shear generated heat that is available to be conducted
127 away from the fault, and also latent heats that are absorbed during shearing: for example the heat
128 of fusion (as in equation (1) [*Di Toro et al.*, 2005]), heat of reaction during other phase changes
129 [e.g., devolatilization *Brantut et al.*, 2011], and the creation of surface energy that results from
130 wear and comminution [*Lachenbruch and Sass*, 1980]. The average shear stress on the fault is
131 related to the total work $\bar{\tau} = E_T / A\Delta\delta$. Following the definition of heat above, equation (1b),
132 define a shear ‘strength’ $\hat{t} = Q_{all} / AD\delta$; \hat{t} is the stress measure of energy dissipated and stored
133 in the source, spatially- and slip-averaged over the entire source region [*McGarr*, 1999; *Beeler*,
134 2006]. It is a representative sliding strength of the fault, associated with energy distributed within
135 the source, including heat, latent heat associated with chemical reactions and with the creation of
136 surface energy. Using the standard definition of apparent stress as the stress measure of radiated
137 energy $\tau_a = E_s / A\Delta\delta$, the balance (2) can be rewritten in stress units as

$$138 \quad \bar{\tau} = \hat{t} + t_a, \quad (3)$$

139 [*Savage and Wood*, 1971; *McGarr*, 1994]. The energy budget can be graphically expressed using
140 a stress versus displacement diagram (**Figure 2**) [*McGarr*, 1994]. The figure presents the
141 definitions of stress quantities used throughout this paper. In particular the average stress is the
142 difference between the static stress levels before and after the earthquake, $\bar{\tau} = (t_0 + t_1) / 2$, where
143 t_0 is the initial stress on the fault prior to the earthquake and t_1 is the stress after seismic slip.

144 Equating the shear strength that produces melt, equation (1b), to this stress measure of all the
145 energy that is not radiated, $\hat{t} = \hat{t}_m$, is the first crucial assumption in our analysis. Making this
146 assumption presumes, for example, that any off-fault damage makes a negligible contribution to
147 the energy budget. This is an assumption that is difficult to verify [*Pittarello et al.*, 2008] and not
148 without associated controversy [e.g., *Wilson et al.*, 2005; *Chester et al.*, 2005; *Ma et al.*, 2006].
149 Some of the limitations and implications if this assumption is relaxed are detailed in the
150 Discussion section below. Recent field studies of pseudotachylite, e.g., *Di Toro et al.* [2006a],

151 have equated the fault shear strength $\hat{\tau}$ inferred from thickness-displacement ratios (1c) (**Figure**
 152 **1**) with the average crustal shear stress $\bar{\tau}$. The average shear strength and the static shear stress
 153 are approximately equivalent under special circumstances, as noted by *McGarr* [1994; 1999].
 154 This analysis to recover shear stress has been repeated elsewhere [e.g., *Barker*, 2005; *Ujiiie et al.*,
 155 2007; *Andersen et al.*, 2008; *Billi and Di Toro*, 2008]. That relationship is valid only if $\hat{\tau} = \hat{\tau}_m$,
 156 as we have assumed, and if the apparent stress, τ_a , is relatively small. For shear melting there are
 157 no published proportions of radiated energy and heat from laboratory measurements. There is
 158 also little knowledge of partitioning between heat and radiated energy from seismology or field
 159 relations; however combining lab and field studies for granitic rock and considering the source
 160 properties of earthquakes observed seismically, the possible range of energy partitioning for
 161 shear melted granitic faults can be addressed as we show next.

162 Earthquakes show a wide range of relationships between shear strength and shear stress during
 163 rupture. The difference can be parameterized to some degree by the slip overshoot [*Savage and*
 164 *Wood*, 1971; *McGarr*, 1994],

$$165 \quad \xi = \frac{\hat{\tau} - \tau_1}{\Delta\tau_s} \quad (4)$$

166 where $\Delta\tau_s$ is the static stress drop, the difference between the initial and final stresses (**Figure 2**).
 167 Throughout the following analysis we take the initial stress to be approximately equal to the
 168 static fault strength; this is the second crucial assumption. This is controversial, especially for
 169 plate boundary-scale faulting [*Lapusta and Rice*, 2003; *Noda et al.*, 2009]. The assumption also
 170 differs from the general example in **Figure 2** where the initial stress is lower than the static fault
 171 strength (the peak strength τ_p in **Figure 2**). Such differences and the implications when this
 172 assumption is relaxed are dealt with in the Discussion section below.

173 The static strength of the andesitic and granitoid rocks of the motivating studies of *Sibson*
 174 [1975] and *Di Toro et al.* [2005] follow Byerlee's law approximately [*Byerlee*, 1978] (**Figure 1**).
 175 To estimate the stresses at depth we use guidance from the field studies of *Di Toro and*
 176 *Pennacchioni* [2004] and *Di Toro et al.* [2006] who used Andersonian assumptions for strike-slip

177 faulting [Anderson, 1951]. We use the mean depth of 10 km, a lithostatic stress gradient of 26
178 MPa/km and assume that the intermediate principal stress is equal to the mean stress and then
179 average the results for hydrostatic pore pressure and dry conditions. The details of the estimate
180 are in the Appendix. The effective normal stress is 122 MPa resulting in an initial stress of $\tau_0 =$
181 104 MPa for a Byerlee friction of 0.85. According to the regression of *Di Toro et al.* [2006] at 10
182 km depth the average dynamic strength is $\hat{\tau} = 10.6$ MPa. This coseismic shear strength is lower
183 than the mean value inferred from the field study, $\hat{\tau} = 26.8$ MPa. Here and throughout we report
184 stress estimates to the tenths of MPa. This choice should not be interpreted as the accuracy of the
185 estimate which is unlikely to exceed a few MPa. However, we are interested in seismologic
186 stress measurements, particularly stress drop, that can often be two to three orders of magnitude
187 smaller than the above quoted initial stress (see the subsequent **Figure 3**). As a consequence the
188 apparent accuracy of stresses in this report is required to estimate stress drop in our analyses.
189 Typical stress drops are a few MPa and our reported stresses are to the order of 10% of that.

190 In the following we consider four possible scenario earthquake source parameters for shear
191 melting at this depth. The scenarios are intended to span the range of plausible seismically
192 observed source properties. For all four scenarios we calculate source parameters using the
193 average field measured shear strength of 26.8 MPa. These results are described in the
194 immediately following text and listed in **Table 1**.

195
196 Scenario 1 is the Orowan condition where the stress drops exactly to the dynamic fault
197 strength $\hat{\tau} = t_1$ [Orowan, 1960; Kanamori and Heaton, 2000], then $\Delta\tau_s = 77.2$ MPa, the
198 overshoot (4) is zero, $\bar{\tau} = 65.4$ MPa and $\tau_a = 38.6$ MPa. This would be a case of high seismic
199 efficiency relative to that which has been assumed for pseudotachylite [Di Toro et al., 2006a],
200 $h = t_a/\bar{\tau} = 0.59$; 59% of the total energy would be radiated. Because the Orowan condition is
201 the most often used assumption in studies of the earthquake energy budget, such as in a number
202 of seminal contributions, compilations and reviews [e.g., Kanamori and Heaton, 2000,

203 *Venkataraman and Kanamori, 2004; Kanamori and Brodsky, 2004; Abercrombie and Rice,*
204 *2005; Viesca and Garagash, 2015*], it is useful for placing estimated source parameters and their
205 uncertainty in context. For example, had we used the upper limit of the field estimated fault
206 strength (48 MPa) rather than the average, the resulting seismic efficiency of 37% would still be
207 much higher than typical seismological estimates [e.g., *Wyss, 1970; McGarr, 1999*].

208

209 Scenario 2 is complete stress drop, then, $\bar{\tau} = 52$ MPa, $\xi = 0.26$, and $\tau_a = 25$ MPa, again, a case of
210 high seismic efficiency $h = t_a / \bar{\tau} = 0.48$.

211

212 Both of these scenarios 1 and 2 would be out of the range of typical earthquake source
213 properties, as follows.

214 In the following analysis we use the stress drops of a recent global compilation [*Allman and*
215 *Shearer, 2008*] for reference. These are determined from seismically inferred corner frequencies
216 (f_c) using the Madariaga source model [*Madariaga, 1976*]. Because stress drops depend on
217 $(f_c/C)^3$ where C is a model-dependent scalar, small differences in the scalar (model) produce
218 much large differences in stress drop, up to a factor of 5.5 [e.g., *Kaneko and Shearer, 2014*].
219 Thus constraints on source properties from stress drop are weak. Specific differences between
220 models and the difficulties that arise in using stress drop in studies of source physics are
221 discussed in section 3.2 below. Typical values of stress drop are a few MPa albeit with
222 significant logarithmic variability (**Figure 3**, after *Allman and Shearer [2009]*). The dashed lines
223 that are superimposed mark 99, 95, and 90% of the stress drops in the Allman and Shearer
224 dataset. For instance, 1% of the earthquakes have stress drops larger than the 99% line, and so
225 on. The 99, 95 and 90% lines are associated with stress drops of 110 MPa, 40 MPa and 23 MPa,
226 respectively. Stress drops as large as those in scenarios 1 and 2 are found only in a few percent or
227 less of natural earthquakes. This apparent inconsistency between seismologically inferred values
228 of MPa static stress drop and the ~ 77 MPa dynamic stress drop from the field and extrapolated

229 from laboratory observations of melting (**Figure 1**) is a paradox long expected from theoretical
 230 considerations of shear heating [Sibson, 1975; Lachenbruch, 1980; Rice, 2006; Noda et al.,
 231 2009]. Similar, but potentially stronger constraints on source properties come from apparent
 232 stress because it is not model dependent. For comparison with the scenario estimates of apparent
 233 stress, **Figure 4** shows apparent stresses compiled by *Baltay et al.* [2010]. The estimated
 234 apparent stresses using Orowan's (Scenario 1) and the complete stress drop (Scenario 2)
 235 assumptions are outside the range of these seismic observations that lie between 0.1 and 10 MPa
 236 (**Figure 4**).

237 We also consider the implied overshoot of these scenarios (**Table 1**). The energy balance with
 238 stress as the dependent variable (3) can be rewritten in terms of stress drop, overshoot and
 239 apparent stress as

$$240 \quad \frac{\tau_a}{\Delta\tau_s} = 0.5 - \xi \quad (5)$$

241 [*Savage and Wood*, 1971; *McGarr*, 1994; 1999]. Keep in mind that the model dependence of
 242 stress drop means that bounds on overshoot are dependent on the choice of source model; for all
 243 the standard source models stress drop tends to be a fixed factor of apparent stress [e.g., *Singh*
 244 *and Ordaz*, 1994; *Kaneko and Shearer*, 2014]. Since both stress drop [*Hanks*, 1977] and apparent
 245 stress [*Ide and Beroza*, 2001] are arguably magnitude independent, earthquake overshoot is also
 246 magnitude independent according to (5). For the Madariaga model at 0.9β , slip overshoots the
 247 static value by 20% [*Madariaga*, 1976], which corresponds to a stress measure of overshoot (4)
 248 of 0.17 which is not so different from scenario 2. Because they involve restrictions on stress
 249 drop, with the exception of overshoot, the source parameters from scenarios 1 and 2 are
 250 independent of the choice of source model; this is not the case for scenarios 3 and 4 that follow.

251
 252 Scenario 3 is typical stress drop. Instead of complete stress drop or Orowan's assumption, take
 253 the stress drop to be $\Delta\tau_s = 3.8$ MPa, then, $\bar{\tau} = 102$ MPa, $\xi = -19.3$, and $\tau_a = 75$ MPa. This would be
 254 a case of extreme undershoot; undershoot larger than can be inferred from seismic observations

255 (see analysis of data of *Venkataraman, and Kanamori*, [2004], in *Beeler*, 2006), and again, high
256 seismic efficiency $h = t_a/\bar{\tau} = 0.73$.

257

258 Scenario 4 is typical overshoot, $\xi = 0.17$, leading to $\Delta\tau_s = 93$ MPa, $\bar{\tau} = 57.5$ MPa and $\tau_a = 30.7$
259 MPa, this too would be a case of high seismic efficiency $h = t_a/\bar{\tau} = 0.53$.

260

261 To put the scenarios in context with seismological observations they are plotted versus seismic
262 moment in **Figures 3 and 4** by assuming a circular rupture. Using the average slip from the
263 exhumed pseudotachylites of 0.59 m [*Di Toro et al.*, 2006], and the stress drops from **Table 1**,
264 we can calculate the radius

265
$$r = \frac{7\pi\mu\Delta\delta}{16\Delta\tau_s}, \quad (6)$$

266 (area $A = \pi r^2$ and seismic moment $M_0 = \mu A \Delta\delta$, **Table 1**). For all scenarios the apparent stress is
267 outside the typical values. All the stress drops except for the case where a typical value was
268 assumed are in the upper few percent of the observations. More extreme earthquake source
269 properties result if the lab-inferred value of the melt shear strength is used instead of the field
270 values.

Discussion

271 Partitioning of radiated and thermal energy during earthquake slip might be most easily
272 considered by normalizing equation (3) by the average stress, defining a total thermal efficiency,

273
$$\frac{\hat{t}}{\bar{\tau}} = 1 - \eta, \quad (7)$$

274 the ratio of the average dynamic shear strength to the average co-seismic shear stress, where η is
275 the seismic efficiency as defined above. As noted by *McGarr* [1994; 1999], for dynamic rupture
276 controlled by low temperature friction at very small displacements, the thermal efficiency is
277 high, for example, greater than 90% [*Lockner and Okubo*, 1983], and the seismic efficiency is
278 less than 10%. However, for much more extreme dynamic weakening, such as seen for shear

279 melts with low dynamic shear strength, so long as the initial stress is high, the seismic efficiency
280 must be significantly larger than it is in low temperature friction experiments.

281 In this context, we can draw a number of conclusions about earthquake source properties
282 associated with the pseudotachylites. Based on our four scenarios, we expect that radiated energy
283 will be similar to or exceed shear heating during the earthquake-generated formation of natural
284 shear melts, equivalently the seismic efficiency is similar to or exceeds the thermal efficiency. A
285 related conclusion is that, because the radiated energy is large, from equation (3), fault shear
286 stress during earthquakes cannot be estimated from exhumed pseudotachylite; the estimates from
287 previous studies assumed negligible radiated energy and directly equated shear stress with the
288 field-measured strength. Thus the estimates from prior studies are likely an implausible lower
289 bound on the shear stress and if so the field studies of exhumed pseudotachylite have
290 underestimated stress. The degree that stress differs from strength depends on how much the slip
291 overshoots (or undershoots) the value that would result from the dynamic stress drop alone (the
292 difference between the final stress and the shear strength) and also on the ‘strength excess’ (how
293 much the failure strength of the fault exceeds the initial stress, see discussion below). Our
294 calculations suggest underestimation by 1.9 to 2.8 times. Overshoot is not determined in the
295 existing shear melting laboratory experiments but it is an active target for laboratory
296 investigation [e.g., *Sone and Shimamoto, 2009; Di Toro et al., 2011a*]. Overshoot might
297 reasonably be inferred from careful measurement in subsequent tests or in relatively simple
298 calculations of dynamic shear melting. According to this analysis, earthquakes that produce
299 pseudotachylite are outside the range of seismic observations of apparent stress (**Figure 4**).

Reconciling the energy balance.

300 There are, however, a number of ways in which our energy accounting may have gone astray.
301 Much uncertainty in our balance is associated with the choice of a Madariaga source model that
302 has the largest stress drop of the conventional models. Still, had we used a dataset in which the
303 stress drops were determined using the Brune model that has the lowest stress drops, apparent

304 stress still would be out of the bounds of the *Baltay et al.* [2010] dataset for all four scenarios,
305 and the discrepancy between the predicted and observed stress drops would be even larger. As
306 above, while acknowledging that the choice of source model has first order implications for
307 earthquake source properties, source model choice does not effect our conclusion that the
308 presence of pseudotachylite implies an unusual earthquake source. Additional discussion of
309 source models is found in section 3.2 below.

310 We now consider whether relaxing the two critical assumptions about initial stress and
311 dissipated energy may allow shear melting to produce more typical earthquake source properties.
312 First, we have assumed that the heat inferred from pseudotachylite is equivalent to all energy that
313 does not go into the radiated field (i.e., $\hat{t} = \hat{t}_m$). This ignores any off-fault damage that may be
314 generated during rupture, such as brittle failure associated with stress concentrations about the tip
315 of the propagating rupture [*Andrews*, 1976; 2005] or from slip on rough fault surfaces [*Chester*
316 *and Chester*, 2000; *Dieterich and Smith*, 2009; *Dunham et al.*, 2011]. Such energy is most often
317 partitioned into a 'shear fracture energy' term in an expanded energy balance [e.g., *Tinti et al.*,
318 2005]. Fracture energy is heat and latent-heat, the energy that goes into the creation of shear and
319 tensile fracture surfaces and into slip on shear fractures in the damage zone about the rupture
320 [*Ida*, 1972; *Andrews*, 1976]. In well-posed dynamic rupture models it is the portion of this
321 energy associated with inelastic deformation about the tip of the rupture that limits the
322 propagation speed [*Andrews*, 1976; 2005]. *Andrews* [2005] has further shown that the size of this
323 energy contribution scales with the dynamic stress drop, thus mechanisms such as shear melting,
324 which produce large strength losses, implicitly require some compensation in off-fault fracture
325 energy as well as in radiation.

326 Second, we have assumed up to this point that the initial stress is approximately equal to the
327 static fault strength which, in the case of the felsic crystalline rocks of the motivating studies,
328 implies high initial stress in the crust. If instead we assume that the initial stress is lower than the
329 failure stress, as depicted in the schematic **Figure 2**, there is a strength excess, S_e defined by the

330 difference between the failure strength and the initial stress [Andrews, 1985]. Such an excess
 331 arises naturally in regions with strength or stress heterogeneity. For example imagine a fault
 332 surface that on average is strong but with a limited contiguous region of weak material. If the
 333 incipient rupture starts in that weak area and that region is sufficiently large and slips far enough
 334 to raise the stress on the adjacent portion of the strong region to its failure stress, then an
 335 earthquake rupture can occur at a lower stress than the average failure strength of the fault.

336 To relax both critical assumptions about initial stress and dissipated energy we modify
 337 equation (3). To consider contributions of damage to source properties it is convenient to use a
 338 stress-measure of fracture energy. Fracture energy, G_e , has the dimensions of energy per unit
 339 area, so the 'fracture stress' then is fracture energy divided by the total slip, $t_c = G_e/Dd$. Replace
 340 the shear resistance in (3) with the sum of that which goes in to shear heat and that which resides
 341 in fracture energy, $\hat{t} = \hat{t}_m + t_c$. To incorporate the strength excess we replace the average stress
 342 in (3) with $t_0 - Dt_s/2$, and replace the initial stress with $t_p - S_e$. Making these substitutions the
 343 balance (3) becomes

$$344 \quad t_c + S_e = t_p - \frac{Dt_s}{2} - t_a - \hat{t}_m. \quad (8a)$$

345 Implementing (8a) for pseudotachylite, $\tau_p = 104$ MPa, and $\hat{t}_m = 26.8$ MPa. To produce a stress
 346 drop within the 95% bound and apparent stresses to be at the upper limit of the observations,
 347 corresponding to $\Delta\tau_s = 40$ and $\tau_d = 10$ MPa, respectively, (8a) is

$$348 \quad t_c + S_e = 47.2 \text{ MPa} \quad (8b)$$

Fracture energy.

349 If the right-hand side of (8b) were all due to fracture energy ($S_e=0$), the fracture stress would
 350 exceed the stress drop. For comparison with typical observations, a measure of the associated
 351 efficiency is the ratio of fracture energy times the fault area to the energy associated with the
 352 stress drop: $h_c = G_e/Dt_sDd$; equivalently the ratio of the fracture stress to the stress drop:
 353 $h_c = t_c/Dt_s$. *Beeler et al.* [2012] compiled some limited and model-dependent data on this

354 efficiency from *Abercrombie and Rice* [2005] and found no natural values greater than 0.5. The
355 minimum fracture efficiency to bring the pseudotachylite data in line with typical earthquakes is
356 1.2. However, as none of the prior estimates of fracture stress or efficiency strictly include off-
357 fault damage or consider the impact of roughness on fracture energy, these remain topics for
358 further research.

The strength excess and fault roughness.

359 Consider instead that all of the right-hand side of (8b) was from the strength excess ($\tau_c=0$),
360 then the difference between the initial stress and the failure strength would be ~ 47 MPa. In that
361 case the heterogeneity would have to be quite high in association with these earthquakes in
362 crystalline rock. Since the source region is a batholith and arguably not highly heterogeneous in
363 elastic or friction properties we can only appeal to stress heterogeneity to produce the necessary
364 strength excess. Some insight into the allowable amplitude of stress heterogeneity may be found
365 in studies of roughness contributions to shear strength [*Chester and Chester*, 2000; *Dieterich and*
366 *Smith*, 2009; *Dunham et al.* 2011; and *Fang et al.* [2013]. The idea is that fault shear resistance
367 consists of two components, the shear resistance due to frictional slip on a planar fault surface,
368 and that which results from fault roughness. Based on measurements of natural fault roughness,
369 the amplitude to wavelength ratio α appropriate for faults that host intermediate sized
370 earthquakes is between 10^{-3} and 10^{-2} [*Power and Tullis*, 1991; *Sagy and et al.*, 2007]. According
371 to the modeled estimates to date, the upper end of this range produces dramatic stress
372 heterogeneity on the fault and significant additional shear strength beyond the interface friction
373 [*Chester and Chester*, 2000; *Dieterich and Smith*, 2009], deemed roughness drag, τ_{drag} , by *Fang*
374 *and Dunham* [2013]. How roughness may define the strength excess would be to allow
375 earthquake nucleation on relatively flat portions of the fault at stress levels equal to the frictional
376 strength.

377 Since roughness drag increases the shear heating above that associated with slip on planar
378 surfaces with the same frictional strength [*Griffith et al.*, 2010], this contribution is included in

379 the pseudotachylite-estimated co-seismic shear strength (1b). Drag may be used to explain the
380 difference between lab and field-measured values of shear strength. Formally

$$381 \quad t_{drag} = \frac{8\rho^3 a^2 G' D d}{l_{min}}, \quad (9)$$

382 where λ_{min} is the minimum wavelength of the roughness and G' is the shear modulus divided by
383 1 minus Poisson's ratio [Fang and Duham, 2013]. Taking the ratio of slip to λ_{min} to be of order
384 one [Fang and Dunham, 2013], the amplitude ratio is $a = \sqrt{t_{drag}/G'8\rho^3}$. Assuming the
385 difference between the lab and field shear strengths (~16 MPa) is the dynamic roughness drag,
386 and $G'=40$ GPa, then $\alpha = 0.0013$.

387 The roughness drag as estimated by Fang and Dunham [2013] (9) and in the prior study by
388 Dieterich and Smith [2009] is calculated for a discontinuity in otherwise intact rock assuming a
389 small amount fault slip relative to the smallest wavelength of roughness, elastic stress transfer,
390 and no dilatancy. Results of these assumptions are that the roughness drag is not pressure
391 dependent and it does not depend on the absolute level of the differential stress. As such the
392 same roughness drag applies to both the sliding and failure strengths, at all depths, so long as the
393 amplitude and characteristics of the roughness are not changed substantially by slip or by
394 ambient stress levels. Accordingly our estimated value of 16 MPa inferred from sliding is also
395 the strength excess due to fault roughness-generated stress heterogeneity. Even if we allow that
396 our failure strength of 104 MPa is overestimated by 16 MPa, that is not enough of a strength
397 excess to bring the pseudotachylite source properties in line with more typical earthquakes.

398 Admittedly these estimates do not consider contributions from material heterogeneity;
399 nonetheless those should be small in the relatively homogeneous source region of the
400 pseudotachylite. Contributions from slip heterogeneity are also not considered. Since those will
401 correlate with fault roughness in a homogeneous material [Duham et al., 2011; Fang and
402 Dunham, 2013] we expect that the difference between our estimate and the needed value of 47
403 MPa precludes reconciling the observations and typical earthquake source properties with this
404 model of the strength excess. Nonetheless, given that our roughness estimate is based entirely on

405 the difference between field and lab melt shear strengths, along with the large uncertainties
406 associated with the field-inferred strength, and our assumption of the high Byerlee failure
407 strength, the strength excess remains perhaps the most poorly constrained of all the poorly
408 constrained earthquake source properties.

409 To assess whether the combined effects of strength excess and fracture energy are sufficient to
410 bring pseudotachylite into line with typical earthquakes, use the strength excess of 16 MPa in
411 (8b) to reduce the needed fracture stress from 47 to 31 MPa. The associated minimum fracture
412 efficiency would be ~ 0.8 , exceeding the limited observations [Abercrombie and Rice, 2005] by a
413 factor of 1.5. Again we conclude that seismically generated pseudotachylite requires atypical
414 earthquake source properties, a result that seems robust even when limitations of the assumptions
415 are taken into account.

Future work on fault roughness.

416 There are physical limits on the estimate of roughness drag in equation (9). The underlying
417 theory breaks down at high but realistic amplitude ratios [Dieterich and Smith, 2009; Fang and
418 Duhnam, 2013], especially at near surface and intermediate depths. For example, at a modest
419 effective normal stress of 100 MPa the strength of intact granite is about 150 MPa while the
420 frictional strength is about 85 MPa. From (9), using the same slip and elastic assumptions as
421 previously, the roughness drag of a fault at the upper end of the natural amplitude ratio range,
422 $\alpha = 0.01$, is 990 MPa, more than ten times the frictional strength and approximately six times the
423 intact rock strength. Empirically this is out of bounds and arises mostly because the estimate
424 forbids the dilatancy that limits rock and fault strength in the first place [Brace et al., 1966;
425 Escartin et al., 1997]. Similarly at more modest values of the amplitude ratio but at greater depth
426 where the normal stress is high, according to (9), friction will dominate the shear resistance as
427 friction increases with normal stress while the roughness contribution does not. This is hard to
428 reconcile with existing laboratory data in which both sliding friction and intact rock strength
429 increase with confining pressure. In practice many of these issues with (9) are dealt with in

430 numerical fault models [Fang and Dunham, 2013]. There, the stresses that arise from slip on
431 rough surfaces are calculated incrementally with slip (rather than assuming that $\Delta\delta/\lambda_{min}=1$) and
432 when the drag stress reaches the failure strength of surrounding rock the material yields via a
433 separate pressure dependent plasticity relation.

434 Simpler models of rough faults and the bounds on the resulting stress heterogeneity might be
435 constructed using existing laboratory data. Among the non-physical aspects of the theory
436 underlying (9) are: no dilatancy and that the fault is zero-thickness and fully localized resulting a
437 stationary shear zone. On the latter, natural fault zones have finite thickness that likely provides
438 some degree of freedom to deform internally to accommodate roughness of the fault bounding
439 rock. On the former, disallowing rigid and fracture dilatancy on a fault between rock surfaces is
440 contrary to the most basic physical observations of brittle deformation and frictional slip [e.g.,
441 Brace et al., 1966; Marone et al., 1990]. Because of these issues we suggest that the contribution
442 of roughness to fault shear resistance is inherently pressure dependent, such that it is smaller than
443 (9) at near surface conditions where, in the presence of very low normal stress and distributed
444 shear, roughness likely leads to rigid dilation rather than damage in the surrounding rock, and
445 also so that the contribution from roughness does not diminish relative to friction at elevated
446 confining pressure. Furthermore the roughness contribution is bounded by existing experimental
447 data to be less than or equal to the strength of intact rock minus the frictional failure strength at
448 the confining pressure and temperature of interest. Future experiments on faults with amplitude
449 ratios between 0.01 and 0.001, at effective normal stresses and temperatures spanning those of
450 the brittle crust should better establish the contributions of roughness to fault strength.

Stress drop and the choice of source model.

451 Choice of source model has a very large effect on the inferred bounds of static stress drop,
452 such as the 95% bound $\Delta\tau_s = 40$ MP from Allman and Shearer [2009] that is superimposed on
453 **Figure 3**. The Madariaga source produces stress drops that are a factor of 2.6 larger than from
454 the Sato and Hirasawa [1973] model and 5.5 times larger than Brune [1970]. Decreasing the

455 upper bound in **Figure 3** to that which would be inferred from *Brune* [1970], would place all
456 scenarios except #3 further out of range of typical stress drops. This model dependency of static
457 stress drop is a significant barrier to using stress drop as a metric in studies of source physics
458 [McGarr, 1999]. And while there is no strict constraint on stress drops from pseudotachylite, our
459 analysis suggests that regardless of the source model used the stress drops from pseudotachylite
460 are unusual for earthquakes.

461 There are, unfortunately, additional fundamental problems relating the stress drop from
462 standard source models to pseudotachylite. For each of the Brune, Sato and Hirasawa and
463 Madariaga source models, the ratio of apparent stress to static stress drop is fixed with a value
464 $0.22 < \tau_a / \Delta\tau_s < 0.4$. In otherwords, these are all crack-like rupture models that overshoot. In
465 contrast, experimental measurements suggest that the shear melts show rapid 'co-seismic'
466 strength recovery [Di Toro et al., 2011a] that, when extrapolated to a propagating, confined
467 rupture, are more consistent with undershoot and pulse-like propagation. In the absence of a
468 definitive earthquake source model that allows for undershoot or seismic methods that reliably
469 distinguish undershoot from overshoot it will remain difficult to use static stress drops to relate
470 laboratory observations to earthquake seismology.

Source properties of shear melts.

471 The source parameters in scenarios 1 to 4 are perhaps the seismic corollary to the
472 interpretation of the geologic record that pseudotachylite is rare [Sibson and Toy, 2006].
473 Although the interpretation is not without controversy [Kirkpatrick et al., 2009; Kirkpatrick and
474 Rowe, 2013], the corollary is not unexpected. While pseudotachylite is known to form under a
475 wide range of conditions, for example in presence of fluids, in metamorphic terrains and even in
476 large events within melange [e.g. Toy et al., 2011; Bjornerud et al., 2010, Meneghini et al.,
477 2010], the friction melting experiments of Di Toro et al. [2006a] suggest that pseudotachylites
478 are easily formed during imposed localized slip on pre-cut faults in cohesive rocks that are dry.
479 Many field studies also suggest that the typical ambient conditions of pseudotachylite is the dry

480 crystalline basement of the continental crust [Sibson and Toy, 2006], as is the case for most
481 nappes in the Western Alps, where pseudotachylites are not uncommon fault rocks
482 [Pennacchioni et al., 2007]. The higher stress drops characteristic of intraplate earthquakes
483 [Scholz et al., 1986], including those of some very high stress drop earthquakes [e.g., Viegas et
484 al., 2010; Ellsworth et al. 2011] may indicate related properties of the source, once differences in
485 source model are accounted for. Large stresses relative to the failure strength, large stress drops,
486 and relatively low fault roughness may lead to some diagnostic rupture properties associated
487 with pseudotachylite formation. High initial stress levels promote a strong tendency for super-
488 shear rupture up to the compressional wave speed, specifically when the ratio of the strength
489 excess to the dynamic stress drop, S , is lower than 1.77 [Andrews, 1985] as claimed to be
490 observed experimentally by Passelegue et al. [2013]. Taking the 16 MPa strength excess, an
491 initial stress of 104 MPa, and sliding strength of 26.8 MPa, Andrews' S ratio is no higher than
492 0.26 and super shear rupture is expected. A large stress drop, low roughness and high initial
493 stress may also tend to promote propagation as an expanding crack rather than as a slip pulse
494 [Zheng and Rice, 1998].

495 An appealing third idea explaining the difference between typical earthquake stress drops and
496 the ~ 77 MPa values inferred for pseudotachylite dynamic stress drops relaxes our implicit
497 assumption that pseudotachylites are representative of the dynamic properties of the earthquakes
498 that generated them. Sibson [2003] suggested that faults have significant spatially varying
499 dynamic properties, allowing the majority of the shear strength to be concentrated in regions of
500 high geometric complexity (e.g., fault bends or step-overs). Fang and Dunham [2013] reached a
501 similar conclusion when considering large ruptures. This kind of model, where part of the fault is
502 dynamically weak but most of the shear strength is concentrated elsewhere, perhaps in relatively
503 limited areas, is similar to the numerical fault models with heterogeneous stress conditions that
504 allow fault slip at low average stress levels [Lapusta and Rice, 2003; Noda et al., 2009]. Under
505 the Sibson [2003] conceptual model, pseudotachylite is generated on parts of the fault that are

506 geometrically simple prior to rupture, but it does not contribute significantly to the dynamic
507 shear strength of the entire fault. Our scenario 3 where we have imposed a typical stress drop is
508 related to this kind of event. Doing so requires that the rupture dimension is much larger than the
509 other scenarios, producing an M6 earthquake. In any event, the Sibson model would remove the
510 discrepancy between typical earthquake stress drops and the implied strength loss by
511 pseudotachylite in granite rock and would allow pseudotachylite to be more common as
512 advocated by *Kirkpatrick and Rowe* [2013]. Meanwhile the mechanical properties of
513 pseudotachylite would be largely irrelevant to the average seismically-inferred source properties
514 such as static stress drop and apparent stress. Testable implications of this model would be that
515 during seismic slip the majority of shear generated heat would be concentrated in distinct local
516 regions of low stress drop. In cases where the stress is high, regions of low shear strength due to
517 the formation of pseudotachylite would appear as 'asperities' in seismic inversions where the
518 stress drop and radiated energies are high [e.g., *Kanamori, 1994; Bouchon, 1997; Kim and*
519 *Dreger, 2008*]. A hope is that the character of radiated energy from such asperities could be
520 quantitatively related to laboratory and field studies of fault properties and in some cases related
521 to a particular shear deformation mechanism in the fault zone (e.g., melting, thermal
522 pressurization). This would require particular mechanisms to have characteristic source
523 properties, for example a distinctive frequency content. Making such a link between various
524 source properties and source mechanisms might be made using synthetic seismograms generated
525 by spontaneous dynamic rupture simulations [e.g., *Andrews, 2005; Harris, 2004*], as
526 developments in that field are directed specifically at the physics within the source [*Harris et al.,*
527 2009].

Conclusions

528 The analysis of the energy budget and source properties of pseudotachylite-generating
529 intermediate sized earthquakes of the Gole Larghe fault zone in the Italian Alps where the
530 dynamic shear strength is well-constrained by field and laboratory measurements suggests these

531 earthquakes have unusual source parameters. The assumptions are: that seismically determined
532 corner frequency relates to stress drop by the *Madariga* [1976] relation, that the heat inferred
533 from pseudotachylite thickness and fault displacement is equivalent to all energy that does not go
534 into the radiated field, and that the initial stress is approximately equal to the static fault strength.
535 For the felsic crystalline rocks of the source region, the final assumption results in an initial shear
536 stress on the order of 100 MPa. Stress drops and apparent stress are larger than a few 10 's of
537 MPa, unlike typical earthquakes, and the radiated energy equals or exceeds the shear-generated
538 heat. Relaxing these assumptions, the observations still cannot be reconciled with typical
539 earthquake source properties unless fracture energy is routinely significantly greater than in
540 existing models, pseudotachylite is not representative of average fault shear strength during the
541 earthquake that generated it, or unless the strength excess is larger than we have allowed.

542

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545

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Appendix - Estimated initial stress

784 The hypocentral source region of the pseudotachylite at Gole Larghe was at approximately 10
785 km depth, in a strike-slip faulting regime in Tonalite [Di Toro and Pennacchioni, 2005; Di Toro
786 *et al.*, 2005]. To estimate the ambient stress level we follow these cited prior studies and assume
787 an Andersonian strike-slip regime [Anderson, 1951] in which the lithostatic stress from
788 overburden σ_L is the mean of the greatest and least principal stresses $S_L = S_m = (S_1 + S_3)/2$. The
789 fault is optimally oriented for failure in the stress field and assumed to limit the stress level in the
790 surrounding rock. These conditions are depicted in the Mohr diagram (**Figure A1**), where the
791 fault is assumed to be cohesionless with a friction coefficient $\mu = \tau/\sigma_e$, defining the friction angle
792 $\mu = \tan \phi$, τ is shear stress, σ_e is the effective normal stress ($\sigma_e = \sigma_n - p$), σ_n is normal stress and
793 p is pore fluid pressure. Here the ratio of pore pressure to the lithostatic stress is denoted by the
794 ratio $\lambda = p / \sigma_L$ [Sibson, 1974]. From the Mohr construction (**Figure A1**), effective normal stress
795 is

$$796 \quad \sigma_e = S_L (1 - \lambda) \cos^2 \phi. \quad (\text{A1})$$

797 The lithostatic gradient is taken to be 26 MPa/km and $\sigma_L = 260$ MPa. To estimate a representative
798 effective normal stress we follow Di Toro *et al.* [2005] and average the results from assuming
799 the pore pressure is hydrostatic with pore pressure gradient 10 MPa/km, with those from
800 assuming dry conditions. That is, using $\lambda = 10/26$ and $\lambda = 0$ in (A1), resulting in $\sigma_e = 93$ and 151
801 MPa, and a representative $\sigma_e = 122$ MPa for $\mu = 0.85$ [Byerlee, 1978] that is appropriate for
802 crystalline rock. These assumptions correspond to a shear resistance at failure of $\tau = 104$ MPa.

803 **Table 1.** Possible earthquake source properties for shear melting at 10 km depth, effective normal stress = 122 MPa and initial stress
 804 of 104 MPa.

scenario	average strength, (MPa) \hat{t}	static stress drop, (MPa) $D t_s = t_0 - t_1$	apparent stress, τ_a (MPa)	average stress, (MPa) $\bar{t} = \hat{t} + t_a$ $\bar{t} = (t_0 + t_1)/2$	overshoot, $\chi = 0.5 - t_a/D t_s$	seismic efficiency, $h = t_a/\bar{t}$	thermal efficiency, \hat{t}/\bar{t}	r (m)	A (m ²)	Moment (Nm)
Orowan $\hat{t} = t_1$	26.8	77.2	38.6	65.4	0	0.59	0.41	78.7	3.1e5	1.1e15
complete stress drop $D t_s = t_0$	26.8	104	25.2	52	0.26	0.48	0.52	61.8	1.7e5	1.3e15
typical stress drop $\Delta \tau_s = 3.9$ MPa	26.8	3.9	75.3	102	-19.3	0.73	0.27	2575.8	1.2e8	1.2e18
typical overshoot $\chi = 0.166$	26.8	93	30.7	57.5	0.166	0.53	0.47	59.0	2.1e5	3.2e15

805 Four scenarios are considered and source parameters are tabulated for an average shear strength of 26.8 MPa (field). For each
 806 scenario the assumed values are in bold in the Table. The values for the stress parameters in the Table can be derived directly from
 807 the initial, and average strength, the definitions in the column headers, and the assumptions that are listed in the scenario rows, using
 808 the assumed (bold) table values.

Figure Captions

809

810 **Figure 1.** Natural and laboratory observed shear strength of granitic melt. Shown for reference
811 is the approximate static strength of pre-existing faults in granitic rocks (solid line) [Byerlee,
812 1978]. The dashed line is the regression of experimental data from *Di Toro et al.* [2006b],
813 extrapolated to higher normal stress. The field inferred shear strengths of *Di Toro et al.* [2005;
814 2006a], that are calculated from measured thickness-displacement ratios using equation (1c), are
815 plotted as the open symbols at the inferred mean normal stress. The box shows the range of
816 possible field-inferred shear and normal stresses.

817

818 **Figure 2.** Earthquake stress versus slip diagram after *McGarr* [1994]. Fault strength is shown
819 as the heavy black line while shear stress is the heavy black dashed line between τ_0 and τ_l , the
820 starting and ending stresses. The average stress, $\bar{\tau}$, is denoted by the heavy grey dashed line and
821 the average fault strength, $\hat{\tau}$, by the grey dashed line. The apparent stress is the difference
822 between these lines. This example is a case of overshoot [*Savage and Wood*, 1971] where the
823 final stress is less than the average strength. This is also a case where the starting stress is lower
824 than the failure strength τ_p , defining a strength excess S_e .

825

826 **Figure 3.** Variation of stress drop with seismic moment. Stress drops from the previous studies
827 of *Abercrombie* [1995], *Tajima and Tajima* [2007] and *Allman and Shearer* [2009]. Here all
828 stress drops are calculated using the *Madariaga* [1976] model. In the case of *Tajima and Tajima*
829 [2007], the stress drops were calculated using their tabled moment and corner frequency, f_c ,
830 using $Dt = M_0 (f_c / (0.42b))^3$ and $\beta = 3.9$ km/s, assuming rupture propagation at 0.9β , as in
831 *Allman and Shearer* [2009]. An implication of these and other compilations [e.g., *Hanks*, 1977;
832 *Baltay et al.*, 2011] is that stress drop is moment independent. The dashed lines are the 99, 95,
833 and 90% boundaries from the global dataset of *Allman and Shearer* [2009] (solid circles). For

834 example 1% of the stress drops are larger than the 99% line (110 MPa). The 95 and 90% lines
835 are stress drops of 40.3 and 22.9 MPa, respectively. Stress drops from exhumed pseudotachylite
836 for the scenarios listed in **Table 1** are shown in grey. Moment is calculated assuming a circular
837 rupture, equation (6) in the text, a shear modulus $\mu = 30,000$ MPa, the average slip from the
838 exhumed pseudotachylite (0.59 m) and the stress drops for each scenario (**Table 1**), see text.

839

840 **Figure 4.** Variation of apparent stress with seismic moment. Compilation of apparent stress
841 (right axis) from *Baltay et al.* [2010; 2011]. The dashed lines are for 10 and 0.1 MPa and are the
842 approximate bounds on the observations. The implication of this and other compilations [e.g.,
843 *Ide and Beroza, 2001*] is that apparent stress is moment independent. Apparent stresses for
844 exhumed pseudotachylite for the scenarios listed in **Table 1** are plotted in grey. Seismic
845 moments for the pseudotachylite are calculated as described in the caption to **Figure 3**.

846

847 **Figure A1.** Schematic Mohr diagram of the estimated initial stress state for pseudotachylite at
848 the Gole Larghe fault zone (see Appendix text for description).