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25 November 2021

Version of attached file:

Accepted Version

Peer-review status of attached file:

Peer-reviewed

Citation for published item:

Ferrand, Thomas P. and Nielsen, Stefan and Labrousse, Loïc and Schubnel, Alexandre (2021) 'Scaling Seismic Fault Thickness From the Laboratory to the Field.', *Journal of Geophysical Research: Solid Earth*, 126 (3). e2020JB020694.

Further information on publisher's website:

<https://doi.org/10.1029/2020JB020694>

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1 Scaling seismic fault thickness from the laboratory to the field

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10 Abstract

11 Pseudotachylytes originate from the solidification of frictional melt, which transiently forms and lubricates
12 the fault plane during an earthquake. Here we observe how the pseudotachylyte thickness a scales with
13 the relative displacement D both at the laboratory and field scales, for measured slip varying from microns
14 to meters, over six orders of magnitude. Considering all the data jointly, a bend appears in the scaling
15 relationship when slip and thickness reach ~ 1 mm and $100 \mu\text{m}$, respectively, i.e. $M_w > 1$. This bend can be
16 attributed to the melt thickness reaching a steady-state value due to melting dynamics under shear
17 heating, as is suggested by the solution of a Stefan problem with a migrating boundary. Each increment of
18 fault is heating up due to fast shearing near the rupture tip and starting cooling by thermal diffusion upon
19 rupture. The building and sustainability of a connected melt layer depends on this energy balance. For
20 plurimillimetric thicknesses ($a > 1$ mm), melt thickness growth reflects in first approximation the rate of
21 shear heating which appears to decay in $D^{-1/2}$ to D^{-1} , likely due to melt lubrication modulated by melt +
22 solid suspension viscosity and mobility. The pseudotachylyte thickness scales with moment M_0 and
23 magnitude M_w ; therefore, thickness alone may be used to estimate magnitude on fossil faults in the field
24 in the absence of displacement markers within a reasonable error margin.

25 **1. Introduction**

26 Originally evidenced in quartz-bearing rocks (Shand, 1916; Philpotts, 1964) and for large earthquakes
27 (McKenzie & Brune, 1972), frictional melting of the fault surface at depth is a common phenomenon (e.g.
28 Toyoshima et al., 1990; Kanamori et al., 1998; Otsuki et al., 2003; Hirose & Shimamoto, 2005a; 2005b;
29 Spray et al., 2010; Beeler et al., 2016; Ferrand et al., 2017; Aubry et al., 2018). The result of this melting
30 and fast quench is the formation of pseudotachylytes (e.g. Shand, 1916; Sibson, 1975; Ueda et al., 2008;
31 Obata & Karato, 1995; Andersen & Austrheim, 2006; Andersen et al., 2014; Deseta et al., 2014a, 2014b).
32 Located both in the fault vein and injection veins, they consist of either glassy or cryptocrystalline rocks
33 (e.g. Toyoshima, 1990; Lin, 1994; Dobson et al., 2018; Ferrand et al., 2018) originating from the
34 solidification of the frictional melt (Sibson, 1975), along with clasts remaining from the host rock.
35 Pseudotachylytes have also been described as the result of landslides (Masch et al., 1985) and meteoritic
36 impacts (Dietz, 1961; Wilshire, 1971), which are not the topic of the present study. An alternative
37 mechanism has been proposed to explain the formation of pseudotachylytes through intense
38 comminution leading to slow mechanical amorphization (e.g. Wenk, 1978; Pec et al., 2012; Hayward et al.,
39 2016; Marti et al., 2020), which is unlikely especially when either asymmetrical damage/injection (e.g.
40 Griffith et al., 2012; Ferré et al., 2015; Thomas et al., 2017; Petley-Ragan et al. 2019) or fractionated
41 crystallization (e.g. Warr & van der Pluijm, 2005; Ferrand et al., 2018) are observed.

42 Alteration or retrograde metamorphism during the exhumation from great depth (up to >80 km) often
43 affects the preservation of pseudotachylyte (e.g. Kirkpatrick et al., 2009; Kirkpatrick & Rowe, 2013; Phillips
44 et al., 2019). Nevertheless, pseudotachylyte remnants have been observed in various places around the
45 world, especially within lower-crust and upper-mantle outcrops (e.g. Sibson et al., 1975; Obata & Karato,
46 1995; Barker, 2005; Nielsen et al., 2010b; Di Toro et al., 2006; Andersen et al., 2014; Scambelluri et al.,
47 2017). Rheological transitions and fluid-rock interactions tend to entirely transform the mineralogy of fault

48 itself but a significant part of the damage zone, containing millimetric and micrometric faults, can remain
49 unaltered (e.g. Sibson, 1975; Scambelluri et al., 2017; Ferrand et al., 2018).

50 Comparing natural pseudotachylytes to their experimental analogues is key for understanding
51 earthquakes mechanics (e.g. Di Toro et al., 2006). Laboratory studies have been reporting
52 pseudotachylytes in various lithologies (Table 1; Table 2; Fig.1; Fig. 2), during rotary-shear experiments
53 (e.g. Spray, 1987; Di Toro et al., 2006; Hirose & Shimamoto, 2005a; Del Gaudio et al., 2009; Niemeijer et
54 al., 2011), stick-slip experiments (e.g. Passelègue, 2014; Passelègue et al., 2016a; Brantut et al., 2016;
55 Hayward et al., 2017; Lockner et al., 2017) and high-pressure experiments (Incel et al., 2017; Ferrand et
56 al., 2017). Rotary-shear experiments have been able to reproduce “seismic” slip velocities (i.e. $\geq 1 \text{ m}\cdot\text{s}^{-1}$)
57 and metric displacements characteristic of earthquakes of $M_w > 4$ (i.e. $D > 2 \text{ cm}$), providing data that scale
58 with natural earthquakes (Nielsen et al., 2010b; Hirose & Shimamoto, 2005b; Di Toro et al., 2011). Rotary-
59 shear experiments shed new insights on the weakening mechanisms but explain only the frictional part of
60 the seismic process, i.e. they do not include dynamic propagation, and are undertaken at normal stresses
61 that are far lower than would be expected at earthquake hypocentres. Stick-slip experiments encompass
62 both the seismic rupture nucleation and propagation stages (Passelègue et al., 2016a; 2016b; Aubry et al.,
63 2018; Marty et al., 2019) but remain limited by the millimetric to pluricentimetric size of the sample, so
64 that the laboratory magnitudes rarely exceed $M_w -3$ (Passelègue et al., 2016a). In consequence, connecting
65 laboratory results with seismological and field observations remains difficult and controversial.

66 The moment magnitude M_w of an earthquake is directly related to the average displacement D on the
67 fault plane, which allows geologists to give reasonable estimates of M_w based on direct field observations.
68 However, deep lithologies often exhibit a lack of clear displacement markers. Based on field measurements
69 on pseudotachylytes in the Outer Hebrides, Scotland, Sibson (1975) showed that, at the field scale, the
70 relationship between a and D follows a power law. However, field observations of pseudotachylyte
71 thickness are to be considered carefully, because discrete accumulations of glassy or fine-grain material,

72 either in pull-aparts, dilational jogs (e.g. Sibson, 1975) or tensile fractures (e.g. Di Toro et al., 2005; Ferrand
73 et al., 2018) have to be fully **observed/grasped** for a thorough average thickness estimate. **Earthquake**
74 **scaling may help understand slip weakening mechanisms (Abercrombie & Rice, 2005)**. Large and small
75 earthquakes present striking similarities (e.g. Kita & Ferrand, 2018; Ide, 2019; Brodsky, 2019) and appear
76 to nucleate similarly (Ide, 2019; Abercrombie, 2019), which supports that laboratory analogues (e.g. Mw -
77 6, fault = 1mm; Ferrand et al., 2017) may be representative of natural processes (e.g. Mw 6+, fault > 1km;
78 Ferrand et al., 2018). However, it is not clear which parameters control the evolution of small-scale
79 ruptures into larger events. Understanding to what extent laboratory analogues of earthquakes are
80 representative of larger-scale natural events is key for the current advances in earthquakes physics and
81 earthquake magnitude prediction (Kilb & Gombert, 1999; Wyss, 2001).

82 **Slip on a plane lubricated with a silicate melt depends on the effective thickness and continuity of the**
83 **melt layer, and on the viscosity of the melt. In nature, however, faults are not planes. Initial fault geometry**
84 **is controlled by pre-seismic creep and gouge evolution (e.g. Bayart et al., 2018) and significantly impacts**
85 **fault rheology (e.g. Johnson et al., 2001; Nielsen et al., 2016; Romanet et al., 2018), as it may define the**
86 **rock volume to be melted. Slip lubrication leads to an asperity-scale dynamic stress drop much higher than**
87 **seismological values (e.g. Passelègue et al., 2016a). Geometrical heterogeneities should control the**
88 **cumulative melt volume. Strong-motion studies show that the rise time, i.e. required for slip to occur, is**
89 **significantly shorter than the overall rupture duration (Beroza & Mikumo, 1996), which supports a “pulse”**
90 **rupture type (Heaton, 1990).**

91 Here, we compile a number of laboratory and field observations of **pseudotachylyte** thicknesses versus
92 their associated displacements (**Table 1; Table 2; Fig.1; Fig.2**), for slip values ranging from < 10 μm to >
93 1m. Based on energy considerations, we investigate the possible scaling of **pseudotachylyte** thickness with
94 displacement and moment magnitude. This may help understand why some small ruptures propagate into

95 larger ruptures when other ruptures do not, and possibly provide magnitude estimates based on the
96 **pseudotachylyte** thickness.

97 **2. Field and laboratory observations**

98 As illustrated in **Fig. 1**, several works have reported fault thickness a and relative displacement D since
99 the observations of Sibson (1975). We compile documented D and a data (**Fig. 2**) from both natural and
100 experimental earthquakes, deriving from various geodynamic contexts (e.g. mantle vs. crust) or
101 experimental techniques (e.g. stick-slip, rotary shear and D-DIA; **Table 1**).

102 **2.1. Data compilation**

103 Both field observations (e.g. Sibson, 1975; Ueda et al., 2008; Obata & Karato, 1995; Andersen &
104 Austrheim, 2006; Nielsen et al., 2010a; 2010b; Andersen et al., 2014; Scambelluri et al., 2017; Ferrand et
105 al., 2018) and laboratory analogues (e.g. Passelègue et al., 2016a; Incel et al., 2017; Brantut et al., 2016;
106 Ferrand et al., 2017; Aubry et al., 2018) have revealed that the displacement D is closely related to the
107 thickness a of fault veins.

108 Friction – until melting temperature is reached – and subsequent viscous shear in the molten layer
109 provide a heat source, which is counteracted by the latent heat of melting H , the thermal diffusion into
110 the country rock, depending on its heat capacity C_p , and advection with the loss of melt toward lateral
111 tensile cracks or zones of dilation (i.e. jogs or stepovers). In the following, we note w the average melt
112 volume per unit fault surface (including the injection veins, e.g. Di Toro et al. 2011) and a the residual
113 thickness after melt injects in the cracks of the damage zone. If such injection is more frequent for large
114 **pseudotachylytes** ($w > 1$ mm) than for small **pseudotachylytes** ($w < 1$ mm), then we may see a discrepancy
115 where $a < w$ for the former and $a \sim w$ for the latter (**Fig. 1a**).

116 In **Fig. 2** the log-log plot shows different log-linear trends and a break in slope ($D \approx 1$ mm, $a \approx 100$
117 μm) suggesting a change in thermal or hydraulic regime of the dynamic system. Except for Barker (2005)

118 and Di Toro et al. (2006), large-scale data, i.e. $D > 1$ mm and $a > 100$ μm , show a relatively close fit to
119 the power law described by Sibson (1975):

$$D = \beta \cdot a^2 \quad (1)$$

120
121 where $\beta = 43600 \text{ m}^{-1}$. On the other hand, for $D < 1$ mm and $a < 100$ μm , the data deviate from the log-
122 linear trend of **eq. (1)**. Note that the discrepancy between w and a is not sufficient to account for the bend
123 at $a \sim 1$ mm. For example, if the scaling observed at the sub-metric scale was valid at a larger scale, the
124 Balmuccia HP **pseudotachylyte** ($1.6 < D < 1.9$ m; $a \approx 5$ mm; Ferrand et al., 2018) should have an effective
125 thickness $w \approx 1$ m. There is no trace of such abundance of melt at the outcrop scale; in addition, according
126 to first-order calculations the magnitude associated to $w \approx 1$ m would be larger than 9.

127 The data from Sibson (1975) and most data from Barker (2005) are apparent offset only, since
128 displacement markers were not systematically identified on an exposed surface parallel to the slip
129 direction. Theoretically, an offset S corresponds to a true displacement $D = [S; \infty[$. The large datasets of
130 Nielsen et al. (2010b) and Di Toro et al. (2006) correspond to true displacement data, i.e. taking into
131 account the geometry of the fault and the observed finite offset. Regarding the fault network from
132 Scambelluri et al. (2017), the metagabbro sample was cut perpendicular to shear, which means that the
133 observed offsets are true displacements with a negligible error. Regarding the HP fault described by
134 Ferrand et al. (2018), the average apparent offset is ≈ 1.2 m for the main slip surface, which corresponds
135 to a displacement of 1.75 ± 0.15 m, i.e. $\approx 45\%$ larger than the measured offset. Considering that the rest of
136 the dataset (red crosses, **Fig. 2**) follows the same offset-displacement relationship, a small uncertainty is
137 taken into account (very small in log-log). For other field studies (Obata & Karato, 1995; Andersen et al.,
138 2014), only few displacement values are provided, which we associate with uncertainties of + 50%.

139 Only a few studies (Di Toro et al., 2005; 2006; Pittarello et al., 2008; Ferrand et al., 2018) report the
140 average thickness of **pseudotachylytes** measured over several meters of fault length. For the other studies,
141 estimates can still be cautiously considered, either regarding the large number of data collected on the

142 fault network (e.g. Nielsen et al., 2010b) or taking into account the internal consistency of the dataset (e.g.
143 Barker, 2005). 3D photogrammetry for instance, will definitely allow future studies on **pseudotachylytes**
144 to reach a sounder statistical approach, including the determination of the actual volume and 3D
145 distribution of **pseudotachylyte** networks. Regarding the study of Obata & Karato (1995), only the
146 minimum and maximum values of a range of observed thicknesses is provided, which we tentatively
147 associate in **Fig. 2** with the associated displacement estimate of ≈ 3 m. Rotary-shear experiments
148 constitute the most reproducible experimental dataset of seismic slip (e.g. Hirose & Shimamoto, 2005a;
149 2005b; Di Toro et al., 2006; Violay et al., 2015) and is for now the only technique reaching slip value typical
150 of natural earthquakes with $M_w > 4$ (> 2 cm). In particular, this technique evidences **dynamic friction values**
151 **approaching** zero during melt lubrication (Di Toro et al., 2004). However, in rotary-shear experiments, the
152 slip surface is not confined within a larger sample (D-DIA experiments for instance) or a jacket (stick-slip
153 experiments). Melt is therefore easily extruded, leading to much smaller final melt thickness (a^* ; **Fig. 2**;
154 **Table 2**). Nonetheless, some publications provide a record of the axial shortening during rotary-shear
155 experiments (Di Toro et al., 2006; Niemeijer et al., 2011), which actually relates to the cumulative melt (**+
156 clasts**) thickness (w^* ; **Fig. 2**; **Table 2**).

157 Regarding other experimental studies, the observations correspond to displacement as long as the
158 samples are cut perpendicular to the fault and the uncertainty mostly relies on the “single-jerk”
159 assumption, especially for the stick-slip design (**section 4.1**). Based on a limited number of FIB sections, D-
160 DIA studies consistently report values of some tens of nanometers. Values as small as 20 nm are associated
161 to local larger amorphous pockets along the fault, up to 200 nm thick (Ferrand et al., 2017).

162 **2.2. Scaling of **pseudotachylyte** thickness under the adiabatic assumption**

163 **Considering the rock volume enduring frictional heating as a closed system within a very short timescale,**
164 **the adiabatic assumption can be considered (Sibson & Toy, 2006; Rempel & Rice, 2006). For each**
165 **increment of rock on the rupture path, the duration of shear is extremely short (Heaton, 1990; Beroza &**

166 Mikumo, 1996), i.e. not the full slip timescale but the timescale of the initial shear and rupture. In other
 167 words, the local onset of efficient melt lubrication, which favors slip, actually corresponds to the local
 168 onset of cooling. Such short timescales also imply undrained conditions likely applying to the dynamic melt
 169 layer, at least for small events ($w < 1$ mm).

170 Assuming a Newtonian melt, negligible fracture energy and negligible viscoelastic effects (section
 171 4.2), the total shear heating Q results from the time integral of shear power (Nielsen, 2008):

$$Q = \int_0^t \tau(t') \cdot V(t') \cdot dt' = \int_0^t \tau(D') \cdot \frac{dD}{dt'} \cdot dt' = \int_0^D \tau \cdot (D') \cdot dD' = \bar{\tau} \cdot D \quad (2)$$

172 with $\bar{\tau}$ the average shear stress during the slip D at a velocity V . The heat Q can be partitioned into specific
 173 heat, latent heat for phase transitions (e.g. melt) and other heat sinks (e.g. diffusion, melt extraction).

174 Considering an adiabatic pseudotachylyte formation (Sibson 1975), all the heat Q_a generated in the
 175 process zone is dissipated within the molten layer on the rupture timescale and is directly related to its
 176 thickness w_a :

$$Q_a = \bar{\tau} \cdot D = \rho \cdot (C_p \cdot \Delta T + H) \cdot w_a \quad (3)$$

177 with H the latent heat of fusion, ρ the density, C_p the heat capacity and $\Delta T = (T_m - T_i)$ the process-zone
 178 heating during the dynamic rupture propagation (T_m : melting point; T_i : initial host-rock temperature). This
 179 implies that the thickness of the molten layer is a linear function of D as follows:

$$w_a = A_1 \cdot D \quad (4)$$

180 with $A_1 = \bar{\tau} / [\rho \cdot (C_p \cdot \Delta T + H)]$. As shown on Fig. 2, w_a strongly depends on $\bar{\tau}$. Assuming that $H = 0.3$
 181 MJ·K⁻¹, $\rho = 3000$ kg·m⁻³, $C_p = 1$ kJ·kg⁻¹·K⁻¹ and $\Delta T = 1000$ K then $A_1 \approx 3 \cdot 10^{-3}$ for $\bar{\tau} = 10$ MPa.

182 The adiabaticity is satisfied if w is larger than the heat diffusion distance within the duration of the slip
 183 ($w \gg \sqrt{\pi \cdot \kappa \cdot t}$). However, w is dynamically growing during the slip, and the ratio w/t is not necessarily
 184 constant throughout the spectrum of observed vein sizes.

185 Looking at the entire dataset, the adiabatic scaling reproduces the observations pretty well (**Fig.2**).
186 This is particularly true at small scale ($D < 1$ mm), because of small associated time scales (10 to 1000 μ s
187 only), during which sliding is likely to occur under quasi-adiabatic and undrained conditions.

188 On the contrary, at a larger scale ($D > 10$ mm), melt can advect and temperature diffuse, to such an
189 extent that the effective thickness w of the silicate melt layer increases less with sliding (Passelègue et al.,
190 2016b). **Effects** of both temperature diffusion and melt advection should depend on time and space scales,
191 respectively via thermal diffusivity and melt viscosity. **Notably, the concept of “seismic suction pump” was**
192 **proposed to explain near-fault drops of fluid pressure during rupture propagation (Sibson, 1987) and has**
193 **recently been confirmed in laboratory experiments producing stick-slip events systematically associated**
194 **with near-instantaneous drops in fluid pressure (Brantut et al., 2020). This effect has been extended to the**
195 **mobility of the rupture-induced transient melt observed at high pressure (Ferrand, 2017; Ferrand et al.,**
196 **2018).** The link observed by Sibson (1975) between D and a could be the result of a combination of the
197 dynamic rupture process, as observed in the laboratory (e.g. Passelègue et al., 2016a) and of melt sucking
198 in tension cracks (Ferrand, 2017). Thus, the deviation in the **pseudotachylyte** scaling law could be
199 interpreted as due to both scale-dependent **heat conduction** and melt pumping effects. Therefore, in
200 **section 3** we **remove** the adiabatic assumption.

201 **3. Pseudotachylyte formation with heat diffusion**

202 The bend observed in the scaling law can be attributed to the **effect** of melting dynamics under shear
203 heating at large scales. At first order, the dataset seems to follow the adiabatic trend (**eq. (4)**) at small
204 scales ($D < 1$ mm) and Sibson’s empirical law (**eq. (1)**) at larger scales (**Fig.2**). Separating the dataset
205 between low confining pressures (< 0.5 GPa; **Fig.3**) from high confining pressures (> 0.5 GPa; **Fig.4**), the
206 intersection/transition from **eq. (4)** and **eq. (1)** appears less well defined, for several reasons discussed

207 later (**section 4.1**). Nevertheless, at high pressures, linking experimental and field data clearly highlight a
208 discrepancy between $D < 1$ mm and $D > 1$ mm (**Fig.4**).

209 To better understand the apparent discrepancy between small-scale and large-scale observations, it is
210 important to come back to the definition of the seismic rupture, which is the dynamic migration of a
211 mechanical instability (e.g. Ohnaka, 2003; Passelègue et al., 2016a), thus implying a migration boundary
212 of the melt front. The full problem of melting coeval with heat diffusion requires solving a Stefan problem,
213 where the heat balance is defined at the advancing melt front (Carslaw & Jaeger, 1959; Nielsen et al.,
214 2008). In **section 3.1**, we develop the model, whose results in **section 3.2** show that the key is in the first
215 moments of the sliding and that Sibson's empirical law differs from the adiabatic model (**section 2.2**) for
216 large events in case of inefficient lubrication.

217 **3.1. Stefan problem: heat balance at the advancing melt front**

218 Following Nielsen et al. (2010b), a Crantz-Nicolson finite difference scheme is used to numerically solve
219 the 1-D diffusion equation with an advection term, coupled to the boundary condition:

$$\rho \cdot H \cdot v = \kappa \cdot \rho \cdot C_p \cdot \left(\left. \frac{\partial T}{\partial z} \right|_+ - \left. \frac{\partial T}{\partial z} \right|_- \right) \quad (5)$$

220 The +/- signs refer to the heat flow at the melting front toward the rock and toward the melt,
221 respectively, and v is the advancement velocity of the melting boundary, i.e. the boundary between melt
222 and country rock.

223 For simplicity we assume that density and diffusivity are the same in either melt or solid phases, and
224 constant. It can be shown (Nielsen et al. 2010) that:

$$\left. \frac{\partial T}{\partial z} \right|_- \approx \frac{\tau(t) \cdot V(t)}{2 \cdot \kappa \cdot \rho \cdot C_p} \quad (6)$$

225 resulting in:

$$\rho \cdot H \cdot v = \kappa \cdot \rho \cdot C_p \cdot \left. \frac{\partial T}{\partial z} \right|_+ + \frac{1}{2} \cdot \tau(t) \cdot V(t) \quad (7)$$

226 A nearly steady-state situation can be achieved after an initial transient, provided that $\tau \cdot V$ changes
 227 gradually. In such case $\left. \frac{\partial T}{\partial z} \right|_+ = \frac{-v}{\kappa} \cdot \Delta T$ (Nielsen et al., 2008) and **eq. (7)** reduces to:

$$v(t) = \frac{\tau(t) \cdot V(t)}{2 \cdot \rho \cdot (H + C_p \cdot \Delta T)} \quad (8)$$

228 Importantly, time integration of **eq. (8)** yields **eq. (4)**, indicating that the steady-state replicates the
 229 adiabatic situation. Indeed, the heat loss through diffusion is retrieved when the melting boundary
 230 advances, thus resulting in no net loss in steady-state. However, an initial transient can be predicted based
 231 on dimensional considerations and simple analytical solutions of the diffusion equation (Nielsen et al.,
 232 2008, eq. 24) where no steady-state has yet been reached:

$$D_c = 8 \cdot \frac{\tau}{V} \cdot \left(\frac{\rho \cdot (H + C_p \cdot \Delta T)}{\tau} \right)^2 \quad (9)$$

233 Substituting $\tau = 500$ MPa, $V = 1$ m/s, $\Delta T = 1000$ K and peridotite parameters in **eq. (9)** we obtain
 234 $D_c = 0.59$ mm, in agreement with $D_c \approx 0.1-1$ mm in **Fig. 4**.

235 **3.2. Modelling results**

236 For sake of clarity, solutions of the Stefan problem are given and compared to observational data
 237 separating low ($P < 0.5$ GPa) and high-pressure ($P > 0.5$ GPa) events, respectively in **Fig. 3** and **Fig. 4**. No
 238 steady-state is assumed in these models, where we use **eq. (7)** to obtain the velocity v , and a numerical
 239 solution of the diffusion equation to obtain the term $\left. \frac{\partial T}{\partial z} \right|_+$ (Nielsen et al., 2010b). The melt volume per unit
 240 fault surface is thereafter obtained by $w(t) = \int_0^t v(t') dt'$.

241 A time-dependent heat source $\tau(t) \cdot V(t)$ is imposed and the resulting $w(t)$ is compared to the
 242 observed **pseudotachylyte** thickness $w(D)$ (gray curves, **Fig. 3** and **Fig. 4**) under the assumption that
 243 $w(D) = w(V \cdot t)$; $V = 1$ m \cdot s $^{-1}$. We use values of H , ρ , C_p and κ which are broadly relevant for the
 244 variety of cases discussed here, but more closely related to peridotite.

245 We test the model using a trial heat which remains relatively high and approximately constant during
246 a very short initial slip stage where $\tau \cdot V = q_0$, subsequently switching to of the form $\tau \cdot V = q_0 \cdot \sqrt{\frac{D_c}{D}}$ (q_0
247 constant in $\text{J}\cdot\text{m}^{-2}\cdot\text{s}^{-1}$) in agreement with high-velocity weakening observed in rotary-shear experiments
248 (Nielsen et al., 2016) and with the fit initially proposed by Sibson (1975). To further support such choice
249 for $\tau \cdot V$, we note that high-velocity friction can be fitted (Nielsen et al. 2016) by $\tau \propto 1/\sqrt{D}$, resulting in
250 $\bar{\tau} \cdot D \propto \sqrt{D}$. For slip velocity during experiments and earthquakes, we may assume an indicative slip history
251 with an initial acceleration α of slip such that $V = \alpha \cdot t = \sqrt{2 \cdot D \cdot \alpha}$ up to a critical distance D_c , followed
252 by a modest variation such that $V \approx \text{Const} \approx \sqrt{2 \cdot \alpha \cdot D_c}$, resulting in an initial heat power $\tau \cdot V \propto \sqrt{2\alpha} \propto$
253 q_0 , followed by $\tau \cdot V \propto \sqrt{\frac{D_c}{D}}$.

254 The gray curves in **Fig. 3** and **Fig. 4** are obtained from modelling with $q_0 = 500 \cdot 10^6 \text{ J}\cdot\text{m}^{-2}\cdot\text{s}^{-1}$ (as an
255 indication, this value is compatible with $V = 1 \text{ m}\cdot\text{s}^{-1}$ and $\tau = \mu \cdot \sigma_n = 0.5 \times 1 \text{ GPa}$, and it is assumed that
256 the volume loss by injection is relatively limited therefore $a \approx w$). Three different values were modelled
257 for the onset of weakening: [1] $t_c = 7.55 \cdot 10^{-6} \text{ s}$ (i.e. $D_c = 2.88 \cdot 10^{-5} \text{ m}$; corresponding to the onset of
258 melting at $T = T_m$), [2] $t_c = 10^{-3} \text{ s}$ and [3] $t_c = \infty$ (the latter case results in a constant power q_0
259 throughout the slip; arguably, this may account for absence of weakening due to high viscosity of colder
260 or more felsic melt, to match the datapoints in the higher part of the diagram).

261 In all three cases the initial thickness is relatively low, in agreement with D-DIA datapoints, but rapidly
262 increases because the melting front is still in the initial transient stage. At about $D \approx 0.1 - 1 \text{ mm}$ ($a \approx$
263 10^{-4} m) a quasi-steady-state is reached where thickness increases as $\propto \sqrt{D}$ in agreement with the
264 imposed power which decays in $\approx 1/\sqrt{D}$ for curves [1] and [2], therefore closely matching the steady-
265 state/adiabatic approximation of **eq. (6)**. For case [3] the thickness increases as $\propto D$ again in agreement
266 with the steady-state approximation of **eq. (6)** but assuming that the power is roughly constant throughout
267 the slip.

268 Once the steady-state thickness is reached ($D > D_c$), we retrieve the trend compatible with Sibson's
269 observation such that $w_{ss} = A_2 \cdot D^{1/2}$ with $A_2 = \sqrt{1/\beta} \approx 5 \cdot 10^{-3} \text{ m}^{1/2}$, in agreement with stress decay $\bar{\tau} \propto$
270 $1/\sqrt{D}$.

271 As mentioned above, Barker (2005) and Di Toro et al. (2006) reported large-scale **pseudotachylytes** that
272 do not follow Sibson's law, but rather a linear relationship between D and a , suggesting an adiabatic
273 process and inefficient melt lubrication. The data of Barker (2005) indicate $\bar{\tau} > 100 \text{ MPa}$ when the data
274 of Di Toro et al. (2006) would indicate $\bar{\tau} \approx 10 \text{ MPa}$ (**Fig. 2**). Contrary to all other studies focusing on
275 isotropic rocks, Barker (2005) describes **pseudotachylytes** in a quartzo-feldspathic schist, suggesting that
276 the bend is host rock-dependent. The parameters controlling melting point and magma viscosity, and how
277 the latter influence our modelling results, are discussed in **section 4.2**. In any case, according to our
278 modelling results, Sibson's empirical law actually corresponds to efficient melt lubrication case, while
279 scaling relations closer to the adiabatic (linear) trend correspond to events where weakening was not
280 significant.

281 **4. Discussion**

282 The results presented in **section 3.2**, if carefully considered in light of limitations of both the dataset
283 (**section 4.1**) and the model (**section 4.2**), might yield a tool to estimate the magnitude of an earthquake
284 by simply looking at the thickness of the resulting pseudotachylyte (**section 4.3**).

285 **4.1. Limitations of the dataset**

286 Displacement-thickness or offset-thickness datasets are presented along with associated uncertainties
287 (**section 2; Fig.2**). **Here we detail, separately, the factors likely to control displacement and thickness.**

288 Displacement estimates depend on the experimental design: for the D-DIA experiments, several lines
289 of evidence suggest that each transformation-induced fault records a single seismic event (Schubnel et al.,
290 2013; Ferrand et al., 2017; Incel et al., 2017), which is **uncertain** for stick-slip experiments that give an

291 estimate at the scale of the saw-cut after seismic events (Passelègue et al., 2016a; Brantut et al., 2016;
292 Aubry et al., 2018). This could actually explain why displacements are up to one order of magnitude larger
293 in stick-slip experiments compared to faults with the same thickness in the D-DIA (**Fig. 2**), **even though**
294 **differences in machine stiffness may also be at play**. Considering multiple **pseudotachylyte** generation on
295 the same fault plane, a significant increase of D/a **would be observed**. However, publications highlight the
296 high strength of faults welded by **pseudotachylyte** (Di Toro & Pennacchioni, 2005; Mitchell et al., 2016;
297 Proctor & Lockner, 2016; **Hayward & Cox, 2017**). Thus, new ruptures are not expected to affect preexisting
298 **pseudotachylytes** except for stick-slip experiments or **in case of** alteration. In nature, **pseudotachylytes**
299 may also form from shear zones exhibiting a significant displacement before rupture nucleation (e.g.
300 Sibson, 1980; Chattopadhyay et al., 2008; Pittarello et al., 2012, John et al., 2009), but no evidence of it
301 has been reported in the reviewed studies (**Fig. 2**). Notably, either **local** mylonites or cataclasites on the
302 side of a **pseudotachylyte** network can be **coeval with** the **pseudotachylyte** (e.g. horse-tail termination,
303 Ferrand et al., 2018; cataclastic damage zone, **Petley-Ragan et al., 2018**). Such coeval evolution highlights
304 the key role of the strain rate (**variable in time and space**) and has been reported in both field and
305 experimental studies (e.g. Fabbri et al., 2000; Kim et al., 2010).

306 The thickness a measured in the field may vary over more than one order of magnitude along strike,
307 either due to dilational and contractional jogs or because of melt flux into the damage zone (e.g. Sibson,
308 1975; Ferrand et al., 2018). Field studies show that **pseudotachylytes** might be present as discontinuous
309 patches along a fault, **with** melting **only** along contractional jogs (Griffith et al., 2009; Kirkpatrick & Shipton,
310 2009). Experimental data also show significant variation of the **pseudotachylyte** thickness (e.g. **Hayward &**
311 **Cox, 2017**). For stick-slip experiments, only **cumulative** average values are available. For D-DIA
312 experiments, TEM **on FIB sections** is required to measure the fault thickness. In rotary-shear experiments,
313 the sample is only **meant to be** an equivalent of a limited fraction of a fault, while other techniques may
314 encompass complete fractures, **i.e. ruptures area smaller than the sample size**. In addition, the

315 experimental fault segment is not confined, which induces fast extrusion of the melt out of the fault plane.
316 Finally, rotary-shear experiments do not reproduce dynamic ruptures but slip generated artificially by a
317 motor, **although it provides** the dynamic molten layer extrusion (**section 2.1**). For normal stress larger than
318 20 MPa and abrupt accelerations to target slip rates of $3 \text{ m}\cdot\text{s}^{-1}$, **rotary-shear** data are well fitted by the
319 model presented in **section 3 (Fig. 2; Niemeijer et al., 2011)**.

320 **4.2. Limitations in the physical model**

321 The occurrence and efficiency of melt lubrication depend on the melting point and melt viscosity, which
322 are not explicitly taken into account in our model (**section 3**). However, these key factors are implicitly
323 contained within the characteristic time t_c and associated distance D_c (onset of weakening). Studies
324 reporting **pseudotachylyte** thickness and displacement data investigate various lithologies and contexts
325 (**Fig. 1, Fig. 2, Table 1**), but the same physics should apply. In the same conditions of pressure, temperature,
326 stress and strain rate, similar materials rupture in the same way. Hereafter we discuss the **influence** of
327 rock composition, pressure, temperature **and strain rate** on the melting point and melt viscosity.

328 Melt viscosity is linked to rock chemistry and temperature (e.g. Taniguchi, 1992; Spray, 1993; Suzuki,
329 2001). Between 1500 and 2000°C at 1 atm, ultramafic melts are at least one order of magnitude **less**
330 viscous than crustal melts with higher silica contents (Holtz et al., 1999; Karki et al., 2013). At 1800°C, a
331 melt of MgO or Mg_2SiO_4 composition exhibits a viscosity six orders of magnitude lower **than SiO_2** (Karki et
332 al., 2013). Melt viscosity is primarily governed by SiO_2 tetrahedra polymerization, **strongly limited** by cation
333 diversity (e.g. Whittington et al., 2000; Le Losq et al. 2013; Toplis & Dingwell, 2004).

334 The viscosity of silicate melts decreases with pressure (e.g. Gupta, 1987; Xue et al., 1991; Bottinga &
335 Richet, 1995; Suzuki et al., 2002; Liebske et al., 2005) due to a change in silicon coordination (Xue et al.,
336 1991). The viscosity of dacite or albite melts drops by one order of magnitude when the pressure increases
337 by 2 GPa (Tinker et al., 2004). These factors contribute to a drastic reduction of the melt viscosity at depth.
338 **Peridotitic** melts may have a viscosity as low as 1 Pa.s at ambient pressure (Dingwell et al., 2004), which

339 drops to $\sim 10^{-1}$ Pa.s between 1 and 3 GPa (Suzuki et al., 2001; Liebske et al., 2005), equalling diopside
340 viscosity (Taniguchi, 1992), whereas higher SiO₂ contents (e.g. andesite composition) exhibit viscosities of
341 ~ 10 Pa.s (Vetere et al., 2006).

342 At high slip rates, displacement is accommodated over finite distances with extremely high strain rates,
343 i.e. far above the conditions for viscoelasticity. Whether a melt behaves as a solid or a liquid depends on
344 its relaxation time (Dingwell & Webb, 1989; Hayward et al. 2019). Most geological materials have non-
345 Newtonian viscosities, especially strain-rate dependent. For seismic slip and at temperatures above the
346 melting temperature, this could result in viscosities being many orders of magnitude less than standard
347 laboratory measurements (Webb & Dingwell, 1990). In addition, a moderate solid fraction may increase
348 the viscosity of the solid+melt suspension by one order of magnitude (Costa, 2005; Ferrand et al., 2018).
349 The solid fraction is likely to be high in rocks made of minerals with very different melting point (e.g.
350 tonalite; Di Toro et al., 2006), whereas peridotites likely endure congruent melting (Ferrand et al., 2018).
351 Furthermore, exothermic crystallization (Burbank, 1936; Blundy et al., 2006) may inhibit the viscosity
352 decay.

353 Water, either aqueous or structurally bound, is key to activating weakening processes (e.g. Mei &
354 Kohlstedt, 2000a; 2000b; Violay et al., 2014). Dehydration reactions induce dynamic weakening that favors
355 strain localization and seismic sliding (e.g. Brantut et al., 2008). Conversely, during sliding-induced
356 antigorite dehydration, a pressurized low-viscosity silicate melt leads to extremely efficient lubrication
357 (Brantut et al., 2016). In small amounts, water lowers the melting point of peridotite by ~ 100 °C, inducing
358 olivine melting around 1300 °C (San Carlos, 0.32 wt.% H₂O; Gaetani & Grove, 1998). Water also drastically
359 lowers melt viscosity (Kushiro et al., 1976; Richet et al., 1996; Vetere et al., 2006). As a consequence,
360 tension cracks forming in the dilatant quadrant as rupture propagates (e.g. Okubo et al., 2019) generate
361 transient depressions able to suck the melt out of the fault core (Ferrand et al., 2018). Inasmuch as melt
362 viscosity is sufficiently low, the melt flux toward these cracks may also represent an important heat sink.

363 The viscosity in the fault vein is expected to increase due to both extrusion (fault thinning) and enhanced
364 quench (solidification). The suction effect dynamically migrates as long as there is still enough lubricant to
365 propagate the rupture. Alternative lubrication mechanisms, such as thermal pressurization (Rempel &
366 Rice, 2006; Viesca & Garagash, 2015) could play a significant role if melting conditions are not fulfilled.

367 As mentioned in **section 3**, the study of Barker (2005) reported a pseudotachylytes population with D
368 and a consistent with adiabatic formation (**section 2.3, Fig.2**), indicating higher values of average shear
369 stress $\bar{\tau} > 100$ MPa. According to **eq. (3)**, this could also be due to a lower ρ , a lower C_p or a lower ΔT
370 (lower melting point). The data of Di Toro et al. (2006) are also consistent with the adiabatic assumption,
371 which suggests that quartzo-feldspathic rocks would not achieve efficient weakening after the onset of
372 melt-assisted lubrication. Feldspar has a lower melting point ($\sim 1500^\circ\text{C}$; Taniguchi, 1992; Suzuki et al.,
373 2002) than the other minerals constituting the rocks of the dataset, but the high Si/Mg ratio maintains a
374 relatively high melt viscosity at low pressure (< 0.5 GPa; Suzuki et al., 2001).

375 The Woodroffe Thrust fault zone (Australia, e.g. Camacho et al., 1995) appears as a counterexample
376 for the scaling law proposed in this study. The fault zone is ≈ 1 km thick and contains ≈ 4 % of
377 pseudotachylyte veining. Large volumes of pseudotachylytes formed in anhydrous felsic granulites, but in
378 significantly smaller volumes in underlying amphibolite-facies rocks (Camacho et al., 1995). The high Si
379 content and the anhydrous conditions probably maintained a high melt viscosity, which means that a
380 larger melt layer is required for a given amount of slip to account for the same stress drop. In addition, the
381 zone is characterized by numerous seismic events, some of which potentially affecting the same fault
382 segment, consistently with repeating earthquakes inferred by seismologists (Uchida & Bürgmann, 2019).

383 **4.3 Implication: Moment magnitude vs pseudotachylyte thickness**

384 Keeping in mind the limitations detailed above, we can try a tentative scaling of the pseudotachylyte
385 thickness with the earthquake magnitude. The stress drop $\Delta\tau$ during an earthquake is controlled by the
386 fault geometry such that (Kanamori, 1977):

$$\Delta\tau = C \cdot \mu \cdot D/L \quad (10)$$

387 with L is the fault length and μ is the shear modulus, e.g. $\sim 8 \cdot 10^{10}$ Pa for olivine, and C is a geometrical
388 constant ≈ 1 . The moment magnitude M_0 is defined as:

$$M_0 = \mu \cdot D \cdot L^2 \quad (11)$$

389 Combining **(10)** and **(11)**, the displacement D can be expressed as a function of M_0 as follows:

$$D = \left(\frac{\Delta\tau}{C}\right)^{2/3} \cdot \frac{M_0^{1/3}}{\mu} \quad (12)$$

390 As a consequence, for an adiabatic process, w_{adiab} varies as $\sqrt[3]{M_0}$:

$$w_{adiab} = \left(\frac{\Delta\tau}{C}\right)^{2/3} \cdot \frac{M_0^{1/3}}{\mu} \cdot A_1 \quad (13)$$

391 with $A_1 = \bar{\tau}/[\rho \cdot (C_p \cdot \Delta T + H)]$, the thickness w_{adiab} can be expressed as a function of $\Delta\sigma$, $\bar{\tau}$ and M_0 :

$$w_{adiab} = \left(\frac{\Delta\tau}{C}\right)^{2/3} \cdot \frac{M_0^{1/3} \cdot \bar{\tau}}{\rho \cdot (C_p \cdot \Delta T + H) \cdot \mu} \quad (14)$$

392

393 Considering large-scale steady-state, w_{ss} varies as $\sqrt[6]{M_0}$:

$$w_{ss} = \left(\frac{\Delta\tau}{C}\right)^{1/3} \cdot \frac{M_0^{1/6}}{\mu^{1/2}} \cdot A_2 \quad (15)$$

394 To summarize this subsection, w_{adiab} and w_{ss} vary as $\sqrt[3]{M_0}$ and $\sqrt[6]{M_0}$ respectively. In other words,
395 the **pseudotachylyte** thickness saturates for large earthquakes (**Fig. 5**) and cannot be much larger than 5
396 cm, which corresponds to a M_w 9, i.e. tens of meters of slip. (M_w 9). Hence, larger pseudotachylyte veins
397 found on the field should be considered as “injectites” (tension cracks), local pull-apart structures or the
398 result of successive events.

399 We define the crossover moment M_0^c as the intersection between the calculations considering
400 adiabatic and non-adiabatic processes, that can be written:

$$M_0^c = \left(\frac{C}{\Delta\tau}\right)^2 \cdot \left(\frac{\rho \cdot (C_p \cdot \Delta T + H) \cdot \mu}{\bar{\tau}}\right)^6 \cdot A_2^3 \quad (16)$$

401

402 Keeping the same assumptions as for **eq. (9)**, the crossover seismic moment M_0^c would be around
403 $5 \cdot 10^{13}$ to 10^{14} , i.e. $M_w \approx 3$, mostly depending on $\bar{\tau}$ and $\Delta\tau$ (**Fig. 6**). When $M_0 \leq M_0^c$ measuring w could
404 provide a reasonable estimate of M_w , but for larger events efficient melt lubrication would induce

405 significant uncertainties, with fault veins thicknesses of some millimeters potentially corresponding to
406 magnitudes ranging from 2 to 7. This implies that field observations of pseudotachylyte thickness should
407 be used carefully in the establishment of energy balances. The total melt volume produced per surface
408 unit appears much more adequate, which requires a thorough investigation at the outcrop scale to
409 estimate the melt amount intruded within the damage zone. In the field, as large pseudotachylyte
410 exposures are necessarily truncated for large events, it is only possible to give minimum estimates of either
411 D or M_w . Nonetheless, carefully considering all the parameters detailed in this study, notably
412 compositional considerations and P-T conditions, some estimate could be made. Furthermore, in the
413 absence of displacement marker at a large pseudotachylyte fault vein ($a > 1$ mm), studying the damage
414 zone could give some idea of the size of the main rupture surface. Our model could be used to upscale
415 small-scale observations for either adiabatic or diabatic ruptures.

416 **4.3 Implication: energy balance**

417 A recent study showed that the energy dissipated by fracturation during the propagation of the seismic
418 rupture depends on the size of the asperities of the fault surface, following a power law (Passelègue et al.,
419 2016b). The nucleation length directly depends on the fracture energy, which means that it depends on
420 the asperity size (Ohnaka, 2003; Passelègue et al., 2016b). The energy balance of fault slip would suggest
421 that fracture energy G dissipates in fault weakening due to coseismic processes (including melting) that
422 would reduce the heating. Seismological inferences of G do show that in itself it is an increasing function
423 of slip (e.g. Rice, 2006; Viesca & Garagash, 2015).

424 Interestingly, the scaling relationship of G with D also shows a change of slope at $D \sim 1$ mm, varying
425 as D^2 for $D < 1$ mm and as $D^{2/3}$ for $D > 1$ mm (Viesca & Garagash, 2015). Furthermore, D varies as a^2 for D
426 > 1 mm, thus G varies as $a^{4/3}$; and for $D < 1$ mm D may actually vary as $a^{2/3}$, which would mean that G varies
427 as $a^{4/3}$ as well. In other words, the pseudotachylyte thickness would scale with the fracture energy with no
428 impact of scale.

429 **5. Conclusions**

430 Pseudotachylyte thickness scales with slip over six orders of magnitude for measured slip varying from
431 microns to meters, **which can be carefully estimated in the field even in the absence of displacement**
432 **markers**. As a consequence, the thickness scales with moment M_0 and magnitude M_w . Experimental and
433 natural faults show striking similarities. Regardless of design, the scaling law follows the same trend. The
434 bend observed in the scaling law is interpreted as the signature of **pseudotachylyte** thickness saturation
435 for large M_w due to scale-dependent efficiency of melt-assisted lubrication and possibly to variations in
436 viscosity and mobility of the **pseudotachylyte** melt suspension infill. The thickness of the molten layer
437 cannot be much larger than 5 cm, which corresponds to a M_w 9, i.e. tens of meters of slip. Hence, larger
438 reported pseudotachylyte veins most likely correspond to injection veins or local **dilatational** jogs along
439 longer and thinner faults.

440 **6. Acknowledgments**

441 We thank Sarah Incel for providing the SEM image of the dehydrated lawsonite blueschist. This research
442 is supported by the *International Research Promotion Office of the Earthquake Research Institute* and the
443 European Research Council, grant no. 681346-REALISM. All datasets used in this paper have already been
444 used by previous studies and are fully presented in the tables.

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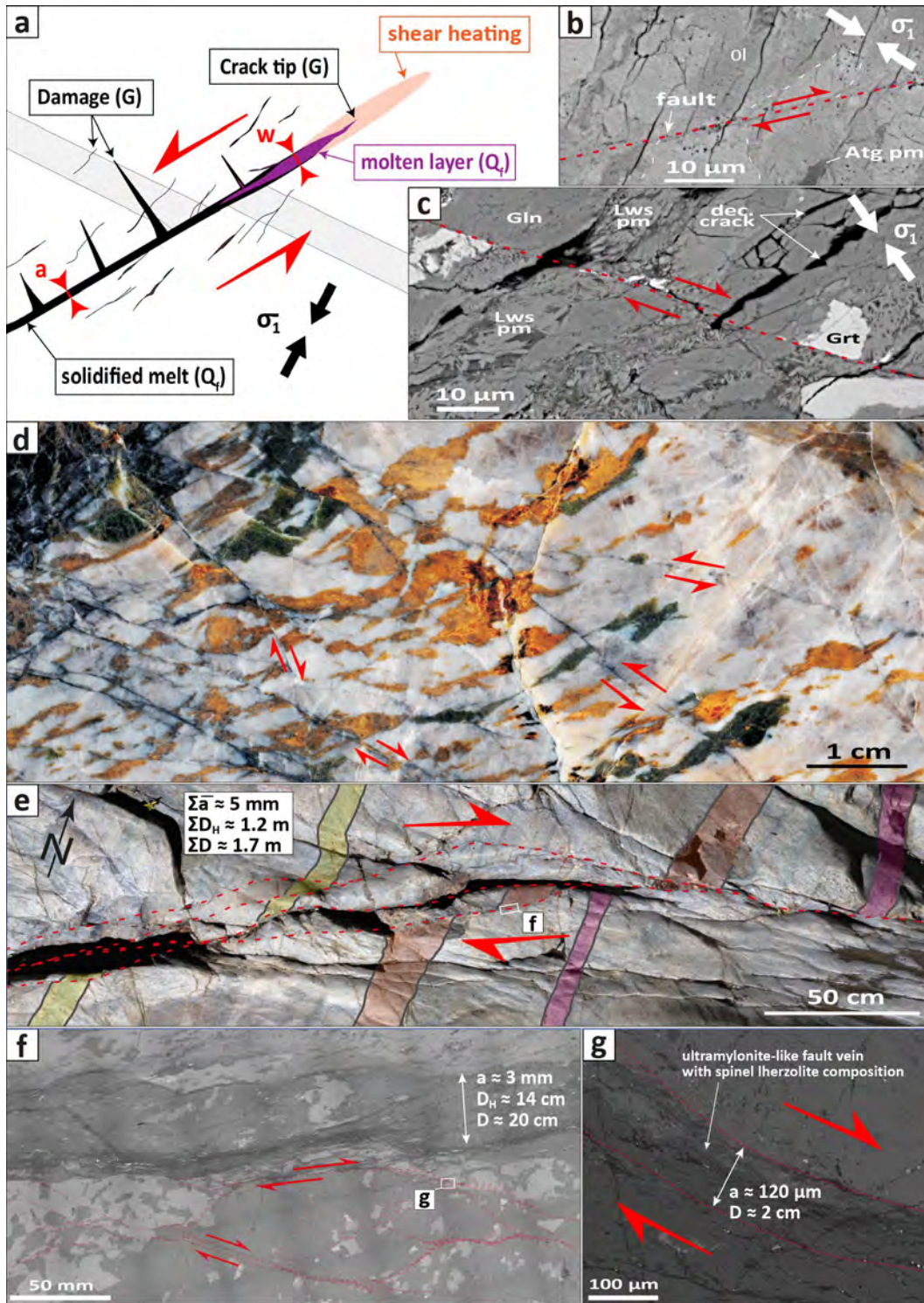
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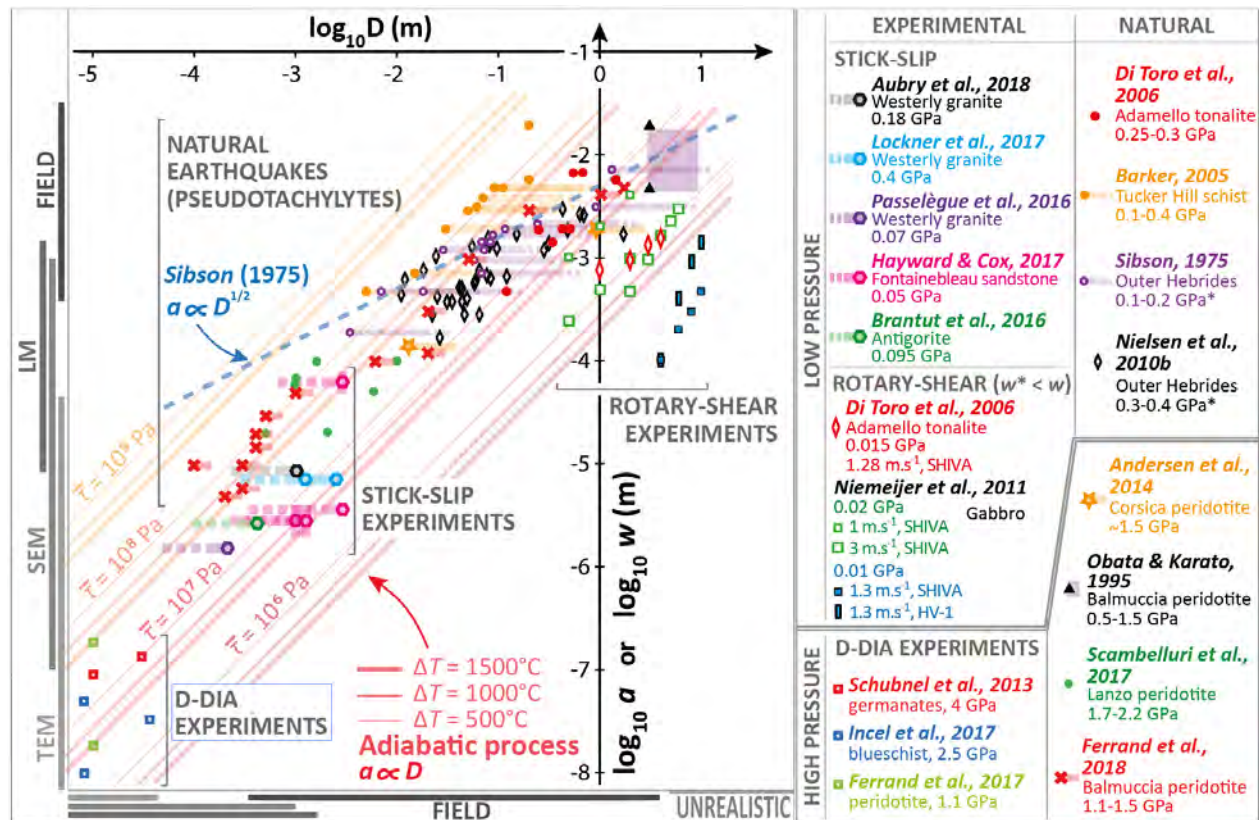
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724 **Figure 1: Laboratory and field observations of pseudotachylytes. a:** Sketch of pseudotachylyte formation

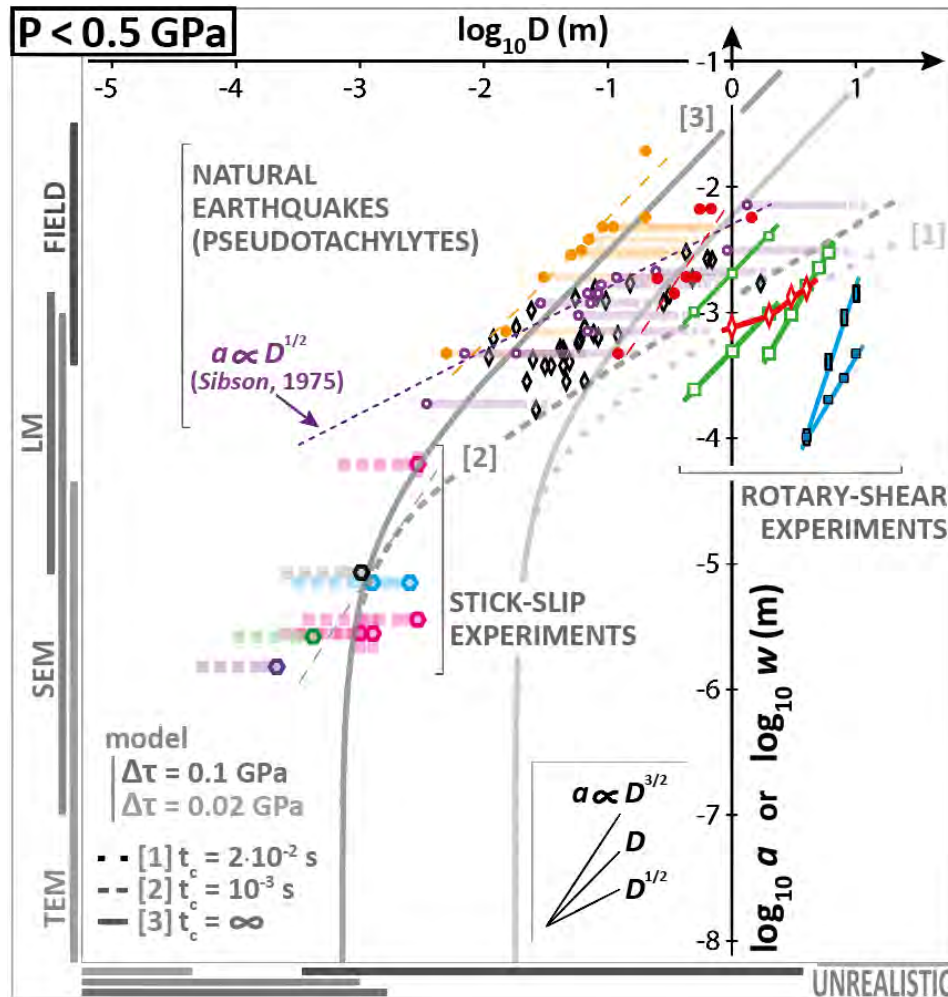
725 within a rock containing a displacement marker (energy partitioning between heat Q and fracture energy

726 G ; distinction between final thickness a and dynamic melt width w . Arrows show the maximum
727 compression direction (σ_1 ; black) and the sense of shear (red). **b**: Experimental pseudotachylyte in olivine
728 due to antigorite dehydration ($D = 10 \mu\text{m}$; $a = 20\text{-}200 \text{ nm}$; Ferrand et al., 2017). White dots define an
729 intensely sheared olivine grain. **c**: Experimental pseudotachylyte in a blueschist due to lawsonite
730 dehydration ($D = 36 \mu\text{m}$; $a = 10\text{-}50 \text{ nm}$). The displacement marker is a garnet grain (Grt). Lawsonite
731 pseudomorphs (Lws pm) consist of epidote needles and omphacite. Gln = glaucophane; dec. =
732 decompression. **d**: Oceanic mylonitic gabbro (Scambelluri et al., 2017) made of green clinopyroxene
733 (green), olivine (orange) and plagioclase (white). Black thin lines are the network of pseudotachylyte fault
734 veins cutting the mylonitic foliation ($D = 1 \text{ cm}$ for $a = 100 \mu\text{m}$; $D = 500 \mu\text{m}$ for $a = 20 \text{ nm}$). **e**: Natural
735 high-pressure fault found in the Balmuccia peridotite (subvertical; $D = 1.7 \pm 0.2 \text{ m}$; $a = 5 \text{ mm}$; $w < 2 \text{ cm}$
736 Ferrand et al., 2018). **f**: Closer view of (e) showing one of the main fault veins ($D = 20 \text{ cm}$; $a = 3 \text{ mm}$). **g**:
737 Network of ultramylonite-like veins highlighted in (f) ($D = 2 \text{ cm}$; $a = 120 \mu\text{m}$). See the associated **Table 1**
738 for details and additional references.

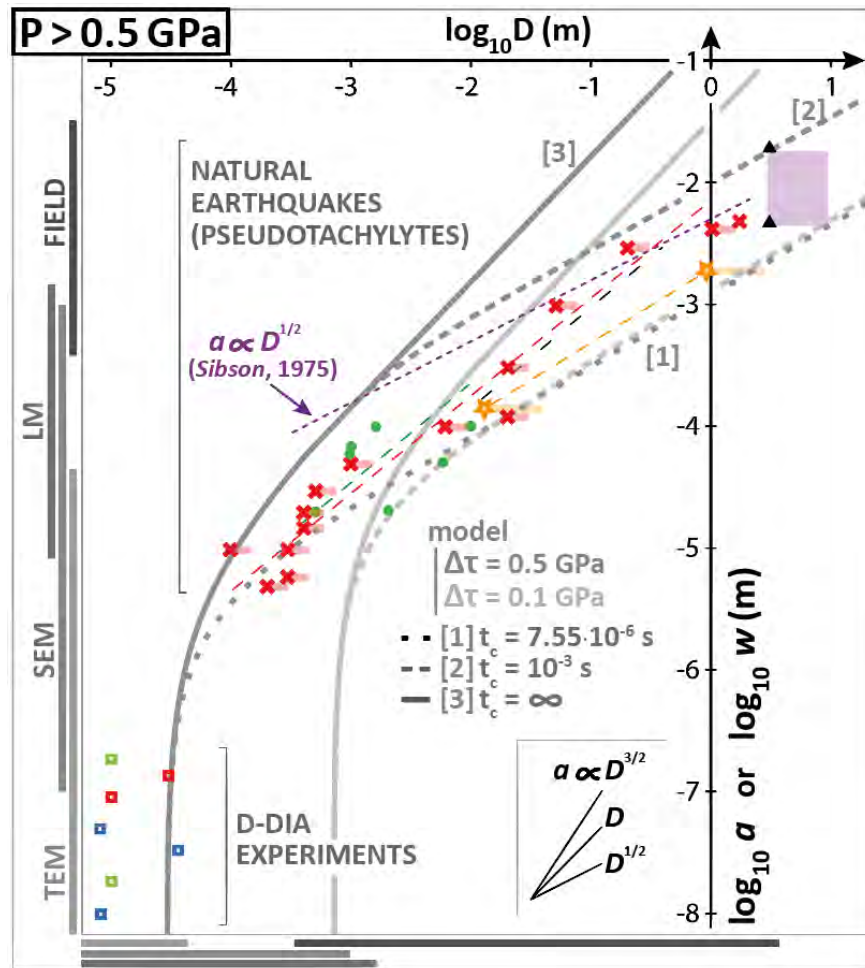


740 **Figure 2: Displacement as a function of fault thickness: datasets and adiabatic assumption.** Observations
741 from field work (1 mm – 10 m) and various microscopy techniques (10 nm – 1 mm). Dataset come from a
742 review of published studies (Table 1). Field-scale data are compared to micrometer-scale and nanometer-
743 scale observations. A distinction is made between the pseudotachylyte thickness a (fossilized trace of the
744 seismic rupture, i.e. observational data) and the dynamic thickness w (volume of melt produced by surface
745 unit, i.e. simple adiabatic model). Values of a and w should slightly differ considering off fault damage and
746 melt intrusion (section 2.2). All the faults are reported as pseudotachylyte, except the ones of Schubnel et
747 al. (2013; “transformational faulting” during olivine-spinel transition). Data from Scambelluri et al. (2017)
748 are from the fault network shown in Fig. 1. The datasets of Barker (2005) and Sibson (1975) are minimum
749 estimates of D , which could be one or two orders of magnitude larger (see discussion in section 2.1).
750 Rotary-shear experiments provide two distinct datasets: sample axial shortening w^* (cumulative melt
751 thickness) and the final pseudotachylyte thickness a^* (Hirose & Shimamoto, 2005; Del Gaudio et al., 2009)

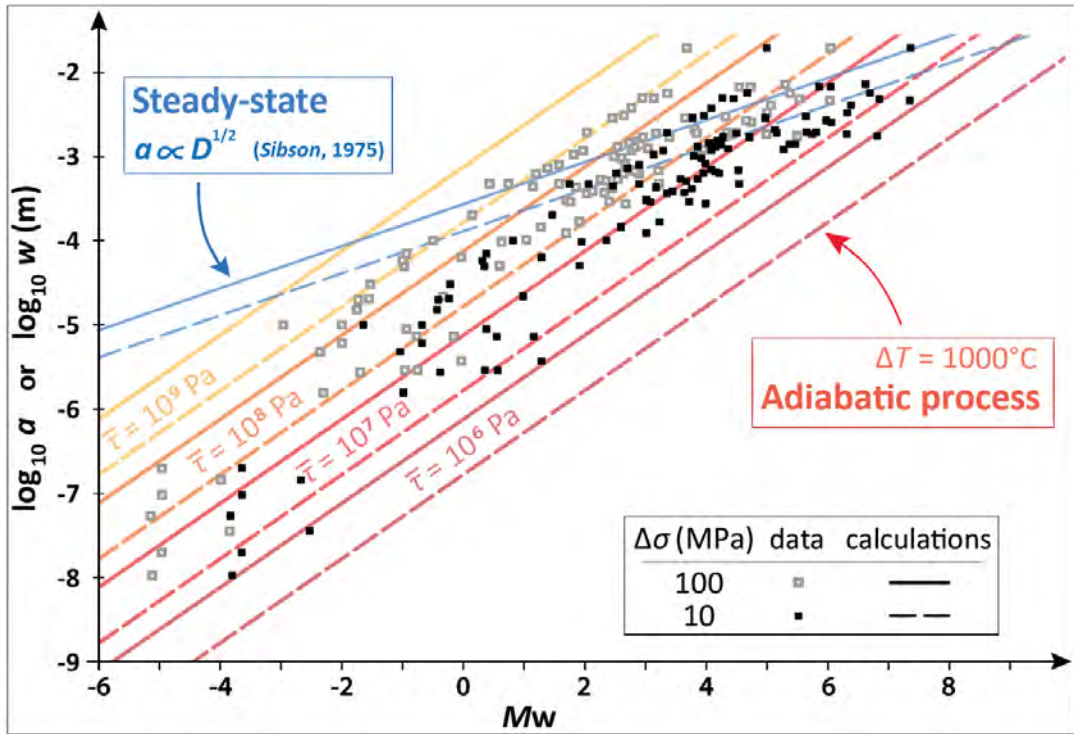
752 that is not considered here ([section 2.1](#)). [Gray sidebars](#) recall the limits of Light Microscopy (LM) and
 753 Scanning (SEM) and Transmission Electron Microscopy (TEM). Calculations [provided in section 2.2](#). The
 754 density is considered to be equal to $2900 \text{ kg}\cdot\text{m}^{-3}$, and its variability does not have much influence on the
 755 results). Uncertainties about the dynamic temperature rise ΔT are considered. The range of average shear
 756 stress $\bar{\tau}$ is consistent with typical seismological estimates. See **Table 1** for information about the datasets.



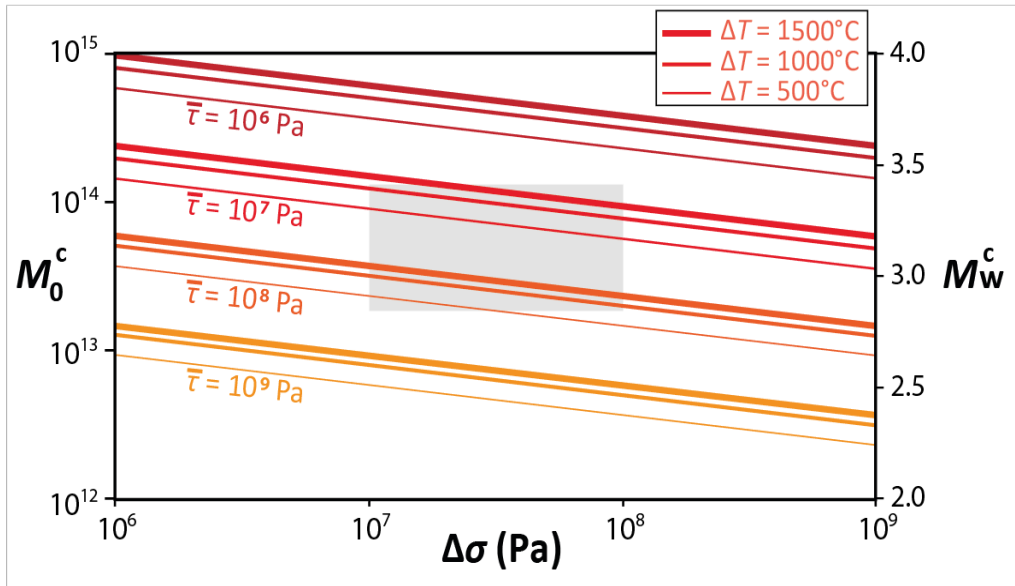
758 **Figure 3: Scaling seismic fault thickness with displacement as a Stefan problem (low pressures).** Gray
 759 curves show the results of the model (see [section 3](#)) considering a peridotite and a constant sliding velocity
 760 $V = 1 \text{ m}\cdot\text{s}^{-1}$ for [1] the onset of weakening ($t_c = 2 \cdot 10^{-2} \text{ s}$), [2] $t_c = 10^{-3} \text{ s}$ and [3] $t_c = \infty$ (constant
 761 power). The calculation is made for values of $\Delta\tau$ relevant for the pressure conditions. The dataset is
 762 presented in **Fig.2**, **Table 1** and **Table2**.



764 **Figure 4: Scaling seismic fault thickness with displacement as a Stefan problem (high pressures).** Gray
 765 curves show the results of the model (see **section 3**) considering a peridotite and a constant sliding velocity
 766 $V = 1 \text{ m}\cdot\text{s}^{-1}$ for [1] the onset of weakening ($t_c = 7.55 \cdot 10^{-6} \text{ s}$), [2] $t_c = 10^{-3} \text{ s}$ and [3] $t_c = \infty$ (constant
 767 power). The calculation is made for values of $\Delta\tau$ relevant for the pressure conditions. The dataset is
 768 presented in **Fig.2** and **Table 1**.



770 **Figure 5: Fault thickness and direct magnitude estimates.** a: pseudotachylyte thickness a (data) and
 771 dynamic molten zone width w (calculations) vs. earthquake magnitude M_w , with $M_w =$
 772 $2/3 \log_{10} M_0 - 6.07$ (section 4.3) and M_0 from eq. (12). Warm colors correspond to the adiabatic process
 773 (eq. 15) whereas the blue color is used for the diabatic process (Sibson's law), which actually corresponds
 774 to the steady-state value of the Stefan problem described in this study (eq. 16).



776 **Figure 6: Crossover moment M_0^c and moment magnitude M_w^c .**
 777 M_0^{co} as function of $\bar{\tau}$ and $\Delta\sigma$ showing limited influence of ΔT . The gray box indicates the range of values
 778 consistent with observation, as discussed in **section 3.3** of the main manuscript.

Table 1 – Dataset collected from various experimental and field works and presented in Fig. 2.

<i>a</i>	<i>D</i>	Material	Reference
D-DIA experiments			
200 nm	10 μ m	Olivine-antigorite aggregates	Ferrand et al. (2017)
20 nm	10 μ m		
10 nm	8 μ m	Lawsonite blueschist	Incel et al. (2017)
50 nm	8 μ m		
30 nm	6 μ m		
100 nm	10 μ m	Ge-olivine	Schubnel et al. (2013)
150 nm	30 μ m		
Stick-slip experiments			
1.5 μ m	250 μ m	Westerly granite	Passelègue et al. (2016a)
8 μ m	1.2 mm		Aubry et al. (2018, supp. Mat.)
7 μ m	1.2 mm		Lockner et al. (2017)
7 μ m	4.2 mm		
2.5 μ m	500 μ m	Alpine Corsica antigorite	Brantut et al. (2016)
2-4 μ m	1.35 mm	Fontainebleau sandstone, > 99% quartz	Hayward & Cox (2017)
2-4 μ m	1.09 mm		
3-4 μ m	2.28 mm		
50-80 μ m	2.28 mm		
Field work			
50 μ m	6 mm	Lanzo peridotite, Italian Alps	Scambelluri et al. (2017)
20 μ m	2 mm		
70 μ m	1 mm		
100 μ m	1.6 mm		
100 μ m	10 mm		
20 μ m	500 μ m		
60 μ m	1 mm		
4 mm	1 m	Balmuccia peridotite, Italian Alps	Ferrand et al. (2018)
3 mm	20 cm		
12 μ m	2 cm		
5 mm	1.7 m		
100 μ m	6 mm		
300 μ m	2 cm		
1 mm	5 cm		
10 μ m	100 μ m		
5 μ m	200 μ m		
6 μ m	300 μ m		
20 μ m	400 μ m		

30 μm	500 μm				
10 μm	300 μm				
6.9 mm	54 cm	Adamello tonalite, Italian Alps	Di Toro et al. (2006)		
2 mm	42.5 cm				
1.5 mm	34 cm				
0.5 mm	12 cm				
7 mm	66 cm				
2 mm	50 cm				
2 mm	25 cm				
5.9 mm	1.44 m				
150 μm	2 mm			Alpine Corsica peridotite	Andersen et al. (2014)
1.29 cm	90 cm				
5 mm	3 m	Balmuccia peridotite	Obata & Karato (1995)		
2 cm	3 m				
0.5 mm	5 mm	Tucker Hill quartzo-feldspathic schist, Central Otago, New Zealand	Barker (2005)		
750 μm	1.5 cm				
2 mm	3 cm				
3 mm	5 cm				
3.25 mm	6 cm				
4 mm	7 cm				
5 mm	9 cm				
5 mm	11 cm				
6 mm	14 cm				
2 cm	20 cm				
500 μm	7 mm	Outer Hebrides gneiss, Scotland	Sibson (1975)		
200 μm	3.4 mm				
1.25 mm	2.8 cm				
500 μm	1.8 cm				
1.5 mm	6.7 cm				
1.75 mm	8.8 cm				
1.5 mm	8.2 cm				
1.25 mm	7.1 cm				
1 mm	5.8 cm				
2 mm	11.7 cm				
750 μm	6.8 cm				
2.25 mm	24,3 cm				
7.5 mm	1.29 m				
3.25 mm	91 cm				
300 μm	2.2 cm		Nielsen et al. (2010b)		
1.1 mm	2.4 cm				
650 μm	1.2 cm				
1.35 mm	5.5 cm				
400 μm	3.1 cm				

180 μm	2.6 cm		
1.4 mm	8.0 cm		
450 μm	2.5 cm		
800 μm	1.8 cm		
800 μm	6.5 cm		
1.4 mm	7.5 cm		
400 μm	4.3 cm		
300 μm	6.5 cm		
600 μm	5.7 cm		
3.1 mm	42 cm		
2.8 mm	63 cm		
1.93 mm	48 cm		
1.63 mm	7.8 cm		
430 μm	4.9 cm		
1.3 mm	9.4 cm		
670 μm	8.2 cm		
1.8 mm	15 cm		
1.44 mm	31 cm		
630 μm	5.9 cm		
1.25 mm	27.5 cm		
2.75 mm	68.5 cm		
400 μm	3.5 cm		
530 μm	4.3 cm		
550 μm	4.1 cm		
720 μm	7.6 cm		
700 μm	12 cm		
300 μm	4.6 cm		
470 μm	1.1 cm		
1.8 mm	1.67 m		
300 μm	2.2 cm		

Table 2 – Dataset from rotary-shear experiments presented in Fig. 2.

α^* or w^*	D	Normal stress	Slip speed	Material	Reference
α^* (mm)	D (m)	Rotary-shear experiments (Final melt thickness α^* as a function of slip)			
0,033	24,5	1.2-1.4 MPa	0.85 m.s ⁻¹	Gabbro, India	Hirose & Shimamoto (2005a & 2005b)
0,084	30,8				
0,104	38,5				
0,116	46,8				
0,108	57,7				
0,135	78,6				
0,0659	1,53	13 MPa	1.14 m.s ⁻¹	Balmuccia peridotite, Italy	Del Gaudio et al. (2009)
0,1265	2,77				
0,0384	5,7				
0,0498	9,9				
0,1311	30,5				
0,112	31,31				
w^* (mm)	D (m)	Rotary-shear experiments (axial shortening w^* as a function of slip)			
3	6	20 MPa	3 m.s ⁻¹	quartz microgabbro "Absolute black"	Niemeijer et al. (2011), Fig. 5
2.3	5				
1.7	4				
1	3				
0.5	2				
2	1	20 MPa	1 m.s ⁻¹	SHIVA, Rome, Italy	Niemeijer et al. (2011), Fig. 8
1	0.5		3 m.s ⁻¹		
4	2				
0.5	1				
0.25	0.5				
1	2	10 MPa	1.3 m.s ⁻¹	SHIVA, Rome, Italy	Niemeijer et al. (2011), Fig. 9 HV-1, Kochi, Japan
0.45	10				
0.3	8				
0.2	6				
0,1	4				
1.4	10	15 MPa	1.28 m.s ⁻¹	Adamello tonalite Italian Alps	Di Toro et al. (2006), Fig. 2 SHIVA, Rome, Italy
0.9	8				
0.4	6				
0.1	4				
0.75	1				
0.9	2	15 MPa	1.28 m.s ⁻¹	Adamello tonalite Italian Alps	Di Toro et al. (2006), Fig. 2 SHIVA, Rome, Italy
1.3	3				
1.5	4				