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1 **Experimental evidence for wall rock pulverization during dynamic rupture at ultra-high**  
2 **pressure conditions**

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## 24 **Abstract**

25 The mechanisms triggering intermediate and deep earthquakes have puzzled geologists for  
26 several decades. There is still no consensus concerning whether such earthquakes are  
27 triggered by brittle or ductile mechanisms. We performed a deformation experiment on a  
28 synthetic lawsonite-bearing blueschist at a confining pressure of 3 GPa and temperatures from  
29 583 to 1,073 K. After deformation, the recovered sample reveals conjugated shear fractures.  
30 Garnet crystals are dissected and displaced along these narrow faults and reveal micro- and  
31 nanostructures that resemble natural pulverization structures as well as partial amorphization.  
32 Formation of such structures is known to require strain rates exceeding  $10^2 \text{ s}^{-1}$  at low  
33 confining pressures and is explained by the propagation of a dynamic shear rupture. The  
34 absence of shearing in the pulverized wall rock is taken as evidence that these structures pre-  
35 date the subsequent heat-producing frictional slip. In analogy to observations at low pressure  
36 we infer that the garnet structures in our experiment result from rapid propagation of a shear  
37 fracture even at the high pressure exerted on the sample and thus suggest that brittle  
38 deformation is possible at lower crustal to upper mantle depths.

39

40 **Keywords:** pulverization, high-pressure deformation, dynamic rupture, lawsonite-blueschist,  
41 DDIA apparatus, acoustic emissions

## 42 **1. Introduction**

43 During subduction at convergent plate margins, intermediate depth (70-300 km) and deep  
44 (>300 km) earthquakes are common. Yet, the processes triggering earthquakes at the high  
45 pressures (>1 GPa) prevailing at these depths are poorly understood. Both brittle (Raleigh and  
46 Paterson, 1965; Kirby, 1987; Green II and Burnley, 1989; Dobson et al., 2002; Hacker et al.,

47 2003b; Schubnel et al., 2013; Okazaki and Hirth, 2016; Ferrand et al., 2017; Gasc et al., 2017;  
48 Incel et al., 2017, 2019; and references therein) and ductile (Braeck and Podladchikov, 2007;  
49 Kelemen and Hirth, 2007; John et al., 2009; Thielmann et al., 2015; Poli and Prieto, 2016;  
50 Prieto et al., 2017; and references therein) mechanisms have been proposed.

51 While the suggested ductile mechanisms involve self-localizing failure by dissipative  
52 heating and thermal runaway situations, the considered brittle mechanisms involve dynamic  
53 rupture. In the former case, one expects significant shear deformation prior to seismic slip,  
54 whereas in the latter case, wall rock damage may occur due to high strain rates and rapidly  
55 changing stresses near a propagating rupture tip prior to frictional heating of the shear fracture  
56 surfaces (Ben-Zion, 2003). Wall rock deformation associated with paleoearthquakes inferred  
57 from the presence of pseudotachylytes, a rock type often assumed to be the result of frictional  
58 melting and subsequent quenching (McKenzie and Brune, 1972; Sibson, 1975), was  
59 interpreted as evidence for a thermal runaway mechanism (John et al., 2009; Deseta et al.,  
60 2014). Studies of pseudotachylyte veins and their surrounding wall rocks in naturally  
61 deformed rocks do, however, pose a number of challenges due to postseismic deformation and  
62 recovery processes with respect to their formation (Guermani and Pennacchioni, 1998;  
63 Mancktelow, 2006; Kirkpatrick and Rowe, 2013). Recent microstructural observations indicate  
64 extremely high stresses in wall rocks around lower crustal earthquake zones, and abundant  
65 fragmentation (Angiboust et al., 2012) without observable shear (Austrheim et al., 2017;  
66 Petley-Ragan et al., 2018). Such fragmentation is often referred to as ‘pulverization’ when  
67 occurring around faults in the shallow seismogenic regime and is assumed to result from  
68 dynamic rupture processes (Dor et al., 2006; Mitchell and Faulkner, 2009; Rempe et al.,  
69 2013).

70 In a previous experimental study, synthetic polycrystalline lawsonite-bearing blueschist  
71 samples were deformed at confining pressures corresponding to lower crustal to upper mantle

72 depths (Incel et al. 2017). Faulting was accompanied by the record of acoustic emissions and  
73 the growth of eclogite-facies minerals monitored using in-situ powder diffraction.  
74 Examination of the recovered run products revealed several conjugated faults decorated with  
75 nanocrystalline eclogite-facies transformation products in samples that entered the stability  
76 field of eclogite. Therefore, Incel et al. (2017) suggested that failure occurred due to  
77 transformation-induced instabilities, a mechanism titled transformational faulting (see also  
78 Kirby, 1987; Green II and Burnley, 1989). In one of these samples (BS\_3\_1073 in Incel et al.  
79 2017), garnet grains that are cut and displaced by the faults show microstructures similar to  
80 what has been described from ‘pulverized’ garnets in natural fault rocks from lower crustal  
81 lithologies (Austrheim et al., 2017; Petley-Ragan et al., 2018). Here, we further investigate the  
82 micro- and nanostructure of different garnets found in this sample. In addition, we model the  
83 spatial relation of the occurrence and absence of garnet fracturing relying on linear elastic  
84 fracture mechanics with the aim to gain further insight into the nucleation and failure  
85 mechanisms of intermediate-depth earthquakes.

## 86 **2. Experimental methods and analytical techniques**

### 87 2.1 Sample description and preparation of the starting material

88 A lawsonite-bearing blueschist from Alpine Corsica served as sample material. To avoid  
89 any initial texture of the starting material a chemically homogeneous part of this blueschist  
90 was crushed and sieved to a grain size  $<38\ \mu\text{m}$ . The major phases are glaucophane and  
91 lawsonite in a ratio of  $\sim 3:2$  making up  $\sim 90\ \text{vol.}\%$  of the powder. Minor and accessory phases  
92 are garnet, omphacite, actinolite, titanite, and phengite. The blueschist powder was hot-  
93 pressed at 3 GPa and 923 K for 24 hours in a piston-cylinder and machined to approx. 2.1 mm  
94 in diameter and 3 mm in height. After hot-pressing, the sample’s phases reveal a

95 homogeneous texture (Fig. 1a). In particular, the garnets are randomly distributed throughout  
96 the sample as evidenced by the Mn-element distribution map (Fig. 1b).

## 97 2.2 D-DIA deformation experiment

98 A 9×9×8 mm sized amorphous Boron-epoxy cuboid was used as pressure medium for the  
99 experiment performed using a D-DIA apparatus. The hot-pressed sample is located in the  
100 middle of this cuboid, sandwiched between two gold foils and two alumina pistons, and  
101 surrounded by a BN sleeve that is inserted into a graphite furnace.

102 The deformation apparatus is mounted on the GSECARS beamline at the Advanced  
103 Photon Source, National Laboratory, Argonne, IL, USA. The use of synchrotron radiation  
104 during deformation made it possible to calculate the differential stress as well as the strain and  
105 the strain rate during deformation. Stress was calculated on lattice planes of glaucophane  
106 using powder diffraction patterns that were taken every five minutes of the deforming sample  
107 and the strain was measured by using radiographs of the sample that were also taken every  
108 five minutes during deformation. Details of the stress and strain calculation are described by  
109 Incel et al. (2017). Additionally, the D-DIA apparatus is equipped with an acoustic emission  
110 (AE) system. Acoustic emissions were recorded using a sampling rate of 50 MHz and in  
111 trigger mode with a trigger threshold of 250 mV on two channels. The duration of the largest  
112 AEs recorded were in the range of a few hundred microseconds. Hence, the interval size over  
113 which the stress measurements are made is around six magnitudes larger than the event  
114 duration. Further details on the experimental and AE setup can be found in Wang et al.  
115 (2003), Gasc et al. (2011), and Schubnel et al. (2013).

116 First, the sample assembly was loaded hydrostatically to a confining pressure ( $P_c$ ) of 3 GPa  
117 (here confining pressure  $P_c$  equals the least principal stress  $\sigma_3$ ). Then, heating was initialized  
118 by increasing the furnace power manually to reach a temperature of 583 K. The sample was  
119 kept at these conditions for 30 min before deformation with a strain rate of approx.  $5 \times 10^{-5} \text{ s}^{-1}$

120 commenced. While deformation of the sample proceeded, the power was increased in 10 W  
121 steps to heat up the sample from initially 583 to 1073 K. Heating steps were initiated at 5, 12,  
122 18, 20, 25, 30, and 35 % axial strain.

### 123 2.3 Analytical techniques

124 Microstructural analyses of the recovered sample were performed using a field-emission  
125 scanning electron microscope (FE-SEM) with an acceleration voltage of 15 kV. To  
126 investigate the nanostructure of this sample, three focused-ion beam (FIB) sections were cut  
127 using a FEI-Helios G4 UC-Dual Beam system for imaging, analysis and transmission electron  
128 microscopy (TEM). The nanostructural analyses were conducted using a FEI Tecnai TEM and  
129 a Jeol JEM 2011 transmission electron microscope. For both machines, the acceleration  
130 voltage was 200 keV.

## 131 **3. Results**

### 132 3.1 Mechanical data and acoustic emissions

133 During the first stage of deformation at a temperature of 583 K, the sample was strained by  
134 5 % and the differential stress increased towards a peak stress of approx. 3 GPa, i.e. the level  
135 of the confining pressure (Fig. 2). The differential stress decreased continuously during  
136 syndeformational heating of the sample. A total axial strain of 40 % was accumulated by the  
137 time the temperature reached 1073 K. In total 10 acoustic emissions were recorded between  
138 10 to 19 % axial strain. Two events recorded at around 12 % and at approx. 19 % axial strain,  
139 respectively, were large events almost reaching voltage saturation of the recording system (5  
140 V).

## 141 3.2 Microstructural analyses

142 Microstructural investigations using the field-emission scanning electron microscope (FE-  
143 SEM) revealed faults oriented at an angle of around  $45^\circ$  to the direction of the axial stress  $\sigma_1$   
144 crosscutting the entire sample (Fig. 3a). It is possible that either some of the fault-filling  
145 material (gouge) of the major faults was lost during sample preparation or the fault surfaces  
146 were separated during decompression. However, some relicts of gouge material are preserved  
147 at the fault borders. These relicts show holes after interaction with the electron beam of the  
148 SEM, implying that this material is poorly crystalline (Fig. 3c).

149 Two different types of garnets can be identified in the deformed sample, (i) garnet crystals  
150 that are dissected and displaced along narrow faults ( $<1 \mu\text{m}$  wide; Figs. 3c; 4a-c) and (ii)  
151 unsheared garnet grains situated at some distance to the nearest fault ( $\sim 0.5$  to  $1.5 \text{ mm}$ ; Figs.  
152 3b, 4d, e). Two example sets of displaced garnet parts show apparent shear displacements of  
153  $42$  and  $58 \mu\text{m}$ , respectively (Figs. 3c; 4a). Back-scattered electron (BSE) imaging reveals that  
154 the displaced garnet halves are fragmented into pieces with diameters  $<1 \mu\text{m}$  (Figs. 3d; 4b)  
155 and some of them were dragged along during slip (red arrow in Fig. 3d). Garnet grains located  
156 further away from the fractures do not seem to be fragmented at this magnification in the  
157 SEM (Figs. 3b; 4d, e).

## 158 3.3 Nanostructural analyses

159 Transmission electron microscope (TEM) analyses were performed at three different sites  
160 (Fig. 3b, c) to investigate the nanometer-scale structures of: (i) the fault-gouge of the narrow  
161 fault dissecting and displacing a garnet crystal, (ii) a garnet crystal that is cut by this narrow  
162 fault, hereafter referred to as damage-zone garnet, and (iii) a garnet crystal located at a  
163 minimum of  $\sim 0.5 \text{ mm}$  from any fault, denoted as host-rock garnet. The fault-gouge contains  
164 garnet crystals with sizes ranging from  $<20 \text{ nm}$  to  $\sim 100 \text{ nm}$ . In bright field mode TEM



165 images, the material embedding the garnets appeared brighter than the garnet grains (Fig. 5a).  
166 An electron diffraction pattern of this area showed a few diffraction spots but also a diffuse  
167 halo (Fig. 5b) implying a combination of domains that have lost their long-range crystalline  
168 order and crystals large enough to produce diffraction spots.

169 Scanning transmission electron microscopy (STEM) of the damage-zone garnet  
170 demonstrates that this grain is completely shattered into small fragments (Figs. 5c, 6a)  
171 surrounded by a fault-filling material exhibiting vesicles and idiomorphic crystals (Fig. 6a).  
172 Using energy-dispersive spectroscopy (EDS) measurements these idiomorphic crystals were  
173 identified as omphacite. The surrounding matrix mainly consists of Si, Al, Na, and Ca in  
174 addition to O. The diffraction pattern obtained from a circular area with a diameter of  $\sim 500$   
175 nm in the shattered damage-zone garnet shows few large and several weak diffraction spots,  
176 indicative of a polycrystalline material, together with a diffuse halo in its center (Fig. 5c, d).  
177 This halo is less prominent than the one obtained from diffraction of the fault-gouge (Fig. 5b).  
178 A bright field and a dark field mode image of the same area within the shattered damage-zone  
179 garnet taken at high magnification document grain-size and crystal-orientation variation,  
180 respectively (Fig. 6b, c). The bright field image shows several grains ranging in size from  $\sim 10$   
181 to  $\sim 50$  nm in diameter (Fig. 6b). A quantitative determination of the grain-size distribution is  
182 hampered by the abundant overlap of small grains. Lattice fringes are clearly visible locally  
183 (Fig. 6c). However, the brightness variation in the corresponding dark-field mode image  
184 suggests variable lattice orientation on the nanometer scale.

185 Nanostructural analysis of the host-rock garnet shows numerous fracture-like features and  
186 possibly subgrains ranging in diameter from several hundred nanometers to  $\sim 5 \mu\text{m}$  (Fig. 5e).  
187 In contrast to the damage-zone garnet, though, very few grains have diameters  $< 100$  nm (Fig.  
188 5c, e). A diffraction pattern of an area with a diameter of  $\sim 500$  nm (Fig. 5e) indicates a high  
189 degree of crystallinity in this zone (Fig. 5f).

## 190 4. Discussion

### 191 4.1 Garnet pulverization due to dynamic rupture propagation

192 Our microscopic analyses revealed extensive fragmentation and grain-size reduction of the  
193 damage-zone garnet. Its diffraction pattern shows numerous weak diffraction spots indicating  
194 the presence of many small crystals. Additionally, a diffuse halo is observed implying that  
195 some subdomains are either amorphous or too small ( $<10$  nm) to produce well-defined  
196 diffraction spots (Fig. 6c; Yund et al., 1990). On the contrary, the host-rock garnet is fully  
197 crystalline and mainly reveals subgrain-formation (Fig. 5e, f). These experimental  
198 microstructures are strikingly similar to what Austrheim et al. (2017) described as  
199 “pulverization structures” in garnets found in close vicinity to a pseudotachylyte produced  
200 during coseismic loading and faulting of granulites from the Bergen Arcs, Norway, and to  
201 those found in garnets from mylonitic micaschists in the Sesia Zone, Swiss Alps (Trepmann  
202 and Stöckhert, 2002).

203 Based on the record of acoustic emissions (Fig. 2), faulting and associated pulverization of  
204 the wall rock occurred at a confining pressure of  $\sim 3$  GPa, a differential stress of  $\sim 2.5$  GPa, in  
205 a temperature range from 640 to 720 K, at an experimentally imposed strain rate of  $5 \times 10^{-5} \text{ s}^{-1}$   
206 and at  $\sim 10$  to 19 % axial strain. One characteristic feature of pulverized structures is the  
207 absence or the low amount of shearing of the fragments (Trepmann and Stöckhert, 2002;  
208 Austrheim et al., 2017). Due to the small fragment sizes of the damage-zone garnet it was not  
209 possible to measure their orientation. It is likely that the fragments experienced some shearing  
210 during further deformation as evidenced by the ‘tailing’ of the damage-zone garnet into the  
211 fault (red arrow in Fig. 3d). However, because the fragments’ arrangement still mimics a  
212 typical garnet crystal shape we can exclude significant shearing of the bulk crystals and their  
213 environment (Figs. 3c, 4a, 5c). As evidenced by the microstructure of the recovered sample

214 showing lawsonite pseudomorphs as well as by in-situ monitoring of the mineral assemblage  
215 during deformation, extensive reaction comprising the dehydration of lawsonite took place at  
216 a later stage during deformation (Incel et al., 2017). Therefore, most of the remaining strain  
217 was accommodated by lawsonite dehydration involving a solid volume change of around -  
218 20 %.

219 When occurring at upper crustal depth (<15 km), wall-rock damage is explained by high  
220 strain rates and stresses around a dynamically propagating rupture tip ( Reches and Dewers,  
221 2005; Dor et al., 2006; Doan and Gary, 2009; Mitchell et al., 2011; Bhat et al., 2012; Rempe  
222 et al., 2013; Aben et al., 2017a, 2017b, 2016; Xu and Ben-Zion, 2017; Griffith et al., 2018).  
223 Due to the much lower strength of rocks in tension than in compression, recently published  
224 studies highlight the impact of isotropic or quasi-isotropic tension on the pulverization of  
225 rocks (Xu and Ben-Zion, 2017; Griffith et al., 2018). In the model of Grady (1982) the author  
226 quantifies the relation between the energy needed to create new fracture surfaces during  
227 fragmentation and the inertial or kinetic energy available due to rapid loading. Later Glenn  
228 and Chudnovsky (1986) added a strain energy term to this model that accounts for the energy  
229 consumed by the solid until reaching its tensile strength. Based on this model, fragment size  
230 will not vary over a wide range of strain rates (strain energy dominated regime). Once the  
231 tensile strength of the material is exceeded, the fragment size exponentially decays with  
232 increasing strain rate (kinetic energy dominated; Grady, 1982). Across natural faults, this  
233 situation is realized in close distance (~5 cm) to the fault plane where strain rates are expected  
234 to be high (Griffith et al. 2018). In the present study, we follow a similar approach as  
235 presented in Griffith et al. (2018) in order to investigate, if the above model can explain the  
236 observed difference in fragmentation intensity between the damage-zone and the host-rock  
237 garnet. First, we check if the calculated fragment size matches our measured garnet fragments  
238 using the Glenn and Chudnovsky (1986) model. Then, we test if the corresponding strain rates

239 fit the predicted strain rates around a dynamically propagating mode II crack tip at the  
240 respective positions of the damage-zone and the host-rock garnet using linear elastic fracture  
241 mechanics (see Freund, 1990). For these calculations, we used a density  $\rho = 3,000 \text{ kg m}^{-3}$ , a  
242 garnet fracture toughness  $K_{IC} = 1.5 \text{ MPa}$  (Mezeix and Green, 2006), and a range in garnet  
243 tensile strength  $\sigma^* = 433 \text{ MPa}$  to  $4.3 \text{ GPa}$  deduced using reported single crystal or aggregate  
244 compressive strengths (Pardavi-Horváth, 1984; Kavner, 2007) assuming that the tensile  
245 strength of a solid is around a third of its compressive strength. We used a shear modulus  $\mu =$   
246  $64 \text{ GPa}$  for glaucophane (Bezacier et al., 2010), a Poisson's ratio  $\nu = 0.22$  (Cao et al., 2013),  
247 and two different rupture velocities  $v_r = 0.8c_s$  and  $v_r = 0.9c_s$  with  $c_s$  being the shear wave  
248 speed. The microstructural observations indicate a coseismic slip of a few tens of micrometer  
249 (Figures 3c; 4a) corresponding to a range in fracture energy  $G_c$  of  $\sim 0.1$  to  $100 \text{ J m}^{-2}$   
250 (Passelègue et al., 2016), deduced from experiments accounting for the uncertainty regarding  
251 the critical slip distance (further explanations in Passelègue et al., 2016).

252 The calculated fragment size distribution matches quite well the measured garnet  
253 fragments of the host-rock and the damage-zone garnet, respectively (Figures 5; 7a). Based on  
254 linear elastic fracture mechanics, strain rates at the position of the host-rock garnet range from  
255  $\sim 10^2$  to  $\sim 10^4 \text{ s}^{-1}$  (Figure 7b). Combining the results of both calculations, the host-rock garnet  
256 plots within the strain energy dominated regime (Figure 7). This fits well the nanostructural  
257 analysis that reveals some fracture-like features, but mostly polygons that are homogeneously  
258 sized resembling subgrains (Figures 5e; 8b). To explain the extensive fragmentation of the  
259 damage-zone garnet, strain rates must have been high enough to exceed garnet's tensile  
260 strength. Based on the measured fragment sizes of the damage-zone garnet, strain rates have  
261 to be at least  $10^8 \text{ s}^{-1}$  (Figure 7a). Since the crack tip passed through this garnet crystal, such  
262 high strain rates are realized within the damage-zone garnet volume in close vicinity to the  
263 rupture tip (Figures 7b; 8a, b).

264 The theory of linear elastic fracture mechanics provides an asymptotic solution for a semi-  
 265 infinite crack that is only valid in the near-tip field (Freund, 1990). This requirement is  
 266 obviously difficult to satisfy regarding the length ratio of the shear fracture relative to the  
 267 respective garnet distances to the fault plane. A previous theoretical study on the relation  
 268 between wall-rock damage and depth reports an increase in the amount of fracture energy  
 269 dissipated in the off-fault medium with increasing depth (Okubo et al., 2019). The authors  
 270 also state that the width of the damage-zone decreases with depth. Consequently, this implies  
 271 that at deeper depth the off-fault damage will be confined to a narrow zone around the fault.  
 272 Showing extensively damaged garnets only in close vicinity to the fault, our experimental  
 273 study confirms these theoretical results.

#### 274 4.2 Frictional melting of blueschist

275 Sliding of the fracture surfaces in mode II causes a temperature increase, which may  
 276 eventually lead to melting of the fracture surfaces. We deduce the presence of a solidified  
 277 melt from the amorphous material present in the fault-filling material (Fig. 5a,b) and in some  
 278 places intruded the shattered damage-zone garnet (Fig. 6a). In this “melting scenario”, the  
 279 vesicles in the amorphous material reflect fluid exsolution during decompression of the melt  
 280 and the observed idiomorphic omphacite crystals nucleate and grow during cooling of the  
 281 melt (Fig. 6a). Assuming a wet basalt solidus temperature  $T_s$  of  $\sim 1,000$  K at  $\sim 3$  GPa (Hacker  
 282 et al., 2003a), a sliding-related increase in temperature of 280-360 K over the temperatures  
 283 prevailing during the AE activity (640-720 K) would be sufficient to cause local melting.  
 284 After Cardwell et al. (1978), the temperature rise  $\Delta T$  on a slipping fault can be expressed as

$$285 \quad \Delta T = \frac{\tau D}{\rho c_p \sqrt{\pi \kappa t_{\text{slip}}}} \quad (\text{eq. 1})$$

286 with shear stress  $\tau$ , shear displacement  $D$ , density  $\rho$ , specific heat capacity  $c_p$  (1,100 J kg<sup>-1</sup> K<sup>-1</sup>  
 287 at  $\sim 1,000$  K; Hartlieb et al., 2016), thermal diffusivity  $\kappa$  ( $\sim 10^{-6}$  m<sup>2</sup> s<sup>-1</sup>), and slip duration  $t_{\text{slip}}$ .

288 The nominal shear stress  $\tau$  and normal stress  $\sigma_n$  acting on the fault are  $\sim 1.25$  GPa and 4.25  
289 GPa, respectively (with  $\theta \approx 45^\circ$ ,  $\sigma_1 = 5.5$  GPa, and  $\sigma_3 = 3$  GPa). Assuming a minimum total  
290 displacement  $D = 42 \mu\text{m}$  (Figs. 3c) and only 10 % of that slip to have happened coseismically  
291 and a sliding velocity of  $\sim 1 \text{ m s}^{-1}$  (for a crack-like rupture; Schubnel et al., 2013) gives a slip  
292 duration  $t_{\text{slip}}$  of  $\sim 4.2 \times 10^{-6}$  s. These estimates result in a  $\Delta T$  of  $> 390$  K indeed exceeding the  
293 difference between prevailing assembly temperature and the sample's solidus temperature.

294 The presence of a melt film on the fault surfaces can lead to fault lubrication (Di Toro et  
295 al., 2006). Dynamic shear strength  $\tau_f$  of a fault with a continuous melt film strongly depends  
296 on the ratio between the width  $w$  of the molten zone that is filling the fault plane and the slip  
297 displacement (e.g., Ferrand et al., 2018)

$$298 \quad \tau_f = \frac{\rho[H + c_p \Delta T]w}{(1 - \eta)D} \quad (\text{eq. 2})$$

299 with the latent heat of fusion  $H$  ( $\approx 3 \times 10^5 \text{ J K}^{-1}$ ) and the radiative efficiency  $\eta$ . In our sample,  
300 the narrow fault that contains molten material shows a width of  $w \approx 100\text{-}500$  nm. Previous  
301 studies showed that the seismic efficiency, as function of the mechanical energy spent on slip  
302 during rupture, ranges between  $0.1 < \eta < 0.5$  (Poli and Prieto, 2016). Within this span, equation  
303 (2) gives a dynamic shear strength of the fault as low as  $\sim 7$  to 66 MPa (Fig. 9a) corresponding  
304 to friction coefficients of  $\sim 0.002$  to 0.015 (Fig. 9b), i.e., significant lubrication (Figure 8c).

## 305 5. Conclusion and implications

306 The micro- and nanostructures observed in the damage-zone garnet, which resemble  
307 pulverization structures in natural rocks at upper as well as at lower crustal depths, can be  
308 explained by extensive fragmentation due to high strain rates associated with a dynamically  
309 propagating shear fracture. Such microstructures are not, however, compatible with fault  
310 models that involve failure by self-localizing thermal runaway mechanisms. In such a

311 situation, one would expect to see evidence of pre-failure shear strain in the wall rocks, and  
312 the local differential stress levels should not rise above the initial externally imposed far-field  
313 stress (John et al., 2009). However, it has been demonstrated that high local stresses, e.g., due  
314 to coseismic loading, are required to fracture garnet (Trepmann and Stöckhert, 2002). After  
315 the passage of the crack tip, frictional sliding of the fault surfaces causes melting and fault  
316 lubrication. Our experimental study emphasizes the importance of dynamic rupture as a brittle  
317 precursor to unstable frictional slip even at upper mantle depths.

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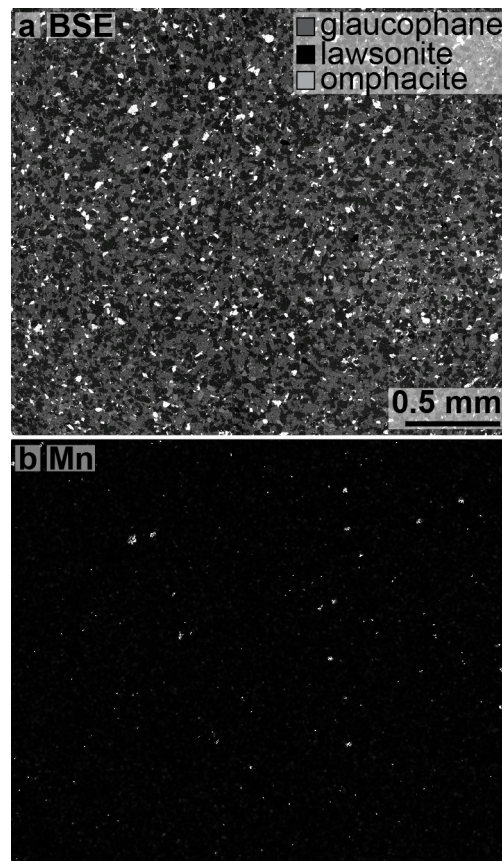
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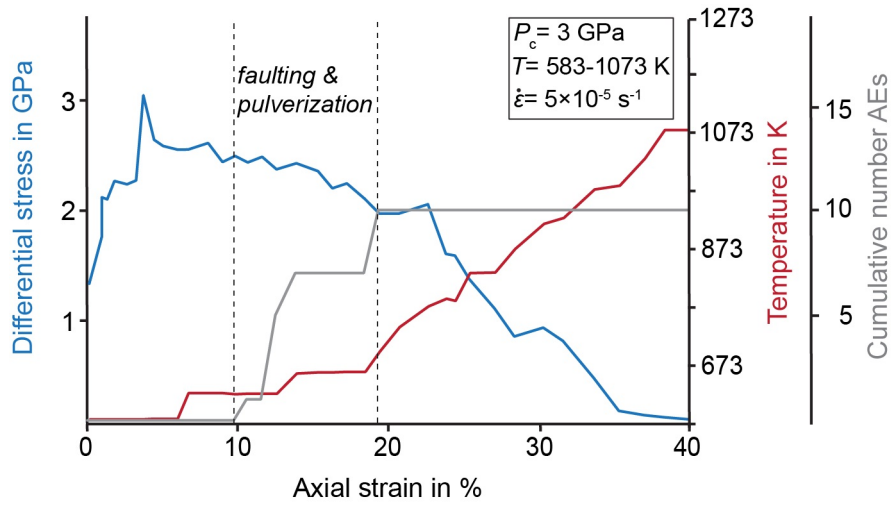
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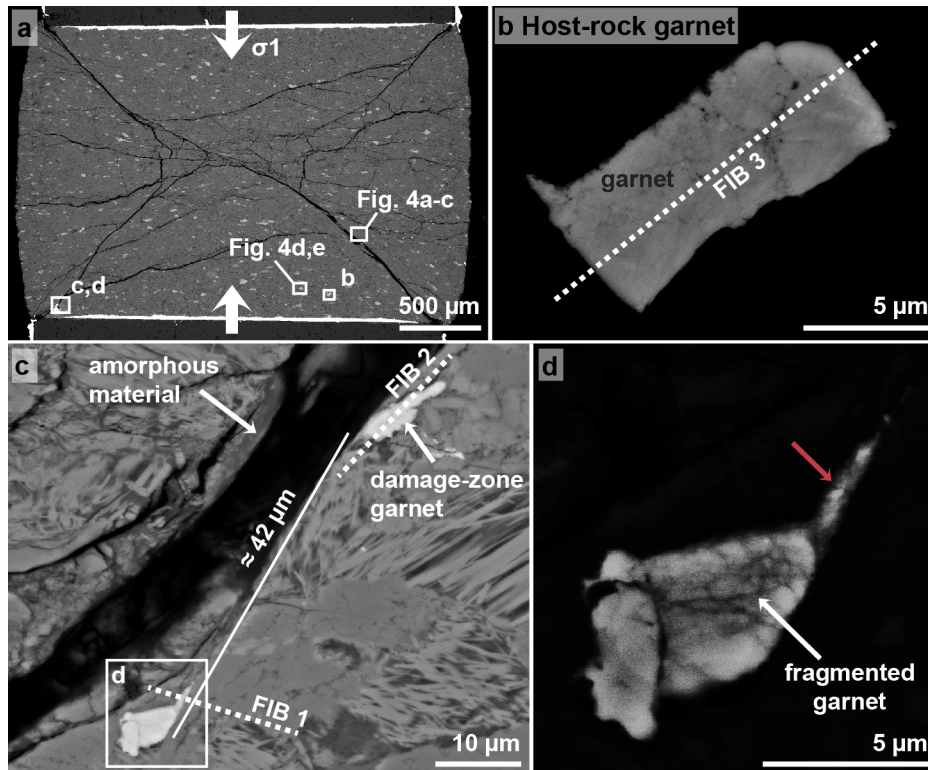


497

498 **Figure 1:** The starting material after hot-pressing. a) Backscattered electron (BSE) image  
 499 showing that the phases are homogeneously distributed throughout the sample. b) A Mn-  
 500 distribution map of the same region was used to highlight the location of garnet crystals.  
 501

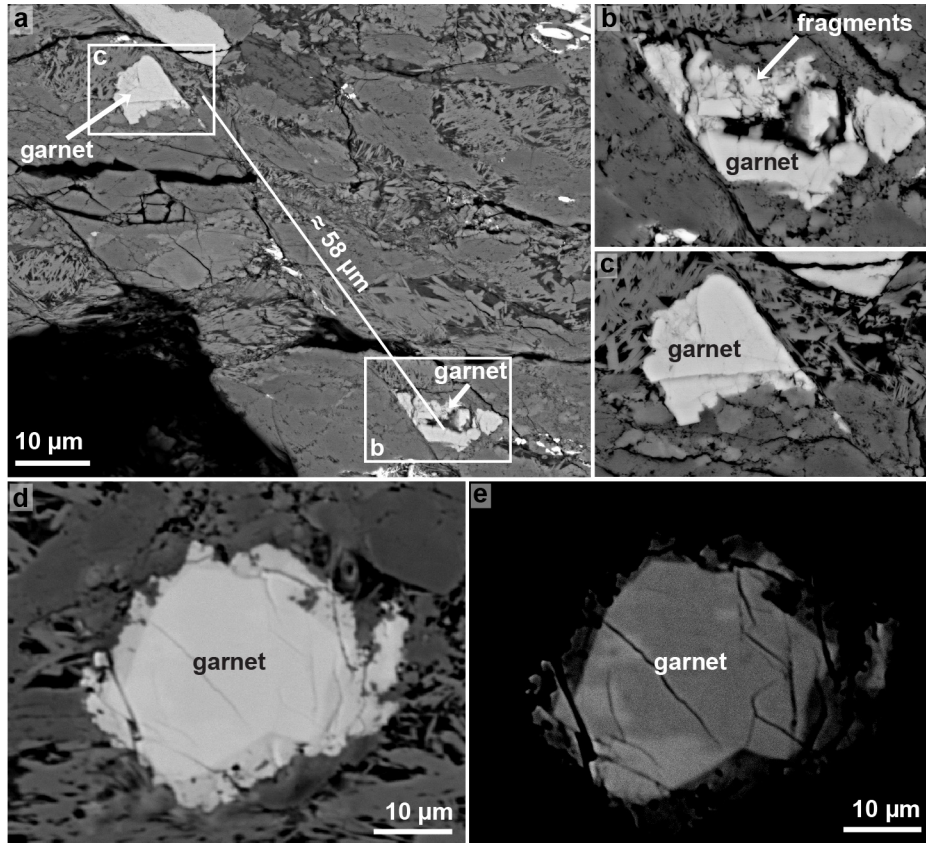


502  
 503 **Figure 2:** Differential stress, temperature, and cumulative number of acoustic emission (AE)  
 504 events over axial strain. Based on the record of AEs (grey curve), faulting and pulverization  
 505 occurs at a confining pressure ( $P_c$ )= 3 GPa, a differential stress of ~2.5 GPa (blue curve), in a  
 506 temperature range of 640 to 720 K (red curve), at an imposed strain rate ( $\dot{\epsilon}$ ) =  $5 \times 10^{-5} \text{ s}^{-1}$ , and  
 507 an axial strain  $\epsilon = \sim 10-19 \%$ .  
 508



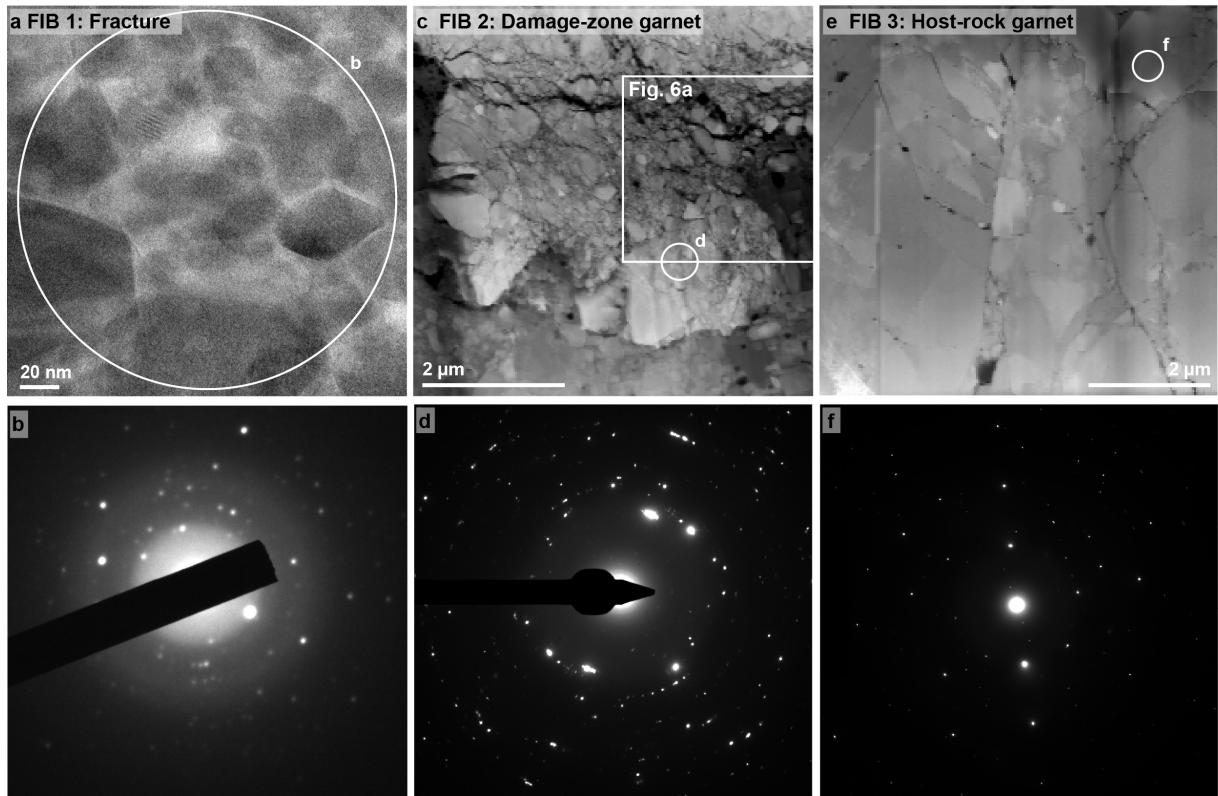
509  
 510 **Figure 3:** Backscattered electron images of the lawsonite-bearing blueschist sample after  
 511 deformation. a) Overview image of the sample showing faults oriented at  $\sim 45^\circ$  towards  $\sigma_1$   
 512 crosscutting the sample. The positions of the high-magnification images in b-d as well as the

513 position for Fig. 4 are highlighted with white rectangles, respectively. b) The host-rock garnet  
 514 crystal that is located at  $\sim 0.5 \mu\text{m}$  to the closest fault. The location of FIB section 3 is shown as  
 515 white dashed line. c) A garnet pair that is dissected and displaced along a narrow fault. The  
 516 apparent displacement is  $\sim 42 \mu\text{m}$ . The locations for FIB sections 1 and 2 are marked by white  
 517 dashed lines. The white rectangle shows the location of a garnet half presented in d. d) At  
 518 high magnification and high brightness contrast the garnet half appears fragmented into  
 519 several pieces  $< 1 \mu\text{m}$ .  
 520



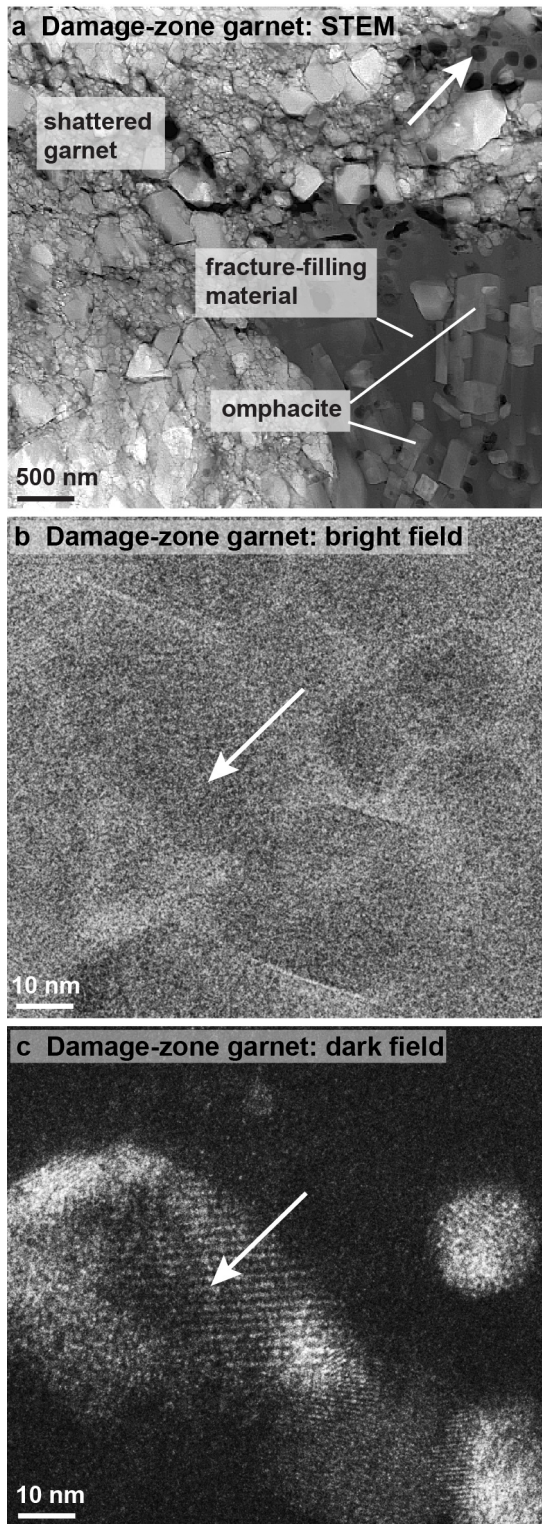
521  
 522 **Figure 4:** a) Backscattered electron images of another pair of dissected and displaced crystals  
 523 that are offset by a narrow fault. The offset along the narrow fault is  $\sim 58 \mu\text{m}$ . b and c) The  
 524 two halves of the garnet pair showing extensive fragmentation in b). d and e) This garnet was  
 525 found at  $\sim 1 \text{mm}$  from the nearest fault and seems relatively intact. e) Same crystal as in d)  
 526 with the image taken at a higher brightness contrast at the SEM.  
 527





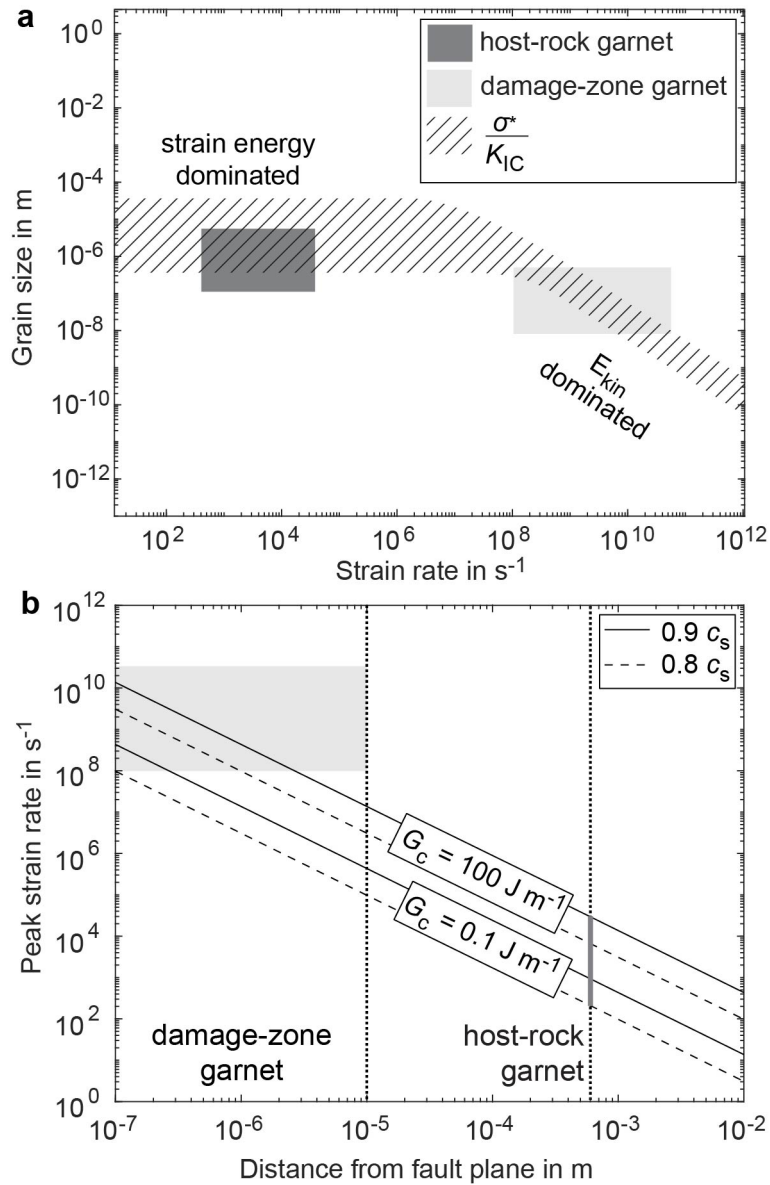
528

529 **Figure 5:** Transmission electron microscope images of the three FIB sections 1-3. a and b)  
 530 FIB 1 revealing the fault-filling material. a) Bright field image showing dark (i.e. crystalline)  
 531 garnet crystals floating in a bright (i.e. amorphous) material. The location chosen for a  
 532 diffraction pattern is highlighted by a white circle. b) Diffraction pattern of the fault-filling  
 533 material exhibiting few large and several weak diffraction spots and a diffuse halo. c and d)  
 534 FIB section 2 cut across the damage-zone garnet next to the narrow fault. c) In STEM mode  
 535 the damage-zone garnet appears to be completely shattered into pieces  $<1 \mu\text{m}$ . The white  
 536 rectangle marks the position of Fig. 6a and the white circle shows the location of the  
 537 diffraction pattern. d) Diffraction pattern of the shattered damage-zone garnet. Many weak  
 538 diffraction spots indicate the presence of numerous small crystals. There is also a diffuse halo  
 539 in the center of the diffraction pattern. e and f) FIB section 3 cut across the relatively intact  
 540 host-rock garnet located at  $\sim 0.5 \mu\text{m}$  from a fault. e) The fragments are much larger than those  
 541 found in the damage-zone garnet. The white circle highlights the location chosen for a  
 542 diffraction pattern. f) The diffraction pattern of the area shown in e) presents a crystalline  
 543 structure.  
 544



545

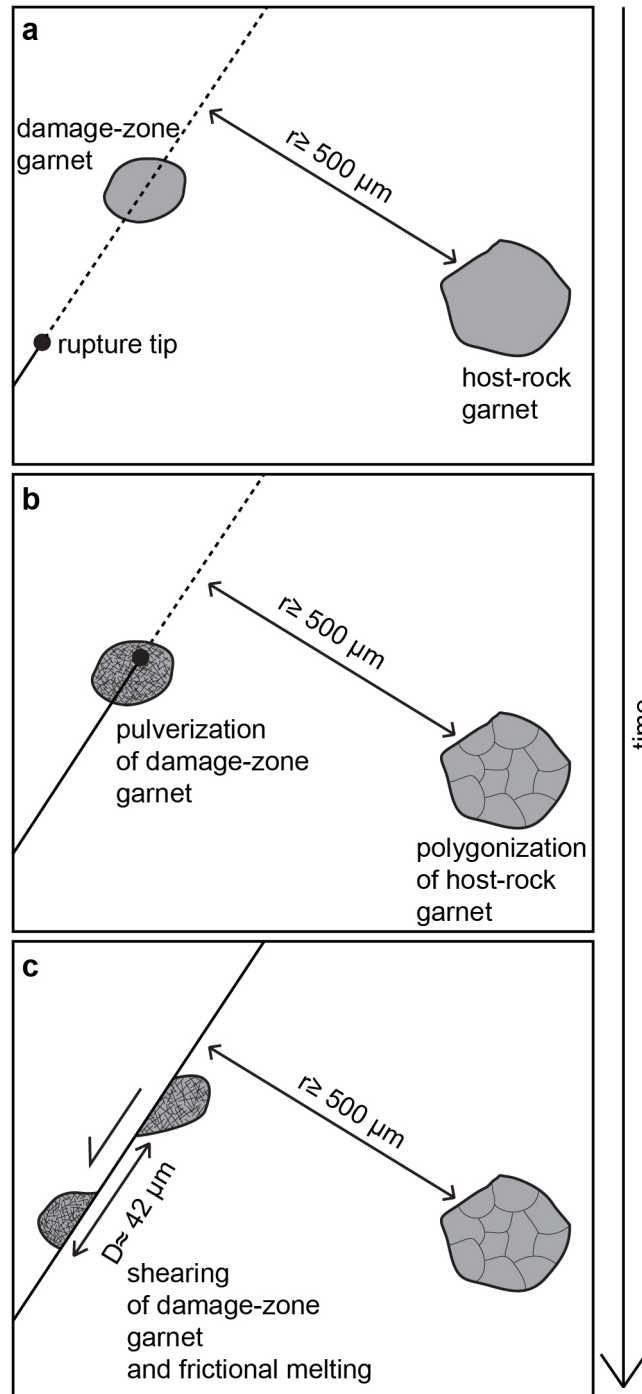
546 **Figure 6:** Transmission electron microscopy images of the shattered damage-zone garnet. a)  
 547 STEM image revealing the fault-filling material surrounding the damage-zone garnet. Within  
 548 this material vesicles (white arrow) and idiomorphic omphacite crystals can be found. b)  
 549 Bright field image taken in the same zone as the diffraction pattern (Fig. 5c, d). c) The same  
 550 area as shown in b) taken in dark field mode. The large grain marked by the white arrow  
 551 shows subdomains ( $\leq 10$  nm) that are slightly tilted.



552

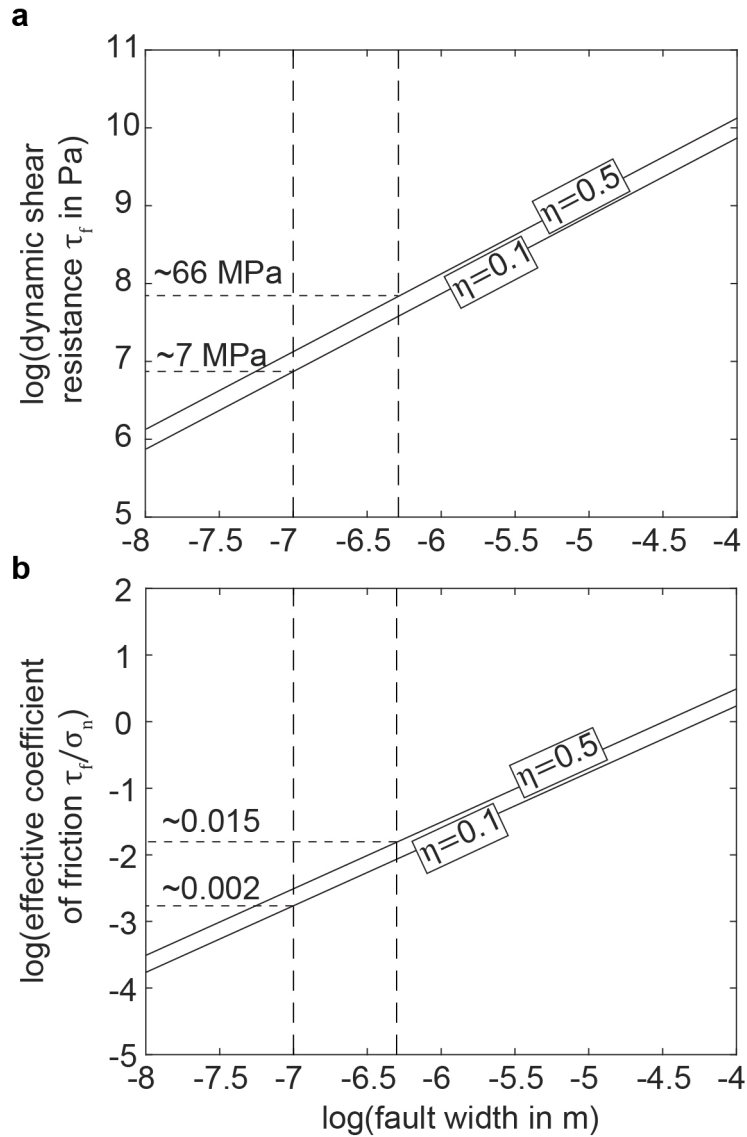
553 **Figure 7:** a) Grain size versus strain rate plot based on the models of Grady (1982) and Glenn  
 554 and Chudnovsky (1986). The measured garnet fragments of the host-rock (dark grey  
 555 rectangle) and the damage-zone garnet fragments (light grey rectangle) fit quite well the  
 556 calculated fragment sizes (dashed area) over the investigated strain rate range. The minimum  
 557 tensile strength for garnet aggregates is 433 MPa deduced from the compressive strength of  
 558 garnet aggregates under confinement (Kavner, 2007). The maximum tensile strength was  
 559 deduced from single crystal hardness measurements reported by Pardavi-Horváth (1984). For  
 560 both we assumed that the tensile strength is around a third of the compressive strength. The  
 561 fracture toughness of garnet is  $\sim 1.5$  MPa (Mezeix and Green, 2006). b) Peak strain rates as a  
 562 function of distance from the fault plane for two different rupture speeds  $v_r$  ( $0.8c_s$  and  $0.9c_s$ )  
 563 and two different fracture energies  $G_c$  ( $0.1 \text{ J m}^{-1}$  and  $100 \text{ J m}^{-1}$ ; Passelègue et al., 2016).

564



565

566 **Figure 8:** a) Prior to rupture, both garnet crystals are intact. b) The rupture tip passes through  
 567 the damage-zone garnet. Pulverization of the damage-zone garnet occurs due to the extreme  
 568 strain rate around the crack tip ( $E_{kin}$  dominated Figure 7a). Because strain rate decays with  
 569 distance, the host-rock garnet that is located at a minimum distance of  $\sim 500 \mu\text{m}$  to the nearest  
 570 fault only shows the formation of subgrains (strain energy dominated Figure 7a). c) Shearing  
 571 behind the rupture tip causes the displacement of the garnet halves and eventually frictional  
 572 melting of the fault surfaces.



573

574 **Figure 9:** a) Dynamic shear resistance and b) effective coefficient of friction versus width of  
 575 the slip-associated melt layer. The vertical dashed lines mark the measured width of the  
 576 molten zone filling out the fault along which a garnet pair is dissected and displaced (Fig. 3c).  
 577 Recent studies show that the radiative efficiency increases with depth ranging between  $\eta= 0.1$   
 578 to 0.5 for intermediate depth earthquakes (50-300 km). The dashed horizontal lines indicate  
 579 the intercept of this radiative efficiency range and the width range of the molten zone ( $\sim 100$  to  
 580 500 nm) measured in the sample. The dynamic shear resistance would be  $\sim 7$  to 66 MPa (a)  
 581 resulting in an effective coefficient of friction of  $\sim 0.002$  to 0.015 (b).