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Ito

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3 **Grain size distributions of fault rocks: a comparison between experimentally and**  
4 **naturally deformed granitoids.**

5

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37 **Grain size distributions of fault rocks: a comparison between experimentally and**  
38 **naturally deformed granitoids.**

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41

42 **1 Abstract**

43 We have investigated the grain size distribution (GSD) of granitoid fault rock by comparing  
44 experimentally produced gouge with fault rock from the Nojima Fault Zone. Triaxial experiments  
45 were carried out on wet and dry intact samples of Verzasca Gneiss at  $T = 300$  and  $500^{\circ}\text{C}$ ,  $P_c = 500$   
46 and  $1030$  MPa,  $\epsilon = 0.013\text{-}1.6 \times 10^{-4}\text{s}^{-1}$ . The GSD has been determined from SEM-micrographs and is  
47 characterized by the slope,  $D$ , of its  $\log(\text{frequency})\text{-}\log(\text{radius})$  histogram. The GSD is not fractal; we  
48 observe two slopes for all GSDs. The larger grains in experimentally deformed samples have a  $D$ -  
49 value,  $D_{>}$ , of 2.04 and 2.26 for feldspar and quartz gouge. Cracked grains yield values of  $D = 1.5\text{-}1.6$ .  
50 Increasing the confining pressure or temperature decreases the  $D$ -value. For grains smaller than  $\sim 2$   
51  $\mu\text{m}$  another  $D$ -value,  $D_{<}$ , of 0.9-1.1 is observed. The grain radius at the slope change,  $r_K$ , corresponds  
52 to the grinding limit of quartz, so that  $r_K$  probably represents a change in the dominant comminution  
53 mechanism from grinding to attrition-processes. The GSD obtained from deformation experiments  
54 agrees well with results for the Nojima Fault Zone:  $D_{>} = 2.02$  for gouge and 1.64 for cracked grains;  
55  $D_{<} = 0.97$ . Grain size reduction in fault zones develops by a two-stage process: rupturing creates  
56 cracked grains; further displacement of fragments causes further comminution by wear and attrition.  
57 Cracked grains have been used to calculate the surface energy associated with faulting; it follows that  
58 this energy forms a small fraction in the total energy-budget of earthquakes.

59

60 **Keywords: faulting, granite, gouge, grinding, surface density, power-law distribution**

61 **2. Introduction**

62 **2.1 Overview**

63

64 Major fault zones and earthquakes in the upper crust typically develop highly localized zones  
65 of deformation. At the macroscopic scale, faults usually appear as narrow, highly deformed  
66 zones or networks embedded in wide damage zones of relatively low deformation.

67 Observations on natural fault zones have shown that this localization of deformation along  
68 faults occurs both in space and time (e.g. Mitra, 1993; Faulkner et al., 2003; Wilson et al.,  
69 2003). Most of the displacement takes place in very narrow zones within the gouge of  
70 typically a few to 100  $\mu\text{m}$  wide, leaving the rest of the gouge inactive (e.g. Wibberley and  
71 Shimamoto, 2005; Chester et al., 2005; Ma et al., 2006). Bos et al. (2000) have made this  
72 same observation for laboratory experiments on halite.

73

74 Within these narrow deformation zones gouge with a wide range of grain sizes has been  
75 produced (e.g. Storti et al., 2003, Boullier et al., 2004a). At the microscopic scale, the fault  
76 rock consists of cataclasite or gouge with properties that change as a function of time and  
77 space. Displacement on a fault zone causes the evolution from a thin fracture to a wider gouge  
78 zone. The microstructures of the fault rocks influence and reflect the mechanics of faulting,  
79 i.e., earthquakes. For example, evidence for coseismic hydraulic fracturing (Boullier et al.,  
80 2004b) or for a change in the frictional properties of faults resulting from their grain shape  
81 (Mair et al., 2002) has been obtained through microstructural studies. Despite experimental  
82 data and field observations on gouge, the properties of fault rock and the evolution of cracked  
83 grains to gouge still remain rather poorly understood.

84

85

86 **2.2 Fault rocks**

87

88 A number of definitions for fault rocks can be found in the literature (e.g. Sibson, 1977;  
89 White, 1982; Wise et al., 1984; Chester et al., 1985; Schmid and Handy, 1991). Usually,  
90 cohesive and noncohesive fault rocks are attributed to cataclasis, a process involving  
91 fracturing, frictional sliding, dilatancy, and rigid body rotation between grain fragments,  
92 grains, or clusters of grains. Sibson (1977) has defined gouge as a cataclastic rock without  
93 cohesion; the cohesive rock is called cataclasite. Many cohesive cataclasites in nature are

94 thought to be the product of syn- or post-tectonic healing processes (Schmid and Handy,  
95 1991).

96

97 In this study, however, we will distinguish different fault rocks on the basis of microstructures  
98 rather than inferred mechanical properties (see Heilbronner & Keulen, 2006). We will use the  
99 term “cracked grains” for angular fragments that have fractured without much displacement  
100 (such that the original geometrical relationship of the fragments can still be recognized). The  
101 term “gouge” will be used for more highly deformed fault rocks where the grains are more  
102 rounded and the original spatial relationship between the fragments is lost. The term “fault  
103 rocks” thus applies to both cracked grains and gouge.

104

### 105 **2.3 Grain size distributions**

106

107 A large number of studies has been dedicated to measuring and analyzing grain size  
108 distributions of fault rocks (e.g. Mitra, 1993; Blanpied et al., 1995; Lin, 1999; Bos et al.,  
109 2000; Stewart et al., 2000; Kato et al., 2003). The grain size distribution (GSD) is considered  
110 as an important parameter for the characterization of the gouge (e.g. Sammis et al., 1987;  
111 Marone and Scholz, 1989; Hadizadeh and Johnson, 2003; Wilson et al., 2005; Chester et al.,  
112 2005). Within a certain size range the GSD may be self-similar (scale invariant) and in this  
113 case can be described by a power-law relationship (Mandelbrot, 1982). The slope of some  
114 power-law distributions,  $D$ , is called the fractal dimension (Turcotte, 1992; Korvin, 1992, and  
115 references in both). This  $D$ -value has often been used to compare the GSD of natural  
116 earthquake and fault zones, like the San Andreas Fault zone (e.g. Anderson et al., 1980;  
117 Blenkinsop, 1991), the Moine Thrust (Blenkinsop and Rutter, 1986) or fault systems in Japan  
118 (Monzawa and Otsuki, 2003; Otsuki et al., 2003) with fault rocks derived from experiments  
119 (e.g. Marone and Scholz, 1989; Morrow and Byerlee, 1989; Amitrano and Schmittbuhl,  
120 2002).

121

122 For the measurements of the GSD, several different methods have been used (see Table 1).

123 The most classical one, originating from sedimentology, is sieving. By this method a  
124 distribution of mass against size is obtained (Anderson et al., 1980; Sammis et al., 1986).

125 Other studies have used the Coulter counter (An and Sammis, 1994; Amitrano and  
126 Schmittbuhl, 2002; Wilson et al., 2005) or optical microscopy on thin sections (Biegel et al.,  
127 1989; Monzawa and Otsuki, 2003); these methods provide a description of the GSD in terms

128 of frequency versus grain size. A disadvantage of sieving and the use of light microscopes is  
129 that the size of smallest detectable grains is  $\sim 2 \mu\text{m}$  (Table 1).

130

131 An easy way to measure the sizes of the grains smaller than  $2 \mu\text{m}$  is scanning electron  
132 microscopy (SEM), as was done by Sammis et al. (1987) and Shao and Zou (1996), but in  
133 their samples no small grains were observed. Olgaard and Brace (1983), Yund et al. (1990),  
134 and Chester et al. (2005) found grains as small as 15-50 nm with the TEM, but the TEM is  
135 unsuitable for measuring large numbers of grains necessary to obtain a size distribution.  
136 Wilson et al. (2005) used a Coulter counter and measured grains of 40 nm to  $300 \mu\text{m}$ .  
137 Nevertheless, the frequency distribution of grains sized between 30 nm and  $2 \mu\text{m}$  is not very  
138 well known. We find that using scanning electron micrographs of impregnated and polished  
139 sections of gouge, reliable GSD in the size range larger than 60 nm can be obtained.

140

141 In a previous study (Heilbronner & Keulen, 2006) two types of gouge can be distinguished:  
142 cracked grains and gouge. Cracked grains and gouge can be discriminated on the basis of  
143 particle shape and grain size distributions. The D-values of gouge are always higher than  
144 those of cracked grains. Composition (quartz versus feldspar) influences various aspects of  
145 the particle shape of cracked grains and gouge (elongation, paris factor, etc.).

146

147 Table 1 shows a compilation of published results on the GSD of fault rocks: values of the  
148 fractal dimension vary between 0.8 and 2.6, but values of D between 1.5 and 2.1 are the most  
149 common for granitoid rocks. High D-values are generally obtained for more recently active  
150 fault zones, though a wide spread of D-values within a single fault zone can be observed.

151

152 As there appears to be a link between fault rock microstructure (shape and GSD) and the  
153 mechanical behavior of faults, and since fossil fault rocks are important clues to seismic  
154 activity in the past, this study focuses on the microstructures of naturally and experimentally  
155 produced faults. In this paper we examine the influence of experimental conditions on the  
156 fault rock microstructure and we compare the results to a natural fault rock cored from a  
157 recently active segment of the Nojima Fault (Japan). The focus of the paper is on the grain  
158 size distributions (D-values cracked grains and gouge) and on the geometry and spatial  
159 distribution of the fault rock in the samples (D-mapping). The aim is to study the influence of  
160 temperature, confining pressure, axial shortening rate, total axial shortening, and fluid content  
161 on the grain size development of gouge material in deformed granitoids.

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163

### 164 **3. Materials and methods**

165

166

#### 167 **3.1 Description of samples**

168

##### 169 3.1.1 Verzasca gneiss:

170 The starting material for the study of experimental fault rocks is cored from a gneiss that was  
171 collected in the Val Verzasca (location: Swiss coordinates 704.65 - 126.30, Figure 1a). The  
172 composition is granitic, consisting of ~29% plagioclase, ~27% K-feldspar, ~35% quartz and  
173 ~7% mica, mainly biotite. The average grain size is ~280  $\mu\text{m}$  and the rock exhibits a faint  
174 foliation and lineation while appearing isotropic at the thin section scale.

175

##### 176 3.1.2 Nojima fault rock:

177 The samples for the study of natural fault rocks are taken from the Nojima fault, which  
178 transects a Cretaceous granodiorite and the overlying sediments of the Kobe Group and Osaka  
179 Group (Figure 1b). The modal composition of the granodiorite is ~33% quartz, ~34%  
180 plagioclase, ~19% K-feldspar, ~11% biotite and ~4% hornblende. One year after the Hyogo-  
181 ken Nambu Earthquake (Kobe Earthquake, 17 January 1995,  $M_w = 7.2$ ) three bore holes have  
182 been drilled through the Nojima fault at different depths (Ito et al., 1999; Ohtani et al., 2000;  
183 Tanaka et al., 2001; Murakami et al., 2002). The fault rock samples studied here (Figure 1c  
184 and 1d) are from the Hirabayashi drill core, which has reached the fault zone at a depth  
185 between 623.1 to 625.3 m (the thin sections are provided by the Geological Survey of Japan).

186

187 The history, as derived from the cores, can be divided into two major deformation episodes  
188 (Boullier et al., 2004a). An earlier deformation episode is related to a left-lateral movement  
189 on the Nojima Fault (Fabbri et al., 2004) prior to the deposition of the Middle to Late Eocene  
190 Kobe Group. Pseudotachylytes associated with this episode have been dated at 56.4 Ma by  
191 fission-tracks on zircons (Murakami and Tagami, 2004). A more recent Quaternary  
192 deformation episode is associated with a right-lateral movement along the Nojima fault and  
193 offsets the Plio-Pleistocene unconformity at the base of the Kobe Group (Murata et al., 2001).  
194 The studied thin sections are from the Quaternary deformation episode.

195

### 3.2 Rock deformation experiments

Samples of Verzasca gneiss measuring 6.39 mm in diameter and 10.1 to 12.9 mm in length ( $L_0$ ) are cored parallel to the mesoscopic lineation. Table 2 lists the used abbreviations. At this size, the sample width corresponds to approximately 25 grains. After polishing their flat ends the samples are dried in an oven at 110°C and atmospheric pressure for at least 24 hours. The samples are wrapped in Ni-foil and placed in a gold jacket. In all cases except for experiment 63nk, a "room-humidity" sample, approximately 0.2 wt% H<sub>2</sub>O (1.68 to 2.14 µl) is added and the jacket is weld-sealed. During welding, the jacket and sample are cooled in order to avoid water loss. The exact water content of the "wet" and "room-humidity" samples has not been determined, and we assume that also in room-humidity samples some water is present at grain boundaries.

Deformation is carried out in a Grigg's solid medium deformation apparatus as described by Tullis & Tullis (1986) and De Ronde (2004), using straight furnaces, alumina pistons and sodium chloride as the confining medium. The pistons are slightly bevelled, thus reducing their diameter from 6.39 mm to 6.2 mm at the sample-piston interface. Temperature is measured with Pt/Pt-10%Rh or Cromel/Alumel thermocouples. The standard or reference deformation conditions are "wet" samples, 300°C temperature, 500 MPa confining pressure,  $10^{-4} \text{ s}^{-1}$  shortening rate, and a total axial displacement of 3.5 - 5.0 mm (samples 38nk and 64nk). In separate runs, one or more parameters are varied: 500°C temperature, 1030 MPa confining pressure, up to 5.8 mm total displacement,  $10^{-6} \text{ s}^{-1}$  shortening rate or no H<sub>2</sub>O is added (see Table 3). During the experiment, pressure, load and displacement are recorded on a strip chart. Using the rigC4 program ([www.unibas.ch/earth/micro/software/](http://www.unibas.ch/earth/micro/software/)), with a correction for rig stiffness, the force-displacement and stress-strain curves are calculated.

After the experiment, the samples are quenched. To avoid loss or damage of the non-cohesive fault rocks, the jackets are punctured and the samples are impregnated with epoxy. The samples are cut along the compressional axis and the deformed length of the sample is re-measured. Thin sections are prepared (polished and carbon coated) for observation in the light and scanning electron microscope.

### 3.3 Grain size analysis



230 For the grain size analysis, images with a wide range of magnifications are used. The  
231 micrographs of the experimental microstructures are collected using a Philips XL30 ESEM  
232 scanning electron microscope (20kV acceleration voltage; backscatter contrast); those of the  
233 Nojima Fault Zone are acquired with a Leica Stereoscan 440 SEM. For each sample a cascade  
234 of micrographs is prepared, starting with overviews at low magnifications (100x) and  
235 zooming in (to 20,000x) on selected areas with a factor of 2 to 2.5 from one magnification to  
236 the next. At each magnification, three or four images are taken of each kind of fault rock  
237 (quartz, feldspar, cracked grains and gouge). The total range of grain sizes (radius,  $r$ )  
238 observable at these magnifications is from  $<20$  nm to  $>200$   $\mu\text{m}$ , however, because of the  
239 scarcity of measurable grains at either end of the distribution, the statistically reliable range  
240 does not exceed three orders of magnitude.

241  
242 We use Image SXM (<http://reg.ssci.liv.ac.uk/>) and ImageJ (<http://rsb.info.nih.gov/ij/>) and  
243 special Macros and Plug-Ins (<http://www.unibas.ch/earth/micro/software>) for segmentation  
244 and image analysis. First, the images are pre-processed (removal of noise) and scaled to  
245 pixels (Figure 2a). Bitmaps of quartz and feldspar grains are obtained by density slicing at the  
246 corresponding grey levels and subsequent separation of grains (Figure 2b). The cross  
247 sectional areas,  $A$  ( $\text{pixel}^2$ ), and the perimeters,  $P$  (pixel), of the grains are measured using the  
248 Analyze menu of the software. After removing grains with areas smaller than  $20 \text{ pixel}^2$ , 300  
249 to 1200 grains per image could be used. Some of the images taken at the highest  
250 magnification could not be processed automatically; in this case the grain boundaries are  
251 traced manually.

252  
253 The resulting files of areas and perimeters were transferred to a spreadsheet program. The  
254 areas were restored by adding the pixel values of  $P$ , i.e., the band of pixels outlining the grain,  
255 to the area  $A$  within the outline. The equivalent radii,  $r$ , were calculated as

256  
257 
$$r = \sqrt{((A+P) / \pi)}$$
 (1)

258  
259 and collected in a histogram (Figure 2c). On a log(frequency)-log(size) plot with 20 bins per  
260 order of magnitude of grain size, the slope ( $D$ ) of the power-law fit is determined (Figure 2d).  
261 The results from all magnifications are then combined into a single plot (Figure 2e),  
262 multiplying the frequency in each bin by a factor that reflects the relative magnification of the

263 images and corrects for areas in the micrograph that were not investigated (e.g. large grains at  
264 edge of the micrograph, cracks). Overlapping bins of grain radii have been analyzed to check  
265 the correct up-scaling; the average frequency of the overlapping bins is used to draw the  
266 combined GSD plot. Figure 2f shows the individual D-values as a function of magnification.

267

268 In all samples, two distinct D-values are obtained: a shallower slope ( $D_{<}$ ) at smaller grain  
269 sizes and a steeper slope ( $D_{>}$ ) at larger grain sizes. The change from  $D_{<}$  to  $D_{>}$  occurs at  $r_K$ .

270 The point  $r_K$  is determined in the following manner: On grain size plots derived from the  
271 magnifications around  $r_K$ , the curves for  $D_{<}$  and  $D_{>}$  are drawn. The intersection of the  
272 corresponding line yields  $r_K$ .

273

274 From the experimentally deformed Verzasca Gneiss, only limited fault rock material is  
275 available. We therefore use only one combined log-log plot and one (average)  $D_{>}$  and  $D_{<}$   
276 value per magnification. In the case of the Nojima fault zone, we have more fault rock  
277 samples. We have been able to evaluate a number of cascades from different samples of the  
278 drill cores and derive 4 to 5  $D_{<}$  and  $D_{>}$  values per magnification.

279

280 For the evaluation of the largest grain sizes, light micrographs (taken with a ZEISS Axioplan  
281 Polarisation microscope) are included. The difference between the slopes ( $D_{>}$ ) determined  
282 from light and scanning electron micrographs is less than 0.05. We have also tested the  
283 reliability of the semi-automatic segmentation by comparing the derived D-values to D-values  
284 obtained from manual tracings of the outlines. On two sets of test images taken at  
285 magnifications of 500x and 2000x, the difference in  $D_{>}$  is less than 0.04.

286

287

## 288 **4. Results**

289

### 290 **4.1 Mechanical data**

291

292 The results of the triaxial compression experiments are shown as force-displacement diagrams  
293 (Figure 3a). With the exception of sample 60nk all samples display the same general  
294 behavior. Yield strength is reached after approximately 1 mm displacement, after that the load  
295 increases steadily, the differential stress decreases slowly after an axial shortening of 8-10%.

296 In some of the experiments deformed at 300°C, reaching the yield strength was accompanied

297 by audible cracking. Sample 60nk (run at the highest confining pressure) is different from all  
298 others in that it has produced distinct acoustic events, which accompany load drops on the  
299 order of 20 kN (corresponding to approximately 500 MPa).

300

301 At 45kN (1000-1200 MPa), sample 70nk, deformed at relatively high temperature (500°C)  
302 and low axial shortening rate ( $10^{-6} \text{ s}^{-1}$ ), is the weakest. At 60-75 kN (1750 MPa) sample 63nk  
303 (room humidity sample) is the strongest. Samples deformed at standard conditions (64nk and  
304 38nk) support loads of 55-65 kN, corresponding to strengths of 1300-1500 MPa.

305

306 For comparison with published data, we also show a stress-strain diagram (Figure 3b). The  
307 differential stress is calculated with a correction for an increase in cross-sectional area,  
308 assuming constant volume. The Verzasca Gneiss has approximately the same strength as  
309 Westerly Granite (Fig. 3b) deformed by Tullis and Yund (1977). The main differences with  
310 respect to Westerly Granite experiments are the higher shortening at yield stress (with typical  
311 values ~10% versus ~5%), the higher shortening rate (mainly at  $10^{-4} \text{ s}^{-1}$ , see Table 3), and the  
312 amount of added water with which the Verzasca Gneiss has been deformed. In both sets of  
313 experiments, a temperature increase from 300°C to 500°C has caused a strength reduction of  
314 approximately 200 MPa. Note that Tullis and Yund (1977) have used pyrophyllite or talc as  
315 confining medium, which means that the true confining pressures probably are ~2/3 of the  
316 published values (J. Tullis, pers. comm., 2004).

317

318 We have noted a difference between the measured lengths of the deformed samples and the  
319 deformed lengths calculated from the total axial displacements derived from the force-  
320 displacement records (compare Figure 3a and Table 3). The calculated displacements exceed  
321 the measured displacements by up to 1 mm. In the rest of the paper, we will therefore use the  
322 more conservative values of the measured displacements, not the displacements shown in  
323 Figure 3. This difference in measured length is not critical for the results, as the displacement  
324 is used semi-quantitatively in this study.

325

## 326 **4.2 Faulted samples**

327

328 In most samples deformed at  $T = 300^\circ\text{C}$  and  $P_c = 500 \text{ MPa}$ , through-going fault zones are  
329 formed accommodating the axial displacement within a single gouge zone. At  $T = 500^\circ\text{C}$ ,  $P_c$   
330 = 500 MPa and occasionally at  $T = 300^\circ\text{C}$ ,  $P_c = 500 \text{ MPa}$ , conjugate fault zones have

331 developed and partitioned the axial displacement over at least two fault zones, leading to a  
332 smaller displacement on any given fault. Sample 60nk deformed at  $P_c = 1030$  MPa shows a  
333 set of parallel faults. The corresponding load-displacement curve (Figure 3) shows an  
334 equivalent number of load drops. It is likely that each drop in the differential stress is related  
335 to the creation of a new fault plane.

336

337 The samples are deformed along relatively wide fault zones of 1.0-1.4 mm width, which  
338 become narrower with higher confining pressure or higher temperature; with increasing  
339 displacement, they tend to broaden. Most of the gouge zones are formed around mica grains  
340 or the connection of two mica grains. At a temperature of  $300^\circ\text{C}$  and shortening rates of  $10^{-4}$   
341  $\text{s}^{-1}$  biotite deforms by gliding, kinking and occasionally by fracturing.

342

343 Using the D-mapping technique described elsewhere (Heilbronner & Keulen, 2006), maps of  
344 local variations of the grain size distribution are prepared for three of the experimental  
345 samples (Figure 4). Cracking and gouge formation occur throughout the samples. In samples  
346 38nk and 102nk gouge regions coalesce; in sample 60nk there is less coalescence. The  
347 volume proportion of gouge formed during the deformation is determined by measuring the  
348 black area (indicating  $D > 1.75$ ) in Figure 4 and dividing this proportion by the total sample  
349 area. For increasing axial displacement,  $d$ , increasing amounts of gouge are formed. After  $d =$   
350 4 mm, sample 38nk (reference sample) has 24.5% of gouge; after  $d = 5.8$  mm, sample 102nk  
351 (deformed at the same conditions) has 54.8%. At higher confining pressures, there is less  
352 gouge: after  $d = 3.5$  mm, sample 60nk ( $P_c = 1000$  MPa) has only 15.5% gouge.

353

354

### 355 **4.3 Microstructures**

356

357 In the experiments at  $P_c = 500$  MPa and  $T = 300^\circ\text{C}$ , the original grains in the gouge zone form  
358 broad and elongated arrays of fractured particles (Figure 5a); at higher confining pressures,  
359 these arrays are much narrower (Figure 5b). Quartz and feldspar are both strongly fractured;  
360 at higher confining pressures, quartz is less intensively fractured than feldspar. A variation in  
361 grain shapes can be observed: quartz grains are more elongated, feldspar grains are more  
362 irregularly shaped (Figure 5c & 5d); the same variation in shape has been observed for a  
363 sample deformed at low PT-conditions (Heilbronner and Keulen, 2006).

364

365 Within the gouge zones, quartz and feldspar are fractured to very small grain sizes in all  
366 experiments (Figure 5e & 5f). The gouge is equidimensional and subangular in shape. In the  
367 cracked grains and the gouge, the grains smaller than 2  $\mu\text{m}$  diameter do not show any  
368 intragranular fractures, while grains larger than 10  $\mu\text{m}$  are usually fractured into a spectrum of  
369 smaller grains (Figure 5g & 5h). Also, grains smaller than 1  $\mu\text{m}$  are equiaxial and appear  
370 weakly indented. Within the gouge these small grains show more rounded edges.

371

#### 372 **4.4 Grain size analysis of experimental fault rocks**

373

##### 374 4.4.1. Dependence of grain size distribution on type and composition of fault rock

375

376 The grain size distributions of all of our experiments display the same general characteristics  
377 (Figure 6). On log-log plots of frequency versus equivalent radius, two distinct parts can be  
378 discriminated (compare also Heilbronner & Keulen, 2006). Both for cracked grains and  
379 gouge, two D-values are obtained: a lower value,  $D_{<}$ , for small grain sizes and a higher value,  
380  $D_{>}$ , for larger grain sizes. The grain size,  $r_K$ , at the intersection of the two curve fits is on the  
381 order of 1 - 2  $\mu\text{m}$ . In gouge, the lower range of grain sizes extends from  $r_{\min} \approx 20 \text{ nm}$  to  $r_K \approx 2$   
382  $\mu\text{m}$  with  $D_{<} \approx 1.0$  and the upper range extends from  $r_K$  to  $r_{\max} \approx 20 \mu\text{m}$  with  $D_{>} \approx 2.0$ . In  
383 cracked grains, the lower range of grain sizes extends from  $r_{\min} \approx 50 \text{ nm}$  to  $r_K \approx 1.5 \mu\text{m}$  with  
384  $D_{<} \approx 1.0$  and the upper range extends from  $r_K$  to  $r_{\max} \approx 100 \mu\text{m}$  with  $D_{>} \approx 1.5$ .

385

386 Comparing feldspar and quartz we note that the  $D_{<}$  values are approximately identical under  
387 almost all experimental conditions ranging from 0.72 to 1.02 in cracked feldspar, 0.74 to 1.09  
388 in cracked quartz, from 0.91 to 1.12 in feldspar gouge and 0.78 to 1.10 in quartz gouge (Table  
389 4). The  $D_{>}$  values, in contrast, are usually higher for quartz than for feldspar ranging from  
390 1.37 to 1.68 in cracked feldspar, 1.44 to 1.72 in cracked quartz, from 1.85 to 2.12 in feldspar  
391 gouge and 1.94 to 2.32 in quartz gouge (Table 4).

392

393 On average, the grain size,  $r_K$ , of cracked grains is slightly larger than in gouge (Table 5).  
394 However, this effect is more pronounced in quartz than in feldspar. The grain size,  $r_K$ , of  
395 cracked quartz ranges from 1.1 – 1.8  $\mu\text{m}$  and is smaller than  $r_K$  of cracked feldspar which  
396 ranges from 1.6 – 2.0  $\mu\text{m}$ . Similarly, in gouge,  $r_K$  of quartz is 1.1- 1.4  $\mu\text{m}$  while  $r_K$  of feldspar  
397 is 1.4 – 2.0  $\mu\text{m}$ .

398

#### 4.4.2. Dependence of grain size distribution on experimental conditions

399

400

401 The GSD of the experimentally deformed granitoid samples are shown in Figure 6 & 7, the  
402 derived values of  $D_{<}$ ,  $D_{>}$ , and  $r_K$  are shown in Tables 4 & 5. The  $d_K$  grain diameter is two  
403 times the  $r_K$  grain radius indicated in Figure 6, and has been used to compare our data to other  
404 studies. We consider the samples 64nk and 38nk as reference samples. The experiments are  
405 performed at  $T = 300^\circ\text{C}$ , confining pressure  $P_c = 500 \text{ MPa}$ , shortening rate  $\dot{\epsilon} = 1.5 \cdot 10^{-4} \text{ s}^{-1}$   
406 and 0.2 %wt  $\text{H}_2\text{O}$  was added to the samples. The total axial displacement,  $\Delta L$ , of 64nk and  
407 38nk is 2.2 and 3 mm, respectively. In Figure 6, each of the plots shows the GSDs resulting  
408 from the change of one of the experimental conditions with respect to 64nk.

409

410 The reference values derived from the GSDs of sample 64nk and 38nk are the following:  $D_{<}$   
411 ranges from 0.95 to 1.07 ( $D_{<}$  of cracked feldspar = 1.00;  $D_{<}$  of feldspar gouge = 1.02;  $D_{<}$  of  
412 cracked quartz = 0.95;  $D_{<}$  of quartz gouge = 1.07), cracked grains have somewhat lower  
413 values than gouge.  $r_K$  is 1.1 – 1.8  $\mu\text{m}$  (cracked feldspar = 1.8  $\mu\text{m}$ ;  $r_K$  of feldspar gouge = 1.6  
414  $\mu\text{m}$ ;  $r_K$  of cracked quartz = 1.6  $\mu\text{m}$ ;  $r_K$  of quartz gouge = 1.1  $\mu\text{m}$ ), quartz yields lower values  
415 than feldspar and cracked grains have higher values than gouge.  $D_{>}$  is 1.5 – 1.6 for cracked  
416 grains ( $D_{>}$  of cracked feldspar = 1.6;  $D_{>}$  of cracked quartz = 1.5) and 2.04 - 2.26 for gouge  
417 ( $D_{>}$  of feldspar gouge = 2.04;  $D_{>}$  of quartz gouge = 2.26), quartz has a lower  $D_{>}$ -value than  
418 feldspar for cracked grains and a higher  $D_{>}$ -value for gouge.

419

420 Increasing the confining pressure from 500 MPa to 1030 MPa leads to a marked reduction of  
421 both  $D_{<}$  and  $D_{>}$ , both for cracked grains and gouge. The grain sizes,  $r_K$ , at the slope break of  
422 cracked grains and gouge do not change significantly, with the possible exceptions of  $r_K$  of  
423 feldspar gouge and cracked quartz: these values increase from 1.6 to 1.8  $\mu\text{m}$  and from 1.5 to  
424 1.8  $\mu\text{m}$ , respectively. This variation is within the error of the measurements.

425

426 Reducing the shortening rate from  $10^{-4} \text{ s}^{-1}$  to  $10^{-6} \text{ s}^{-1}$  does not affect the values of  $D_{>}$ ; of the  $D_{<}$   
427 values only those of cracked feldspar and of quartz gouge appear to be lowered: from 1.00 to  
428 0.86 and 1.07 to 0.97, respectively. The grain size  $r_K$  of feldspar remains unchanged, while  
429 that of cracked quartz and quartz gouge increases from 1.5 to 1.8  $\mu\text{m}$  and from 1.1 to 1.4  $\mu\text{m}$   
430 respectively.

431

432 Increasing the axial shortening from 2.2 mm to 4.8 mm has no significant effect on any of the  
433 measured values. As has been mentioned before, the only difference between samples 64nk  
434 ( $\Delta L = 2.2$  mm), 38nk ( $\Delta L = 3.0$  mm) and 102nk ( $\Delta L = 4.8$  mm) is the width of the fault  
435 zone(s), i.e., the amount of fault rock (cracked grains and gouge) created during deformation.

436

437 Deforming the sample without adding water (under room-humidity conditions) has a marked  
438 effect on the GSD. With the exception of feldspar gouge, all  $D_{<}$  values are lowered  
439 significantly to values as low as 0.72 to 0.83. At the same time, all values of  $D_{>}$  increase. The  
440 grain size at the slope break is not affected except for  $r_K$  of cracked quartz, which is lowered  
441 from 1.5 to 1.2  $\mu\text{m}$ .

442

443 Increasing the temperature from 300°C to 500°C affects the  $D_{>}$  values. In particular, the  
444 values for feldspar gouge and quartz gouge are lowered from 2.04 to 1.95 and from 2.26 to  
445 2.07, respectively. The major effect on  $r_K$  is to increase the values for feldspar and quartz  
446 gouge from 1.6 to 2.0  $\mu\text{m}$  and from 1.1 to 1.4  $\mu\text{m}$ , respectively.

447

#### 448 **4.5 Grain size analysis of natural fault rocks**

449

450 For the Nojima samples the GSDs are given as bulk results, i.e., the different mineral phases,  
451 in particular, quartz and feldspar, were not separated. Comparing the GSDs of cracked  
452 minerals and gouge to the experimental results, we note that the values of  $D_{<}$  (1.64 and 2.02),  
453  $D_{>}$  (0.97) and  $r_K$  (1.9  $\mu\text{m}$  and 1.5  $\mu\text{m}$ ) are very close to the corresponding values obtained for  
454 feldspar cracked grains and gouge (see Table 4, Figure 8).

455

456

#### 457 **4.6 Characteristics of GSDs**

458

459 The GSDs of both the experimental and natural fault rocks display a number of characteristics  
460 that we want to highlight (compare Figure 6, 7 & 8). The lower limit of grain sizes that could  
461 be used for the calculation of statistically reliable distributions is  $r_{\text{min}} = 30$  to 60 nm for  
462 experimentally produced and  $r_{\text{min}} = 100$  to 200 nm for naturally produced gouge and cracked  
463 grains, respectively. Similarly, the upper limit is  $r_{\text{max}} = 50 - 100$   $\mu\text{m}$  for experimentally  
464 produced and  $r_{\text{max}} = 100 - 200$   $\mu\text{m}$  for naturally produced gouge and cracked grains,

465 respectively. In other words  $r_{\min}$  and  $r_{\max}$  of gouge are always smaller than  $r_{\min}$  and  $r_{\max}$  of  
466 cracked grains by a factor of  $\sim 2$ . Note, however, that  $r_{\min}$  is not the absolutely smallest grain  
467 size that can be detected in the SEM (Figure 9). Grains as small as  $r \approx 5$  nm have been  
468 detected both in experimental and in natural fault rocks, but are at the detection limit of the  
469 SEM so that their number is not large enough for a grain size analysis.

470

471 The grain size  $r_K$  at the slope break of the power law fit to the GSD has a remarkably constant  
472 value. In contrast to the marked decrease of minimum and maximum grain size from cracked  
473 grains to gouge, the  $r_K$  values remain approximately constant or decrease by  $< 30\%$  from an  
474 average value of 1.7 to 1.4  $\mu\text{m}$  in experimental and from 1.9 to 1.5  $\mu\text{m}$  in natural fault rocks  
475 (Figure 10).

476

477

## 478 **5 Discussion**

479

### 480 **5.1 Faulting experiments**

481

482 Brittle deformation experiments on granitoid rock at elevated pressure-temperature conditions  
483 have been performed by Griggs et al. (1960); Stesky et al. (1974); Carter et al. (1981);  
484 Blanpied et al. (1995); Kato et al. (2003) and others. Our experiments on Verzasca Gneiss  
485 compare best with the data on Westerly Granite published by Griggs et al. (1960) and Tullis  
486 and Yund (1977, 1980). The geometry and distribution of strain in our samples is variable  
487 (Figure 4). Therefore, we prefer to present our mechanical data in terms of force versus axial  
488 displacement rather than stress and strain (Figure 3a). For comparison with published data, we  
489 have converted our force data to stress data (Figure 3b).

490

491 The experiments on Westerly Granite at  $P_c = 500$  and 750 MPa have been performed without  
492 added water; samples have been pre-dried at a higher temperature (300°C) than ours (Tullis &  
493 Yund 1977). Since pyrophyllite or talc has been used as outer confining medium, the  
494 confining pressure has been reduced to approximately 350 and 500 MPa. Their experiments at  
495 750 MPa are therefore best compared to ours at 500 MPa, because we use NaCl as outer  
496 confining medium. The strengths of Westerly Granite and Verzasca Gneiss are comparable  
497 showing strengths of 1200 to 1600 MPa under similar conditions, although no experiments  
498 have been performed under identical conditions. The variation of the bulk shortening rate



499 between  $10^{-4}$  and  $10^{-6} \text{ s}^{-1}$  does not show any mechanical effect, consistent with the results of  
500 Tullis & Yund (1980, 1992).

501

502

503 The effect of a high confining pressure ( $\sim 1 \text{ GPa}$ ) is comparable to experiments by Tullis &  
504 Yund (1977, 1980). Very narrow slip zones accommodate the axial shortening. According to  
505 Tullis and Yund (1977) and Hirth and Tullis (1994), the formation of these narrow zones and  
506 the decrease in volume of fault rock is related to the transition from mode I to mode II  
507 cracking above  $P_c \approx 600\text{-}800 \text{ MPa}$ . The number of these narrow zones corresponds to the  
508 number of audible cracking events in the experiment 60nk. This observation suggests that the  
509 mechanical behavior is not a case of stick-slip as described in the literature (e.g. Byerlee and  
510 Brace, 1968; Stesky et al., 1974; Lockner et al., 1986). Instead, the apparent stick slip  
511 character of the mechanical data (Figure 3) is produced by multiple rupture events,  
512 demonstrating the great strength of already ruptured material at high confining pressure.

513

## 514 **5.2 Grain size distributions obtained in experiments**

515

516 Changing the physical parameters of experimental deformation also affects the grain size  
517 distribution. The effects are discussed with reference to sample 38nk and 64nk, which have  
518 been deformed under identical temperature, pressure and shortening rate conditions.

519

520 The largest effect is observed for increasing the confining pressure to  $\sim 1 \text{ GPa}$ . It causes a  
521 reduction in the  $D_{>}$ -values (Figure 6, 7, Table 4) and a decrease in the total amount of gouge  
522 and cracked grains in the fault zone (see Figure 4). This smaller amount of gouge is related to  
523 the transition from mode I to mode II cracking: mode I fracturing is dilatant; in mode II  
524 fracturing sharp shear fractures are formed when dilatancy is suppressed. Dilatancy  
525 suppression appears to inhibit the formation of large amounts of surface (small grains) and  
526 thus may reduce the  $D_{>}$ -value.

527

528 At higher temperatures, there is a decrease in the  $D_{>}$  values of the gouge (Figure 7, Table 4).  
529 In feldspar deformed at  $T = 500^\circ\text{C}$  we observe undulatory extinction and deformation bands,  
530 which are developed by micro-cracking at a very small scale, as described by Tullis and Yund  
531 (1987). At a temperature of  $500^\circ\text{C}$ , Hirth & Tullis (1994) have found an increased dislocation

532 activity in quartz. The less efficient grain size reduction at elevated temperature may be  
533 caused by a semi-brittle behavior.

534

535 The effects of not adding water and slower shortening rate are either unsystematic (higher  
536  $D_{<}$  values for feldspar and lower  $D_{<}$  values for quartz in gouge of non-water added sample  
537 63nk; Table 4) or not existent (samples 38nk, 64nk, 102nk; Table 4). The D-values are the  
538 same regardless of the amount of shortening of the sample (Figure 7, Table 3).

539

540 After 4 mm axial shortening a gouge of 1.0 mm thickness has been formed. Once a mature  
541 zone with gouge is established, this zone is growing wider at greater displacement, but  $D_{>}$  of  
542 the gouge is not changing. Marone and Scholz (1989) have also observed similar D-values for  
543 different shear strains. In general, it can be concluded that the effects of changing the physical  
544 conditions of deformation on the grain size distribution are of minor importance except for  
545 the increase in confining pressure, and, to a lesser extent, temperature.

546

### 547 **5.3 Cracked material and gouge**

548

549 All experimental and Nojima fault rock samples show the same type of grain size distribution  
550 (GSD). For small grain sizes ( $d < 2 - 3 \mu\text{m}$ ) the GSD is characterized by D-values of  $\sim 1$ ,  
551 whereas for the larger sizes the D-values range between  $\sim 1.5$  and  $\sim 2.3$  (Table 4). The gouge  
552 always shows higher D-values than the cracked material (Figure 6, 8, Table 4). It is  
553 interesting to note that the values of  $D_{<}$  in cracked material and gouge, in quartz and feldspar,  
554 and in experimental and in Nojima fault rocks all have very similar values and the same range  
555 of variation (Table 4). Furthermore, this grain size range is not affected by changing the  
556 experimental deformation conditions. In contrast,  $D_{>}$  of gouge is the parameter that has been  
557 measured in several other studies before (Table 1). It is the most suitable of the D-values to  
558 compare natural fault zones in different regions or to compare natural faults with  
559 experimentally obtained values.

560

561 Higher D-values indicate a more efficient grain size comminution producing a larger number  
562 of smaller grains. The difference between gouge and cracked grains clearly demonstrates that  
563 the initial rupturing process does not produce the full amount and size fraction of small grains  
564 observed in gouge (Figure 6, 8). Instead, slip on the fault zone causes further comminution of  
565 grains in the gouge and an increase in  $D_{>}$ . Thus, it can be concluded that after the initial

566 fragmentation of the rock by rupturing there is a post-rupture processing during slip which  
567 takes place in the gouge to produce the higher values of  $D_{>}$ .

568

569 Further comminution of grains in gouge has been observed by Sammis et al. (1986, 1987) and  
570 Hadizadeh & Johnson (2003). However, any quantitative characterization of fault  
571 displacement by GSD measurements appears impossible because an increase in sample  
572 shortening from 2.2 to 4.8 mm in our experiments (64nk and 102nk) does not change any of  
573 the  $D$ -values systematically (Figure 6, 9, Table 5). Thus, the more efficient grain  
574 comminution in gouge must take place in the first stages of the fault displacement after  
575 rupturing.

576

577 In the experiments performed under identical conditions (38nk, 64nk, 102nk) the variation of  
578  $D$ -values of the cracked material is always larger than in the gouge (Table 4). The only  
579 systematic difference appears to be the consistently higher  $D_{>}$  values of quartz gouge  
580 compared to feldspar gouge, i.e., for the coarser grain size fraction ( $> 2 \mu\text{m}$ ) quartz shows a  
581 more efficient comminution of grains than feldspar. The  $D_{>}$ -values for quartz and feldspar  
582 gouge are almost identical in these three experiments and indicate that this  $D$ -value is very  
583 well reproducible and therefore forms a reliable parameter to compare the GSD of fault zones.  
584 Another difference is the lower  $r_K$ -value for quartz compared to feldspar. Thus, the change in  
585 the GSD occurs at  $\sim 1 \mu\text{m}$  in quartz, while it occurs at  $\sim 2 \mu\text{m}$  in feldspar (Table 4).

586

#### 587 **5.4 Natural fault rock**

588

589 The range of values in the natural samples from the Nojima fault is approximately the same as  
590 those of the feldspars in experiments (Figure 6, 8, Table 4) so that the natural fault rock  
591 appears to behave in a similar way as feldspar. The granodiorite deformed by the Nojima fault  
592 consists of  $\sim 53\%$  feldspar (plagioclase and K-feldspar); therefore, feldspar is expected to play  
593 a dominant role in the GSD of a mixed gouge.

594

595 The recently formed fault rock of the Nojima Fault zone shows the same  $D$ -values as the  
596 experimentally deformed granitoid rock, especially for feldspar minerals. If we compare the  
597  $D$ -values of the recent deformation episode with feldspar in experiment 64nk and 38nk,  
598 performed at  $T = 300^\circ\text{C}$ ,  $P_c = 500 \text{ MPa}$ ,  $\dot{\epsilon} = 1.5 \cdot 10^{-4} \text{ s}^{-1}$ , with  $< 4 \text{ mm}$  shortening and with 0.2

599 wt% H<sub>2</sub>O added to the sample, all D-values are the same within the error of 0.05.  
600 Temperature and pressure conditions during Nojima deformation were lower than in the  
601 deformation experiments.  
602  
603 D<sub>></sub> of the Nojima gouge (D<sub>></sub> = 2.02) is in fairly good agreement with data from coarser grain  
604 size fractions by Monzawa and Otsuki (2003); they obtain D<sub>></sub> = 2.192 to 2.559, for five  
605 different samples of the surface outcrop of the Nojima Fault Zone, with a mean of 2.347.  
606 Monzawa and Otsuki's (2003) data is more similar to the values of D<sub>></sub> for the experimentally  
607 deformed quartz so that all data appears to be within the variation of the experimental data.  
608  
609 Our results are consistent with studies of other fault zones (Table 1). Blenkinsop (1991) and  
610 An and Sammis (1994) describe both low (D<sub>></sub> < 1.6) and high (D<sub>></sub> > 2.0) values for analyses  
611 of gouge originating from the same fault zone. Several other studies describe high D<sub>></sub> values  
612 (e.g. Blenkinsop, 1991; Hadizadeh and Johnson, 2003). In some gouge the lower limit of D<sub>></sub>  
613 is 1.4 to 1.6, except for the San Andreas Fault Cajon Pass Drillhole granite (D = 0.8;  
614 Blenkinsop, 1991). We speculate that these gouge with D<sub>></sub> below 1.6 might be monomineralic  
615 gouge where fluids and temperature caused healing (e.g. Sammis et al., 1987, Keulen et al.,  
616 submitted) or were only cracked (e.g. Marone and Scholz, 1989).

617

## 618 **5.5. Surface densities of fault rocks**

619

620 The surface energy in gouge can be used to calculate a part of the energy release during an  
621 earthquake event (Kanamori, 1994; Wilson et al., 2005; Chester et al., 2005; Ma et al., 2006).  
622 The grain size distribution allows the calculation of the total surface area of the fault rock  
623 (Chester et al. 2005, Wilson et al. 2005). The estimates of Chester et al. (2005) for the  
624 Punchbowl fault are based on GSDs with an assumed minimum grain size, r<sub>min</sub>, of 1.6 nm and  
625 a constant D-value of 3 (= 2 in 2 dimensions) for the entire grain size range. In contrast, we  
626 find that in all of our experimental and natural samples the D-value below r ≈ 1 to 2 μm is D<sub><</sub>  
627 ≈ 1.

628

629 If the evolution of the GSD is a two-stage process involving initial rupturing and further grain  
630 comminution in the gouge, probably by attrition, shearing, grinding, etc.; it is questionable  
631 whether the surface energy calculated from the GSD can directly be related to the seismic  
632 energy release as proposed by Wilson et al. (2005) and Chester et al. (2005). Neither is the

633 surface energy obtained from the GSD of fault rocks identical to the specific surface energy  
634 as applied in earthquake energy budget calculations (e.g. Ventakaraman & Kanamori, 2004).  
635 For the following calculation we use the surface density of cracked grains, but not of gouge,  
636 to compare to the data of Wilson et al. (2005) and Chester et al. (2005), because we assume  
637 that this is the material, which is created by the rupture process during the seismic event,  
638 while the higher surface density of gouge is due to the combined effect of fracturing and  
639 subsequent deformation during slip.

640

641 To estimate the surface density for our cracked minerals and gouge we use the concept of the  
642 fractal fragmentation of the cube and a simple spreadsheet program. We set up a histogram of  
643 13 (logarithmic) bins, extending from 0.03  $\mu\text{m}$  to 128  $\mu\text{m}$  and from 0.015  $\mu\text{m}$  to 64  $\mu\text{m}$ ,  
644 respectively. Between  $r_{\min}$  and  $r_K$ , the 3-dimensional fractal dimension is set to 2 in both  
645 cases, corresponding to a value of  $D_{<} = 1.0$ . Between  $r_K$  and  $r_{\max}$ , the fractal dimension is set  
646 to 2.5 for cracked material and to 3 for gouge, corresponding to  $D_{>} = 1.50$  and  $D_{>} = 2.00$ ,  
647 respectively. A value of  $r_K = 1 \mu\text{m}$  is assumed in both cases. For each of the bins, the number,  
648 total volume and total surface of the cubes is calculated. Summing over all bins, the surface to  
649 volume ratio of cracked material and gouge is 1.3 and 6.0  $\mu\text{m}^2/\mu\text{m}^3$  i.e.,  $1.3 \cdot 10^6 \text{ m}^{-1}$  and  
650  $6.0 \cdot 10^6 \text{ m}^{-1}$  for cracked material and gouge.

651

652 Using the same approach to calculate surface densities for the Punchbowl Fault using the data  
653 ( $r_{\min} = 1.6 \text{ nm}$  and  $r_{\max} = 100 \mu\text{m}$ ,  $D = 2.0$ ) given by Chester et al. (2005), we obtain  $6.8 \cdot 10^8$   
654  $\text{m}^{-1}$ . If we now compare our value of cracked grains of  $1.3 \cdot 10^6 \text{ m}^{-1}$  to a value of  $4.2 \cdot 10^8 \text{ m}^{-1}$   
655 (corresponding to  $4.2 \cdot 10^5 \text{ m}^2$  in the 1mm thick layer of the fault, given by Chester et al.,  
656 2005) or  $2.7 \cdot 10^8 \text{ m}^{-1}$  (corresponding to  $80 \text{ m}^2 \text{ g}^{-1}$  given by Wilson et al., 2005) we note that in  
657 our case, the surface density, and accordingly, the energy required to generate it, is smaller  
658 than the values assumed by Wilson et al. (2005) and Chester et al. (2005) by a factor of 200 to  
659 300. Finding that the surface densities in cracked rocks and even in gouge are so low suggests  
660 to us that the creation of those surfaces cannot play a major role in the energy budget of a  
661 seismic event.

662

663 The discrepancy between our data and the data published by Chester et al. (2005) can be  
664 explained by the fact that they have measured their small grain size fraction (down to  $r = 25$   
665 nm) in the TEM, which is a notoriously difficult method for obtaining good statistics of any  
666 parameter observed. Thus, they did not detect the change of D-values at  $r \approx 1 \mu\text{m}$ . Using  $D =$

667 2.0 over the entire range of grain sizes results in a total surface area which is 2 orders of  
668 magnitude too high. Furthermore, they assume  $r_{\min} = 1.6$  nm and thus many tiny grains with a  
669 high surface area density. Note that finding lower surface densities rather supports and  
670 strengthens their general conclusion that the surface energies do not contribute in any major  
671 way to the energy dissipated during earthquakes. Pittarello et al. (2006) come to the same  
672 conclusion, using GSD measurements of pseudotachylyte-related granitoid cataclasites.

673

## 674 **5.6 Grinding limit**

675

676 Grains with very small radii formed by comminution have been found before in deformation  
677 experiments ( $d = 10$  nm; Yund et al., 1990), in mining induced fault zones ( $\sim 40$  nm; Olgaard  
678 and Brace, 1983; Wilson et al., 2005) and in natural fault zones (40 nm; Wilson et al., 2005,  
679 Chester et al. 2005). We observe grains of 30 nm in both experimental and natural fault rock  
680 (Figure 9).  $r_K$  is  $1.2 \pm 0.3$   $\mu\text{m}$  for quartz and  $1.8 \pm 0.3$   $\mu\text{m}$  for feldspar grains. The slope  
681 change is observed for both cracked grains and gouge (Figure 6, 8, Table 5). Very small  
682 grains are already formed upon cracking of the grains; they are not only a product of gouge  
683 formation.  $D_<$  is less dependent on deformation conditions, the type of mineral (quartz or  
684 feldspar) and is more similar for cracked grains and gouge than  $D_>$ .

685

686 Kendall (1978) has demonstrated that a critical radius exists below which particles cannot be  
687 comminuted further by grinding. This critical radius, called the grinding limit, is 0.3  $\mu\text{m}$  for  
688  $\text{SiO}_2$  glass, 0.9  $\mu\text{m}$  for quartz and 2.2  $\mu\text{m}$  for calcite (Steier and Schönert, 1972 in Prasher,  
689 1987). Small particles have a different internal stress distribution and are stronger due to  
690 smaller flaw sizes and flaw size density (Prasher, 1987) so that very high stresses are required  
691 to break them (Mitra, 1978). This is consistent with our observation that no intragranular  
692 cracks within grains smaller than 1  $\mu\text{m}$  in radius are observed.

693

694 The  $r_K$ -values for quartz are slightly higher than the grinding limit (given as radius) for quartz  
695 (Figure 10). There is no data for the feldspar grinding limit available but the higher  $r_K$ -values  
696 for feldspar (Figure 10) are consistent with the observation that the critical grain size for  
697 grinding is linearly dependent on the elastic modulus  $E$  of the mineral (Kendall, 1978),  
698 because the bulk  $E$ -modulus of feldspar is approximately 2 times that of quartz (Bass, 1995).

699

700 Thus, it appears likely that  $r_K$  represents approximately the grinding limit in the comminution  
701 of fault rock. For smaller grain sizes, where grinding is inhibited, comminution is less  
702 efficient, leading to a lower  $D_c$ . Grinding is performed by compressional fracturing of grains:  
703 a network of cracks will develop and the grain will break into several smaller ones (Jaeger,  
704 1967). Other processes, like shearing or attrition also produce comminution. Reches and  
705 Dewers (2005) and Dor et al. (2006) describe comminution as dynamic pulverization  
706 (explosive granulation followed by dynamic contraction) at earthquake crack tips. For grains  
707 smaller than  $r_K$  compressional fracturing appears inhibited, but shearing or attrition remains  
708 significant. From the difference between  $D$ -values in cracked grains and gouge, it is clear that  
709 grinding, shearing and attrition during slip cause further comminution in gouge.

710

## 711 **5.7 Fractal dimension**

712

713 Fractal dimensions are used to describe geometries that are scale invariant (Turcotte, 1992).  
714 To describe a GSD as fractal, this distribution should lie on a straight line over several orders  
715 of magnitude in a  $\log(\text{frequency})$ - $\log(\text{size})$  distribution. The GSD of both the experiments and  
716 the Nojima Fault Zone gouge does not show a fractal distribution. The data cover 3.5 orders  
717 of magnitude, but cannot be fitted with a single straight line. Furthermore, there are statistical  
718 problems because most fault rock grain size measurements only cover 1 to 4 orders of  
719 magnitude.

720

721 The advantage of having a fractal distribution is the scale invariance of the measurements.  
722 Analyses performed on laboratory experiments or on hand specimens could easily be  
723 extrapolated to large fault systems. However, Turcotte (1992) emphasizes that not every  
724 power-law distribution can be described by a fractal dimension. Fractal dimensions are  
725 limited to  $D$ -values between 0 and 2 in 2D (Turcotte, 1992), so that gouge with  $D_c$  larger than  
726 2.0 cannot be described as fractal.

727

728 Several authors have commented on the ideal nature of a fractal dimension or  $D$ -value of 1.6:  
729 a GSD with a  $D$ -value of 1.6 appears the most ideal for close packing (Monzawa and Otsuki,  
730 2003). Hoffmann and Schönert (1971) and Sammis et al. (1987) describe a  $D$ -value of 1.6 (for  
731 constrained comminution) as ideal, because the probability that grains of the same size are  
732 neighboring each other and the amount of pore space for a close-packed array of spherical  
733 grains are minimized. Morgan (1999) and Morgan and Boettcher (1999) have shown with

734 numerical simulations of gouge that a  $D$  of 1.6 marks the onset of shear localization, inter-  
735 particle rolling, and a decrease in the general comminution rate. Biegel et al. (1989) and  
736 Marone and Scholz (1989) find a  $D$ -value of 1.6 for the onset of shear localization in their  
737 experimentally formed gouge. It seems that the onset of gouge evolution from cracked grains,  
738 leading to an increase of  $D_{>}$  above 1.6, causes dilatancy (less close packing), which will  
739 increase the possibility for grinding, creating more small grains and thus a higher  $D_{>}$ .  
740 Dilatancy may allow for inter-particle rolling; in this way smaller and rounder grains are  
741 formed, therefore  $D_{>}$  is increased and more grains with  $r \leq r_K$  are formed. The smaller  $D$ -  
742 values of gouge in high confining pressure experiments support such an interpretation,  
743 because of the effective suppression of dilatancy at high pressures.

744

745 Storti et al. (2003) report measurements in carbonate fault zones, in which the  $D$ -values at the  
746 boundary of the shear zone are between 0.9 and 1.4 and evolve to  $D$ -values between 1.6 and  
747 2.5 for the interior shear bands. These observations are consistent with our observations:  
748 gouge ( $D > 1.6$ ) is localized in slip zones, whereas cracked grains ( $D < 1.6$ ) make up a larger  
749 part of the sample without major slip.

750

751

## 752 **6. Conclusions**

753

754 The analyzed fault rocks derived from experimentally faulted granite and from the Nojima  
755 fault zone consist of two different types of material, cracked grains and gouge. These can be  
756 distinguished on the basis of their microstructures and their grain size distribution (GSD). On  
757 log-log plots the GSDs display two distinct power-law fits:  $D_{<}$  for grain sizes smaller than  $r_K$   
758 and  $D_{>}$  for grain sizes larger than  $r_K$ .  $D_{>}$  is the value that corresponds to other published  $D$ -  
759 values (so-called "fractal dimensions").

760

761  $D_{<}$  is 0.9-1.0 for cracked grains and 0.9-1.1 for gouge and more or less independent of the  
762 deformation conditions or the type of mineral. The lower limit of grain sizes in  
763 experimentally and naturally produced gouge is  $r = 15$  nm.

764

765  $D_{>}$  is 1.5-1.6 for cracked grains and 2.0-2.3 for gouge and depends on the deformation  
766 conditions or the type of mineral. The upper limit of grain sizes evaluated in this study,  $r_{\max}$ , is



767 100  $\mu\text{m}$ .  $D_{>}$  for gouge is a good parameter to compare natural and experimental fault rock of  
768 grain sizes smaller than  $\sim 100 \mu\text{m}$ .

769

770 Cracked grains result from initial fragmentation by rupturing. They develop into gouge by  
771 subsequent grain comminution, grinding, attrition, or shear during slip along the fault zone.

772 These processes produce larger  $D_{>}$ -values and therefore represent a more efficient grain size  
773 reduction for quartz and feldspar grains in the size range greater than  $1\text{-}2 \mu\text{m}$ .

774

775 The grain size,  $r_K$ , at the slope break of the  $\log(\text{frequency})\text{-}\log(\text{size})$  histogram of the GSD  
776 occurs at  $1.2 \pm 0.3 \mu\text{m}$  for quartz and at  $1.8 \pm 0.3 \mu\text{m}$  for feldspar. The grain size  $r_K$  coincides  
777 approximately with the grinding limit in quartz and probably corresponds to a change in the  
778 physical process of grain comminution. Attrition and shear may dominate below the transition  
779 value  $r_K$ .

780

781 Most of the experimental conditions during deformation (temperature, confining pressure,  
782  $\text{H}_2\text{O}$ -content, displacement rate) have a minor effect on the GSD. Most noticeable is an  
783 increase in confining pressure ( $\sim 1 \text{ GPa}$ ), which reduces the efficiency of grain comminution.

784

785 The gouge of the Nojima Fault Zone shows  $D_{>} = 2.02$  for gouge and  $1.64$  for cracked grains.  
786 These values are the same as for the experimentally deformed granitoids.

787

788 Experiments and natural seismic fault rocks from the Nojima fault zone show the same GSD,  
789 although the experiments have been carried out at higher PT-conditions and at considerably  
790 slower displacement rates than in nature.

791

792 The development of the GSD is a two-stage process. First rupturing of the rock causes  
793 cracked grains that evolve to a gouge as a result of further movement on the fault zone. The  
794 surface densities calculated from the GSD of cracked minerals and gouge are  $200\text{-}300$  times  
795 lower than published data, confirming that the creation of surface plays a minor role in the  
796 energy budget of earthquakes.

797

798

799

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801

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812

813 **References**

- 814 Amitrano, D., Schmittbuhl, J., 2002. Fracture roughness and gouge distribution of a granite shear band. *Journal*  
815 *of Geophysical Research* 107, 2375–2390.
- 816
- 817 An, L.-J., Sammis, C., 1994. Particle size distribution of cataclastic fault materials from southern California: A  
818 3-D study. *Pure & Applied Geophysics* 143, 203–227.
- 819
- 820 Anderson, J., Osborne, R., Palmer, D., 1980. Petrogenesis of cataclastic rocks within the San Andreas Fault  
821 Zone of southern California, U.S.A. *Tectonophysics* 67, 221–249.
- 822
- 823 Anderson, J., Osborne, R., Palmer, D., 1983. Cataclastic rocks of the San Gabriel Fault- An expression of  
824 deformation at deeper crustal levels in the San Andreas Fault Zone. *Tectonophysics* 98, 209–251.
- 825
- 826 Bass, J., 1995. Elasticity of Minerals, Glasses, and Melts. In: Ahrens, T. (Ed.), *Mineral Physics &*  
827 *Crystallography - A handbook of Physical Constants*, AGU Reference Shelf 2, 45–63.
- 828
- 829 Biegel, R., Sammis, C., Dietrich, J., 1989. The frictional properties of a simulated gouge having a fractal  
830 particle distribution. *Journal of Structural Geology* 11, 827–846.
- 831
- 832 Blanpied, M., Lockner, D., Byerlee, J., 1995. Frictional slip of granite at hydrothermal conditions. *Journal of*  
833 *Geophysical Research* 100, 13045–13064.
- 834
- 835 Blenkinsop, T., 1991. Cataclasis and Processes of Particle Size Reduction. *Pure & Applied Geophysics* 136,  
836 60–86.
- 837
- 838 Blenkinsop, T., Rutter, E., 1986. Cataclastic deformation of quartzite in the Moine thrust zone. *Journal of*  
839 *Structural Geology* 8, 669–681.
- 840
- 841 Bos, B., Peach, C., Spiers, C., 2000. Slip behaviour of simulated gouge-bearing faults under conditions  
842 favoring pressure solution. *Journal of Geophysical Research* 105, 16699–16717.
- 843
- 844 Boullier, A.-M., Fujimoto, K., Ito, H., Ohtani, T., Keulen, N., Fabbri, O., Amitrano, D., Dubois, M., Pezard, P.,  
845 2004a. Structural evolution of the Nojima Fault (Awaji Island, Japan) revisited from the GSJ drillhole at  
846 Hirabayashi. *Earth Planets Space* 56, 1233–1240.
- 847
- 848 Boullier, A.-M., Fujimoto, K., Ohtani, T., Roman-Ross, G., Lewin, E., Ito, H., Pezard, P., Ildefonse, B., 2004b.  
849 Textural evidence for recent co-seismic circulation of fluids in the Nojima fault zone, Awaji island, Japan.  
850 *Tectonophysics* 378, 165-181.
- 851
- 852 Byerlee, J., Brace, W., 1968. Stick slip, stable sliding, and earthquakes - effect of rock type, pressure, strain  
853 rate, and stiffness. *Journal of Geophysical Research* 73, 6031–6037.
- 854
- 855 Carter, N., Anderson, D., Hansen, F., Kranz, R., 1981. Creep and creep rupture of granitic rocks. In: Carter, N.,  
856 Friedman, M., Logan, J., Stearns, D. (Eds.), *Geophysical Monograph: Mechanical behavior of crustal rocks; the*  
857 *Handin volume*, 24, American Geophysical Union, Washington DC, 61–62.
- 858
- 859 Chester, F., Friedman, M., Logan, J., 1985. Foliated cataclasites. *Tectonophysics* 111, 139–146.
- 860
- 861 Chester, J., Chester, F., Kronenberg, A., 2005. Fracture surface energy of the Punchbowl fault, San Andreas  
862 system. *Nature* 437, 133–136.
- 863
- 864 De Ronde, A., 2004. Mineral Reaction and Deformation in Plagioclase-Olivine Composites: An Experimental  
865 Study. PhD thesis, Basel University.
- 866

867 Dor, O., Ben-Zion, Y., Rockwell, T.K. Brune, J., 2006. Pulverized rocks in the Mojave section of the San  
868 Andreas Fault Zone. *Earth and Planetary Science Letters* 245, 642-654.

869  
870 Fabbri, O., Iwamura, K., Matsunaga, S., Coromina, G., Kanaori, Y., 2004. Distributed strike-slip faulting,  
871 block rotation, and possible intracrustal vertical decoupling in the convergent zone. In Grocott, J., Tikoff, B.,  
872 McCaffrey, K., Taylor, G. (Eds.), *Vertical coupling and decoupling in the lithosphere*. Geological Society  
873 London Special Publications, London. 227, 141-166.

874  
875 Faulkner, D., Lewis, A., Rutter, E., 2003. On the internal structure and mechanics of large strike-slip fault  
876 zones: field observations of the Carboneras fault in southeastern Spain. *Tectonophysics* 367, 235-251.

877  
878 Griggs, D., Turner, F., Heard, H., 1960. Deformation of rocks at T = 500 °C to T = 800 °C. In Griggs, D.  
879 Handin, J. (Eds.), *Geological Society of America Memoir*, 79, Waverly Press, Baltimore MD, 39-104.

880  
881 Hadizadeh, J., Johnson, W., 2003. Estimating local strain due to comminution in experimental cataclastic  
882 textures. *Journal of Structural Geology* 25, 1973-1979.

883  
884 Heilbronner, R., Keulen, N., 2006. Grain size and grain shape analysis of fault rocks. *Tectonophysics* 427, 199-  
885 216 doi:10.1016/j.tecto.2006.05.020.

886  
887 Hirth, G., Tullis, J., 1994. The brittle-plastic transition in experimentally deformed quartz aggregates. *Journal*  
888 *of Geophysical Research* 99, 11731-11747.

889  
890 Hoffmann, N., Schönert, K., 1971. Bruchanteil von Glaskugeln in Packungen von Fraktionen und binären  
891 Mischungen. *Aufbereitungstechnik* 9, 513-518.

892  
893 Ito, H., Fujimoto, K., Tanaka, H., Lockner, D., 1999. Proceedings of the international workshop on the Nojima  
894 fault core and borehole data analysis, Geological Survey of Japan Interim report No. EQ/00/1, USGS Open-File  
895 Report 000-129.

896  
897 Jaeger, J., 1967. Failure of rocks under tensile conditions. *International Journal of Rock Mechanics and Mining*  
898 *Sciences* 4, 219-237.

899  
900 Kanamori, H., 1994. Mechanics of Earthquakes. *Annual Review of Earth and Planetary Sciences* 22, 207-237.

901  
902 Kato, A., Ohnaka, M., Mochizuki, H., 2003. Constitutive properties for the shear failure of intact granite in  
903 seismogenic environments. *Journal of Geophysical Research* 108, doi:10.1029/2001JB000791.

904  
905 Kendall, K., 1978. The impossibility of comminuting small particles by compression. *Nature* 272, 710-711.

906  
907 Korvin, G., 1992. *Fractal Models in the Earth Sciences*. Elsevier, Amsterdam.

908  
909 Lin, A., 1999. S-C cataclasite in granitic rock. *Tectonophysics* 304, 57-273.

910  
911 Lockner, D., Summers, R., Byerlee, J., 1986. Effects of temperature and sliding rate on frictional strength of  
912 granite. *Pure & Applied Geophysics* 124, 445-469.

913  
914 Ma, K.-F., Tanaka, H., Song, S.-R., Wang, C.-Y., Hung, J.-H., Tsai, Y.-B., Mori, J., Song, Y.-F., Yeh, E.-C.,  
915 Soh, W., Sone, H., Kuo, L.-W., Wu, H.-Y., 2006. Slip zone and energetics of a large earthquake from the  
916 Taiwan Chelungpu-fault Drilling Project. *Nature* 444, doi:10.1038/nature05253.

917  
918 Mair, K., Frye, K., Marone, C., 2002. Influence of grain characteristics on the friction of granular shear zones.  
919 *Journal of Geophysical Research* 107, ECV 4-1, doi:2001JB000516.

920  
921 Mandelbrot, B., 1982. *The fractal Geometry of Nature*. Freeman and Co., New York.

922  
923 Marone, C., Scholz, C., 1989. Particle-size distribution and microstructures within simulated gouge. *Journal of*  
924 *Structural Geology* 11, 799–814.  
925  
926 Mitra, G., 1978. Ductile deformation zones and mylonites. the mechanical processes involved in the  
927 deformation of crystalline basement rocks. *American Journal of Science* 278, 1057–1084.  
928  
929 Mitra, G., 1993. Deformation processes in brittle deformation zones in granitic basement rocks: A case study  
930 from the Torrey Creek area, Wind River Mountains. In: Schmith, C., Chase, R., Erslev, E. (Eds.), *Laramide*  
931 *Basement Deformation in the Rocky Mountains Foreland of the Western United States*. Geological Society of  
932 *America Special Paper* 280, 177–195.  
933  
934 Monzawa, N., Otsuki, K., 2003. Comminution and fluidization of granular fault materials: implications for  
935 fault slip behavior. *Tectonophysics* 367, 127–143.  
936  
937 Morgan, J., 1999. Numerical simulations of granular shear zones using the distinct element method. 2. Effects  
938 of particle size distribution and interparticle friction on mechanical behavior. *Journal of Geophysical Research*  
939 104, 2721–2732.  
940  
941 Morgan, J., Boettcher, M., 1999. Numerical simulations of granular shear zones using the distinct element  
942 method. 1. Shear zone kinematics and the micromechanics of localization. *Journal of Geophysical Research*  
943 104, 2703–2719.  
944  
945 Morrow, C., Byerlee, J., 1989. Experimental studies of compaction and dilatancy during frictional sliding on  
946 faults containing gouge. *Journal of Structural Geology* 11, 815–826.  
947  
948 Murakami, M., Tagami, T., 2004. Dating pseudotachylite of the Nojima fault using the zircon fission-track  
949 method. *Geophysical Research Letters* 31, doi:10.1029/2004GL020211.  
950  
951 Murakami, M., Tagami, T., Hasebe, N., 2002. Ancient thermal anomaly of an active fault system: zircon  
952 fission-track evidence from Nojima GSJ 750m borehole samples. *Geophysical Research Letters* 29,  
953 doi:10.1029/2002GL015679.  
954  
955 Murata, A., Takemura, K., Miyata, T., Lin, A., 2001. Quaternary vertical offset and average slip rate of the  
956 Nojima Fault on Awaji Island, Japan. *The Island Arc* 10, 360–367.  
957  
958 Ohtani, T., Fujimoto, K., Ito, H., Tanaka, H., Tomida, N., Higuchi T., 2000. Fault rocks and past to recent fluid  
959 characteristics from the borehole survey of the Nojima fault ruptured in the 1995 Kobe earthquake, southwest  
960 Japan. *Journal of Geophysical Research* 105, 16161–16172.  
961  
962 Olgaard, D., Brace, W., 1983. The microstructure of gouge from a mining-induced seismic shear zone.  
963 *International Journal of Rock Mechanics and Mining Science & Geomechanical Abstracts* 20, 11–19.  
964  
965 Otsuki, K., Monzawa, N., Nagase, T., 2003. Fluidization and melting of gouge during seismic slip:  
966 Identification in the Nojima fault zone and implications for focal earthquake mechanisms. *Journal of*  
967 *Geophysical Research* 108, ESE 4–1 – 4–18.  
968  
969 Pittarello, L., Di Toro, G., Bizzarri, A., Hadizadeh, J., Pennacchioni, G., 2006. Seismic energy partitioning  
970 inferred from pseudotachylite-bearing faults (Gole Larghe Fault, Adamello batholith, Italy). *EGU, Vienna*.  
971 *Geophysical Research Abstracts* 8, EGU06–A–01883.  
972  
973 Prasher, C., (1987). *Crushing and Grinding Process Handbook*. Wiley, New York.  
974  
975 Reches, Z., Dewers, T., 2005. Gouge formation by dynamic pulverization during earthquake rupture. *Earth and*  
976 *Planetary Science Letters* 235, 361–374.

977  
978 Sammis, C., Biegel, R., 1989. Fractals, fault-gouge, and friction. *Pure & Applied Geophysics* 131, 255–271.  
979  
980 Sammis, C., King, G., Biegel, R., 1987. The kinematics of gouge deformation. *Pure & Applied Geophysics*  
981 125, 777–812.  
982  
983 Sammis, C., Osborne, R., Anderson, J., Banerdt, M., White, P., 1986. Self-similar cataclasis in the formation of  
984 gouge. *Pure & Applied Geophysics* 124, 53–78.  
985  
986 Schmid, S., Handy, M., 1991. Towards a genetic classification of fault rocks: Geological usage and  
987 tectonophysical implications. In: Müller, D., McKenzie, J., Weissert, H. (Eds.), *Controversies in Modern*  
988 *Geology*, Academic Press, London, 339–361.  
989  
990 Shao, S.-M., Zou, J.-C., 1996. Fractal research of gouge. *Acta Seismologica Sinica* 9, 485–491.  
991  
992 Sibson, R., 1977. Fault rocks and fault mechanisms. *Journal of the Geological Society of London* 133, 191–  
993 213.  
994  
995 Stesky, R., Brace, W., Riley, D., Robin, P.-Y., 1974. Friction in faulted rock at high temperature and pressure.  
996 *Tectonophysics* 23, 177–203.  
997  
998 Stewart, M., Holdsworth, R., Strachan, R., 2000. Deformation processes and weakening mechanisms within the  
999 frictional-viscous transition zone of major crustal-scale faults: insights from the Great Glen fault zone,  
1000 Scotland. *Journal of Structural Geology* 22, 543–560.  
1001  
1002 Storti, F., Billi, A., Salvini, F., 2003. Particle size distributions in natural carbonate fault rocks; insights for  
1003 non-self-similar cataclasis. *Earth and Planetary Science Letters* 206, 173–186.  
1004  
1005 Tanaka, H., Fujimoto, K., Ohtani, T., Ito H., 2000. Structural and chemical characterization of shear zones in  
1006 the freshly activated Nojima fault, Awaji Island, southwest Japan. *Journal of Geophysical Research* 106, 8789–  
1007 8810, 10.1029/2000JB900444.  
1008  
1009 Tullis, J., Yund, R., 1977. Experimental deformation of dry Westerly Granite. *Journal of Geophysical Research*  
1010 82, 5705–5718.  
1011  
1012 Tullis, J., Yund, R., 1980. Hydrolytic weakening of experimentally deformed Westerly Granite and Hale Albite  
1013 rock. *Journal of Structural Geology* 2, 439–451.  
1014  
1015 Tullis, J., Yund, R., 1987. Transition from cataclastic flow to dislocation creep of feldspar: Mechanisms and  
1016 microstructures. *Geology* 15, 606–609.  
1017  
1018 Tullis, J., Yund, R., 1992. The brittle-ductile transition in feldspar aggregates: An experimental study. In: Evans,  
1019 B., Wong, T.-f. (Eds.), *Fault Mechanics and transport properties of Rocks*, Academic Press, San Diego, 89–  
1020 117.  
1021  
1022 Tullis, T., Tullis, J., 1986. Experimental rock deformation techniques. In: Hobbs, B. Heard, H. (Eds.), *AGU*  
1023 *Geophysical Monograph, Mineral and rock deformation: Laboratory Studies - The Paterson Volume* 36, 297–  
1024 324.  
1025  
1026 Turcotte, D., (1992). *Fractals and chaos in geology and geophysics*. University Press, Cambridge.  
1027  
1028 Ventakaraman, A., Kanamori, H., 1992. Observational constraints on the fracture energy of subduction zone  
1029 earthquakes. *Journal of Geophysical Research* 109, B05302, doi:10.1029/2003JB002549.  
1030  
1031 White, S., 1982. *Fault rocks of the Moine thrust zone: a guide to their nomenclature*. Textures and

1032 microstructures 4, 211–221.  
1033  
1034 Wibberley, C., Shimamoto, T., 2005. Earthquake slip weakening and asperities explained by thermal  
1035 pressurisation. *Nature* 436, 689–692.  
1036  
1037 Wilson, B., Dewers, T., Reches, Z., Brune, J., 2005. Particle size and energetics of gouge from earthquake  
1038 rupture zones. *Nature* 434, 749–752.  
1039  
1040 Wilson, J., Chester, J., Chester, F., 2003. Microfracture analysis of fault growth and wear processes,  
1041 Punchbowl Fault, San Andreas system, California. *Journal of Structural Geology* 25, 1855–1873.  
1042  
1043 Wise, D., Dunn, D., Engelder, J., Geiser, P., Hatcher, R., Kish, S., Odom, A., Schamel, S., 1984. Fault-related  
1044 rocks: Suggestions for terminology. *Geology* 12, 391–394.  
1045  
1046 Yund, R., Blanpied, M., Tullis, T., Weeks, J., 1990. Amorphous material in high strain experimental gouge.  
1047 *Journal of Geophysical Research* 95, 15589–15602.  
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1050

## 1051 **Figure captions & Table captions**

1052

1053 Figure 1:

1054 Starting material for experiments and natural fault rocks used in study.

1055 (a) Undeformed sample material from the Val Verzasca, Switzerland; scanning electron micrographs,  
1056 back scatter contrast. This gneiss is isotropic at the scale of a thin section and has a granite  
1057 composition. L = lineation. Minerals, in order of increasing brightness: quartz (Qtz), Na-rich  
1058 plagioclase (Plg), muscovite (Musc), K-feldspar (Kfs), biotite (Bt).

1059 (b) Map of the northern part of Awaji Island, Japan, showing the locality of the Hirabayashi drill site  
1060 in the Nojima Fault [after Ohtani et al. (2000)]. Inset shows the cities Tokyo and Kobe (stars) and the  
1061 enlarged area (black square).

1062 (c, d) Fault rocks of the Nojima Fault Zone (recent event); scanning electron micrographs, back  
1063 scatter contrast. (c) Gouge with rounded siderite fragments (white arrows) and pseudotachylyte  
1064 fragment (black arrow). 623.50 m depth. (d) Cracked grains: individual parts can still be fitted  
1065 together. The white arrow indicates siderite grains. 633.08 m depth.

1066

1067 Figure 2:

1068 Schematic of procedure for grain size analysis.

1069 (a) Starting image: gouge of Nojima Fault Zone; scanning electron micrographs, backscatter contrast.

1070 (b) Binary image. (c) Histogram of equivalent radii (1285 grains). (d) Log-log plot of frequency

1071 versus equivalent radius (20 bins per order of magnitude). The slope of the line fit yields the D-value.

1072 (e) Combined log-log plot for entire range of magnification, showing two distinct average D-values  
1073 ( $D_{<}$  and  $D_{>}$ ) and  $r_K$  (see text). (f) D-values versus magnification. Note that intermediate magnifications  
1074 yield both D-values ( $D_{<}$  and  $D_{>}$ ).

1075

1076 Figure 3:

1077 Mechanical data.

1078 (a) Force versus axial shortening of samples of Verzasca Gneiss. For experimental conditions see  
1079 Table 3, r-h = room-humidity. (b) Differential stress versus axial shortening, calculated using  
1080 program rigC4 (see text). For comparison, 3 experiments on Westerly Granite by Tullis and Yund  
1081 (1977) are included. Since pyrophyllite or talc was used as confining medium, the true  $P_c$ -values  
1082 (indicated in brackets) are approx. 2/3 of the values given in the literature (see text).

1083

1084 Figure 4:

1085 Maps of local variations of grain size distribution.

1086 The radius of the Gauss-filter was 5 pixels (= 40  $\mu\text{m}$ ); for D-mapping technique, see Heilbronner &  
1087 Keulen (2006). D-values > 1.75 indicate gouge, D-values < 1.75 indicate cracked grains. Holes and  
1088 biotite in samples are masked. Three samples are shown: 38nk (reference sample): d = 3 mm; 102nk:  
1089 d = 4.8 mm; 60nk: d = 2.5 mm. Fault surfaces are indicated. Scale applies to all samples.

1090

1091 Figure 5:

1092 Microstructures of experimentally deformed samples of Verzasca Gneiss.

1093 Scanning electron micrographs, backscatter contrast. Minerals, in order of increasing brightness:  
1094 quartz, Na-rich plagioclase, muscovite, K-feldspar, biotite.

1095 (a) Fault zone consisting of gouge with adjacent cracked grains (reference sample 38nk). (b)

1096 Relatively narrow gouge zone (sample 60nk, high confining pressure). (c) Reference sample 64nk. (d)

1097 Reference sample 38nk. (e) Reference sample 38nk. (f) Reference sample 38nk. (g) Cracked quartz

1098 grains at high magnification (high displacement sample 102nk). (h) Quartz gouge at high

1099 magnification (reference sample 38nk).

1100

1101 Figure 6:

1102 Grain size distribution of experimentally deformed Verzasca Gneiss plotted in a log(frequency)-

1103 log(equivalent radius) histogram. Results for quartz and feldspar are shown in separate graphs, results

1104 for cracked grains and gouge are shown with separate symbols. The frequency is normalized to 100

1105 grains with a radius of 10  $\mu\text{m}$  for cracked grains and to 1000 grains with a radius of 10  $\mu\text{m}$  for gouge.



1106 Note the slope change in most of the curves at  $r \approx 2 \mu\text{m}$ . The D-values ( $D_{<}$  and  $D_{>}$ ) and the grain size  
1107 at the slope change,  $r_K$ , are listed in the Table 4 and 5, respectively.

1108

1109 Figure 7:

1110 D-values of the gouge plotted against magnification of analyzed images.

1111 Average D-values are indicated by the horizontal lines. Feldspar and quartz are shown separately.

1112 Analyses at 1000x and 2000x magnification yield a value for  $D_{<}$  and  $D_{>}$ , because  $r_K$  lies in the range  
1113 of analyzed grain sizes .

1114

1115 Figure 8:

1116 Grain size distribution and D-values of the Nojima Fault Zone.

1117 (a) Grain size distribution of cracked grains and gouge (all minerals). The frequency is normalized to  
1118 100 grains with a radius of  $10 \mu\text{m}$  for cracked grains and to 1000 grains with a radius of  $10 \mu\text{m}$  for  
1119 gouge. (b) D-values versus magnification, compiled from 9 cascades (see text). The average values  
1120 per magnification are indicated with solid symbols. Overall average values for  $D_{<}$  and  $D_{>}$  are  
1121 indicated by solid lines. The thin-sections used for the analyses originate from the following locations  
1122 in the cores: a) HR4-1 (522.79 m), HR4-11 (633.08 m) b) HR4-11, HR2-52 (626.90 m), HR256  
1123 (643.10 m) and HR3-10 (623.46 m).

1124

1125 Figure 9:

1126 Scanning electron micrographs of the smallest grains found in gouge.

1127 (a,c) Nojima Fault Zone. (b,d) Experiment 63nk (room-humidity sample). (a,b) Secondary electron  
1128 mode. (c,d) backscatter contrast mode. The pockmark structure in (a) is an effect of the carbon coating  
1129 and does not show individual atoms or another substructure.

1130

1131 Figure 10:

1132 Grain size,  $r_K$ , at the transition from  $D_{<}$  to  $D_{>}$  (slope change in GSD plots) and grinding limit for

1133 quartz, after Steier & Schönert (1972) in Prasher (1987). The uncertainty in the position of  $r_K$  is  $\pm 0.3$   
1134  $\mu\text{m}$ .

1135

1136

1137 Table 1:  
1138 Published data on grain size analyses of fault rocks.  
1139 Method of analysis: LM = light microscope image of thin section, SEM = scanning electron  
1140 microscope image, manual = analogue box counting, digitizer = instrument for digital analysis of  
1141 photomicrographs, digital = analyzed with computer, Coult.count. = laser diffraction particle size  
1142 analysis. The indicated fractal dimension (D-value) is for 2-dimensional analysis. S.A.F. = San  
1143 Andreas Fault, R.S.A. = Republic of South Africa.

1144  
1145 Table 2:  
1146 Abbreviations and definitions of measurements as used in this paper.

1147  
1148  
1149 Table 3:  
1150 Deformation conditions of the experiments on Verzasca gneiss.  
1151 T = temperature,  $P_c$  = confining pressure,  $\dot{\epsilon}$  = axial shortening rate rate,  $L_0$  = starting length of  
1152 sample;  $d_{\text{chart}}$  = axial displacement from chart strip,  $d_{\text{chart} - 1 \text{ mm}}$  = corrected axial displacement,  $H_2O$  =  
1153 amount of added water. Samples were quenched immediately after deformation.

1154  
1155 Table 4:  
1156 D-values ( $D_<$  and  $D_>$ ) of cracked grains and gouge.  
1157 In experiments, quartz and feldspar were measured separately; in natural samples (Nojima), all  
1158 mineral grains were combined.

1159  
1160 Table 5:  
1161  $r_K$  values determined from grain size distributions.  
1162 In experiments, the values for quartz and feldspar were determined separately; in natural samples  
1163 (Nojima), all mineral grains were combined.

1164  
1165  
1166  
1167

Figure01  
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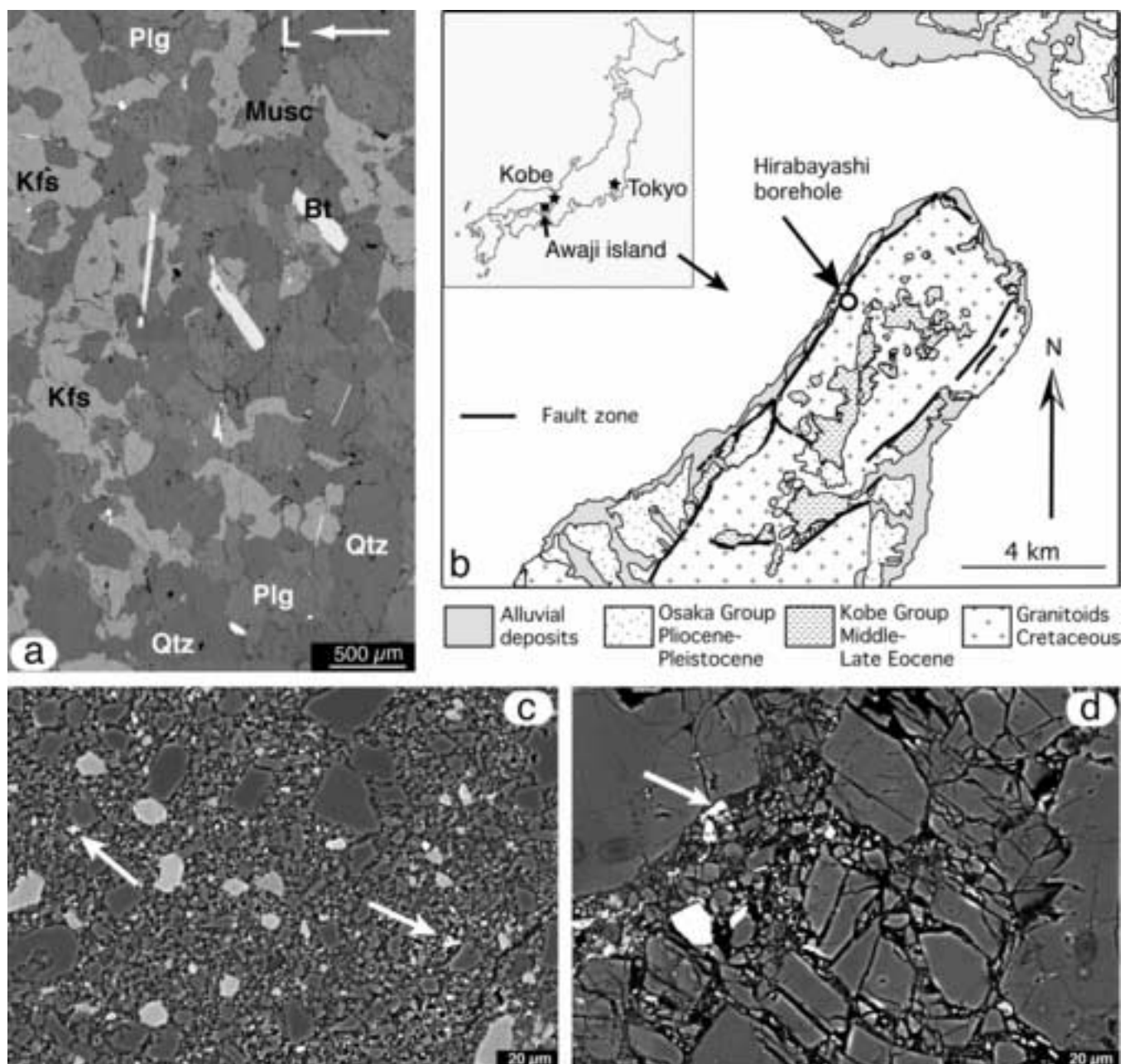


Figure02

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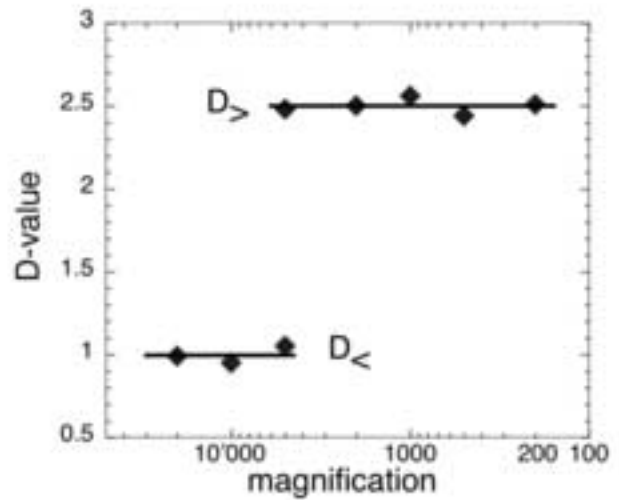
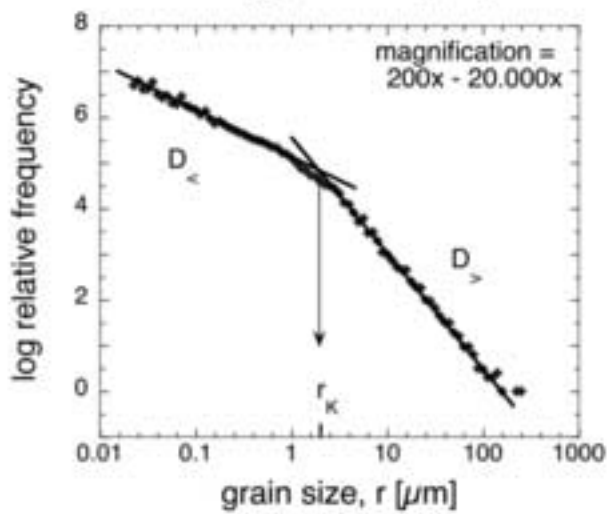
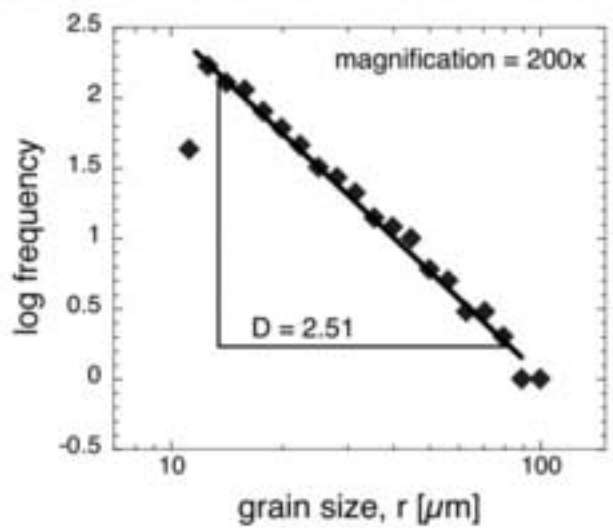
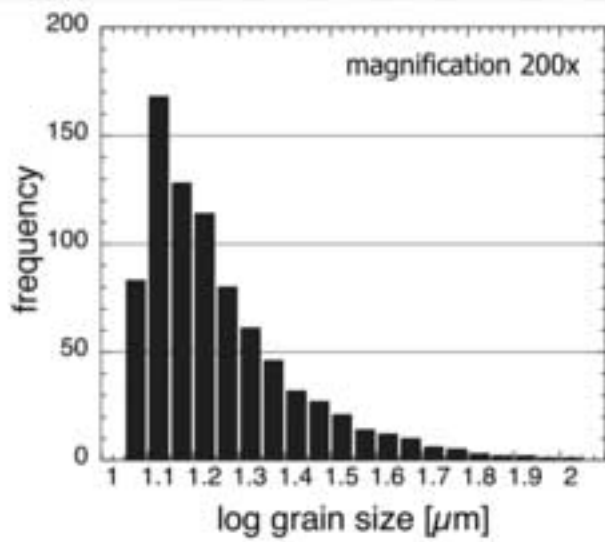
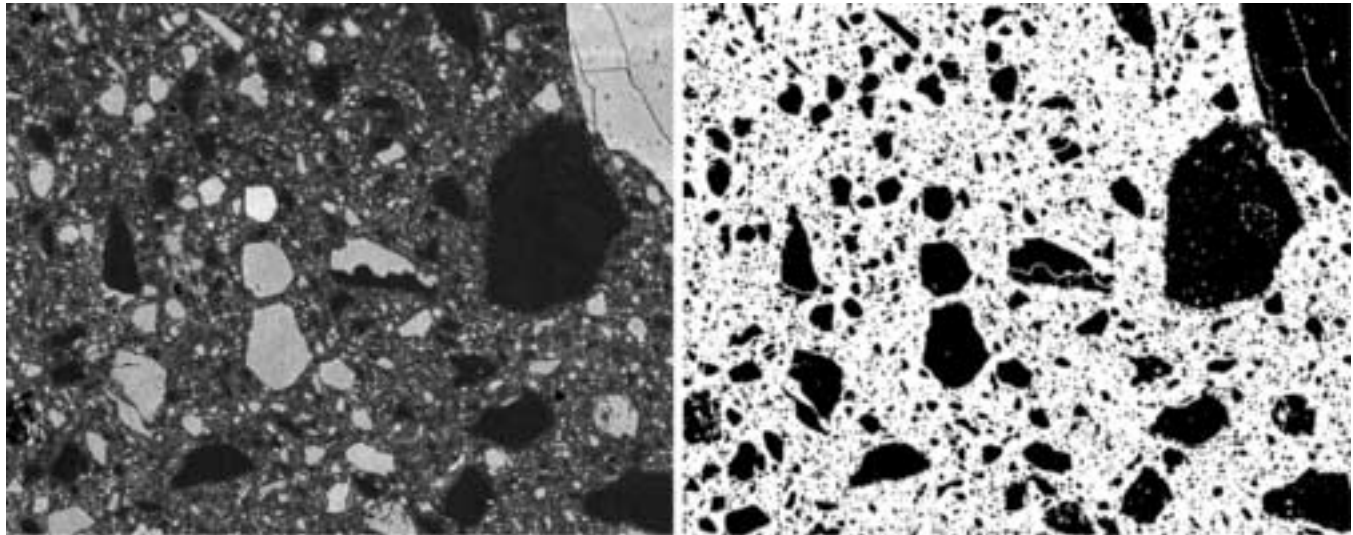
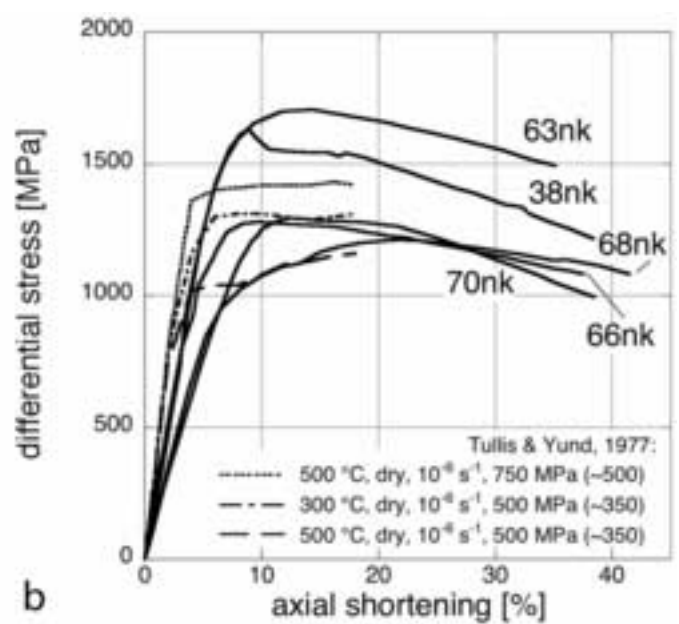
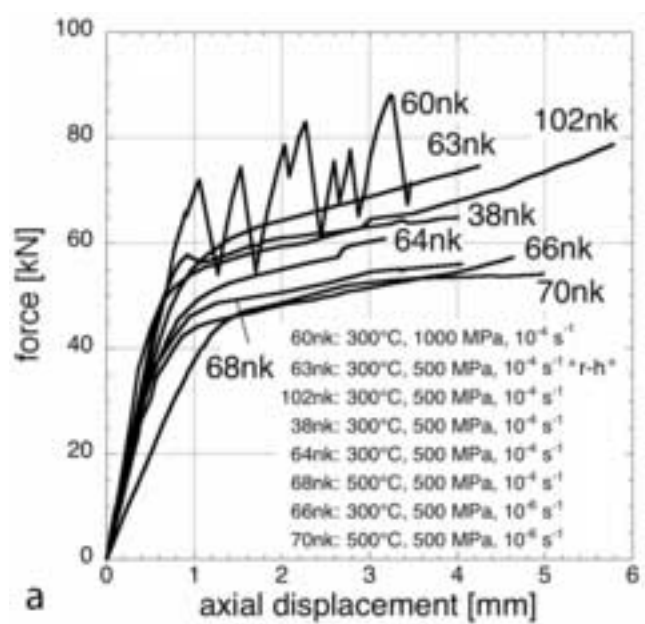
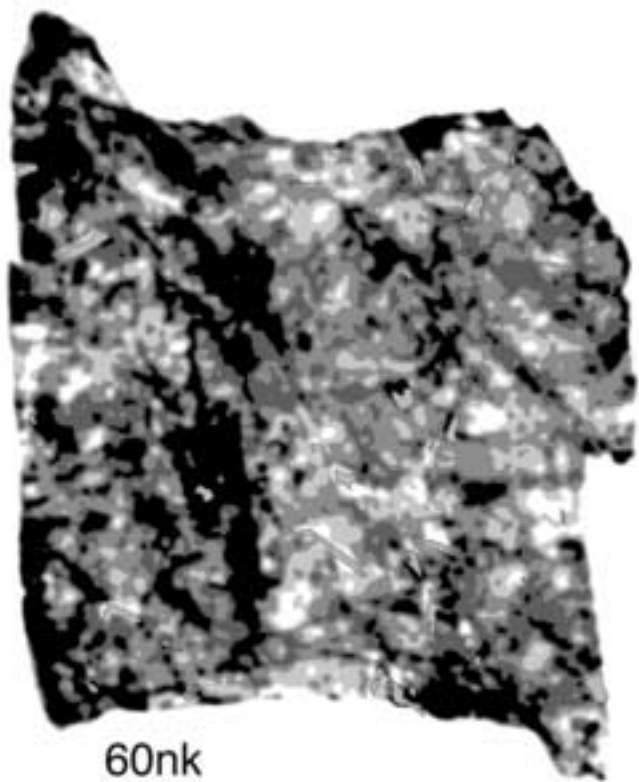
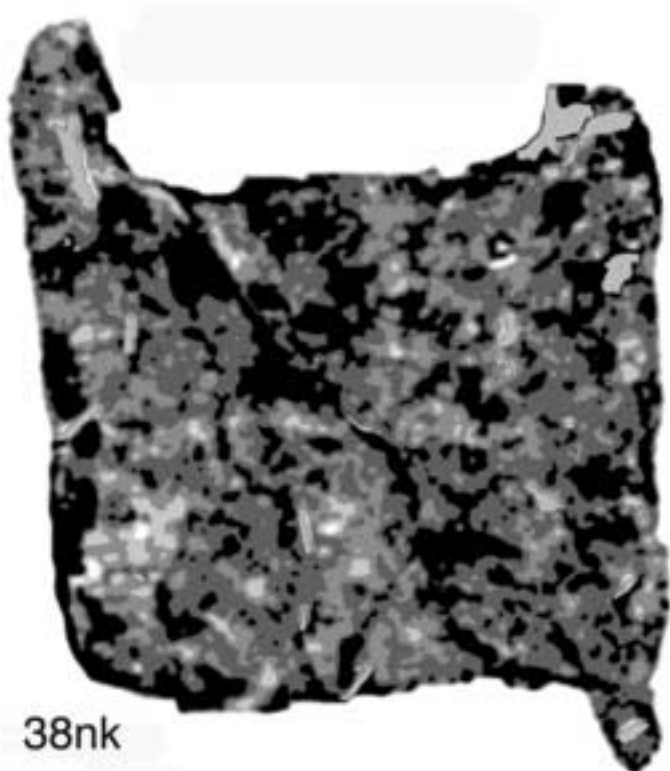


Figure03

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5 mm

D : 1.45 1.55 1.65 1.75



gouge



mask



Figure05

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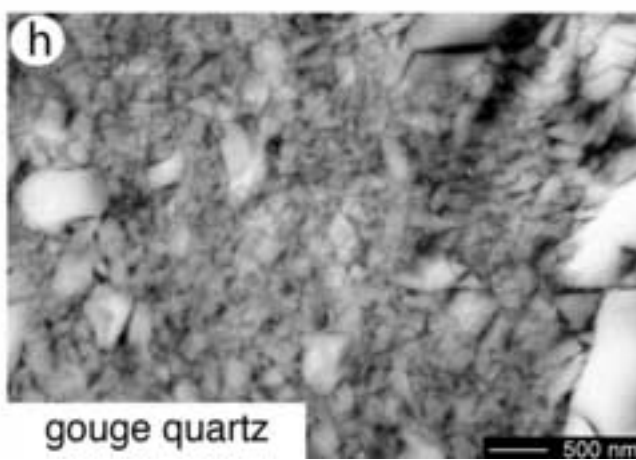
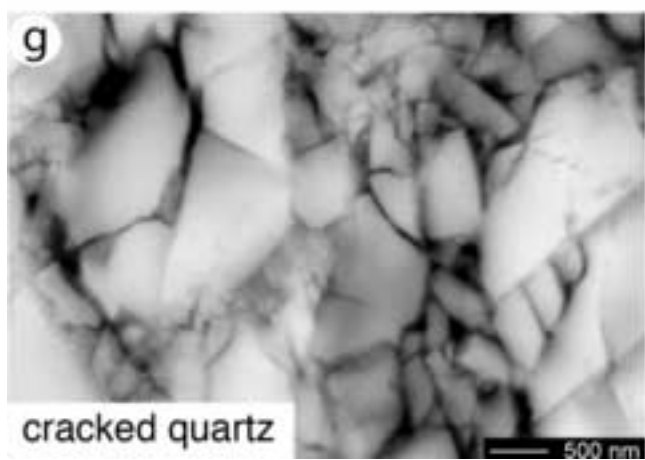
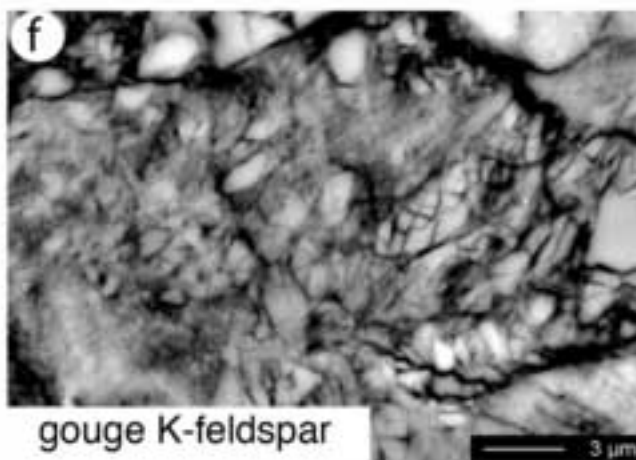
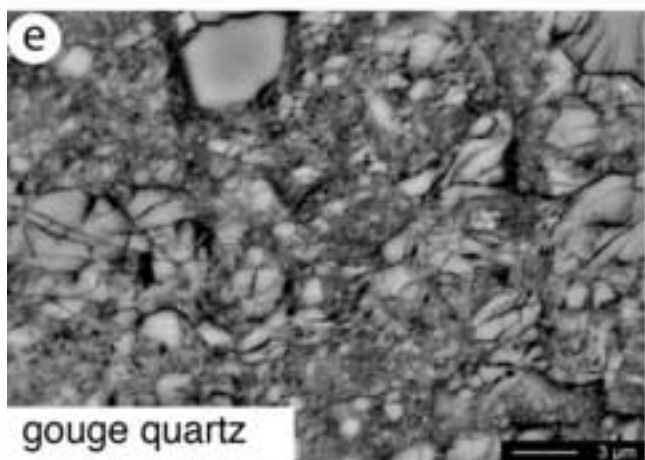
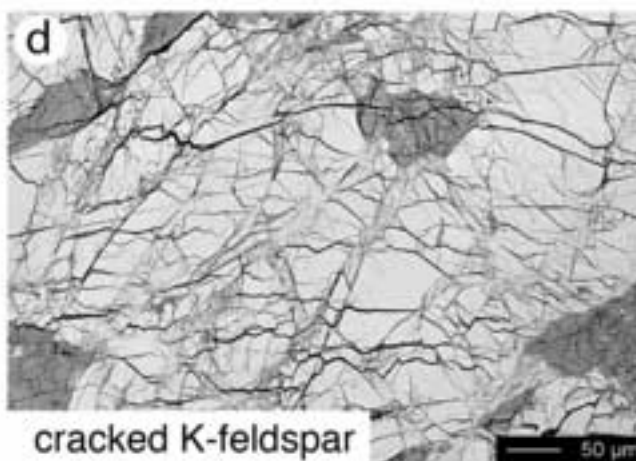
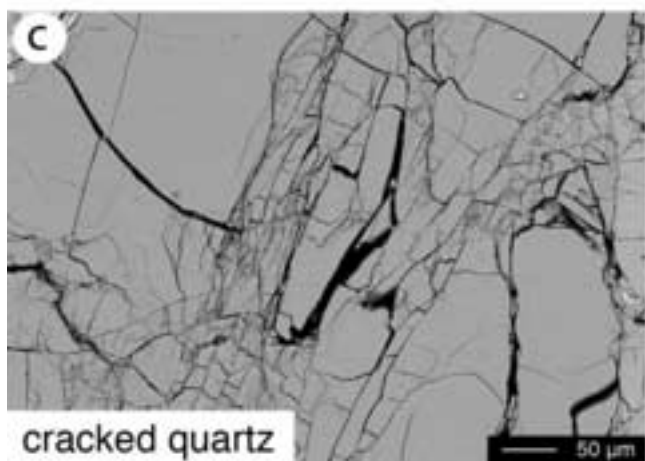
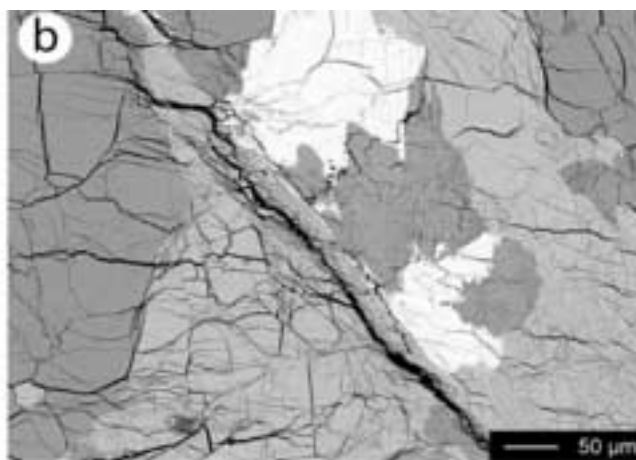
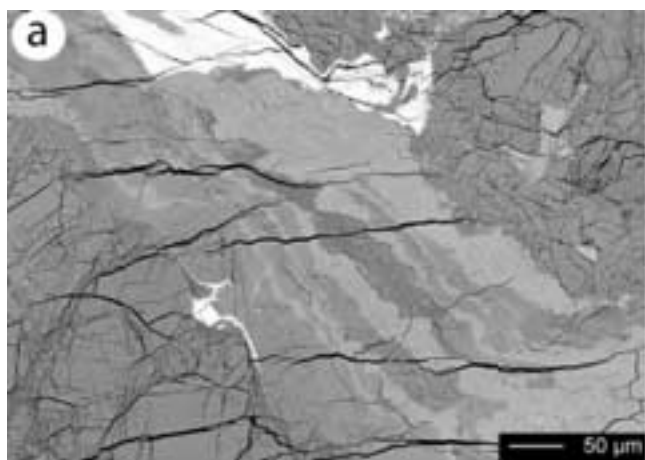


Figure06  
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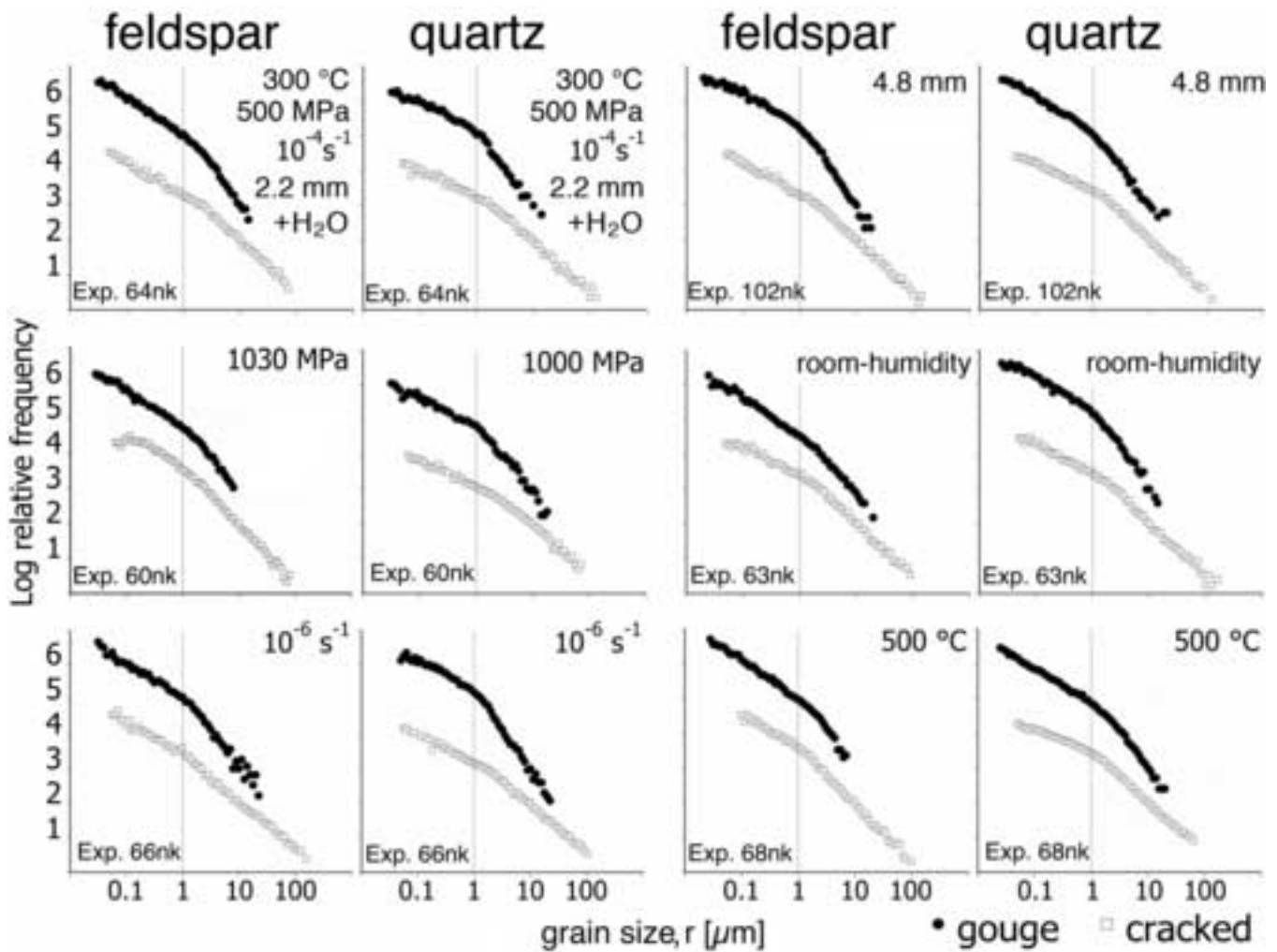




Figure07

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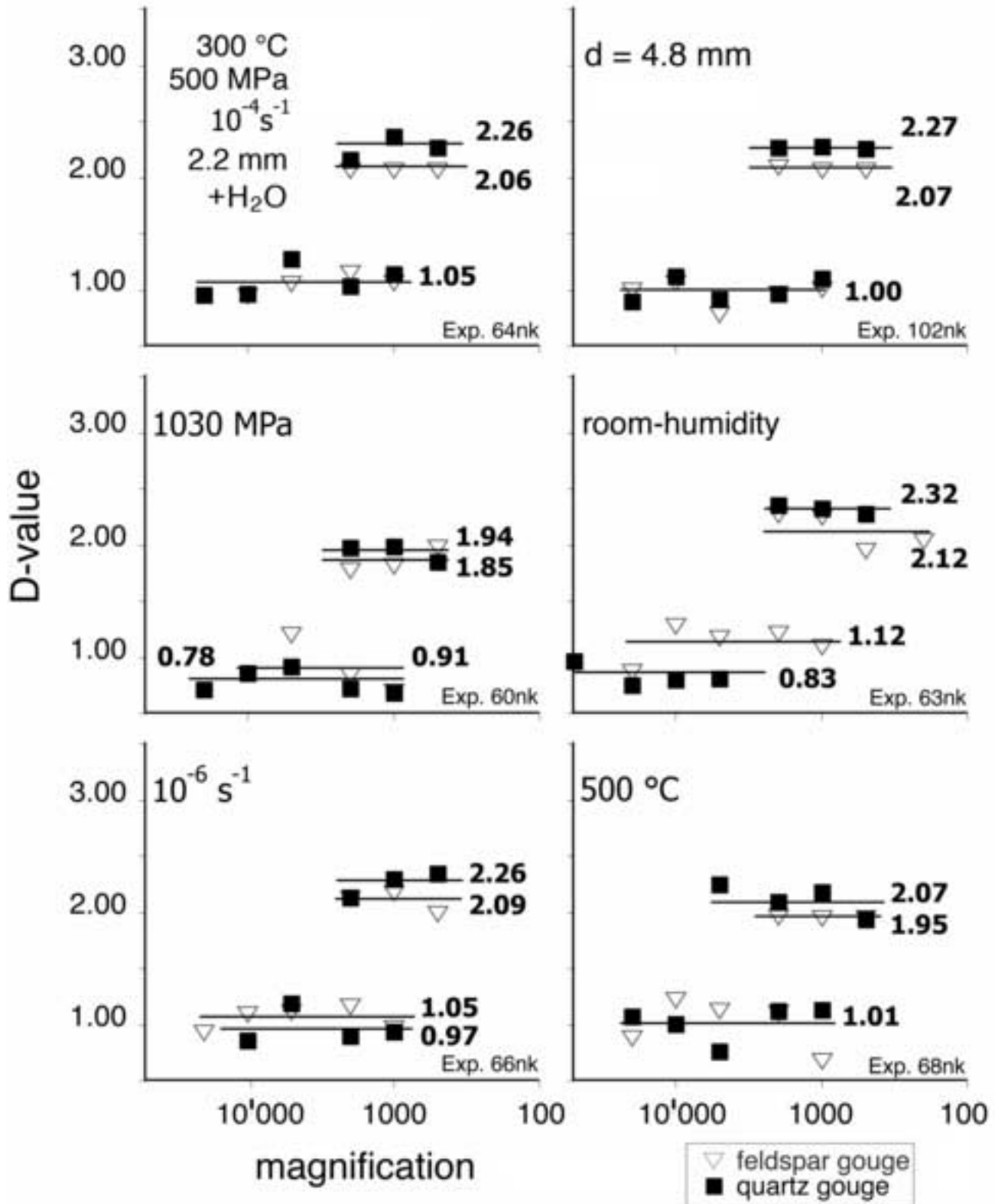


Figure08

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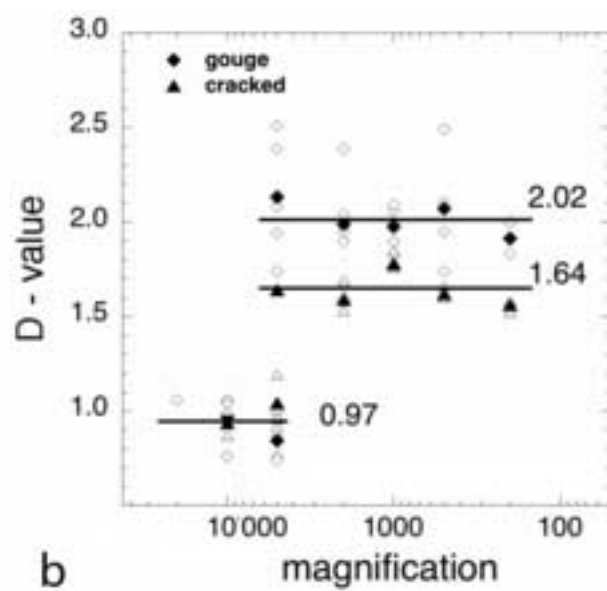
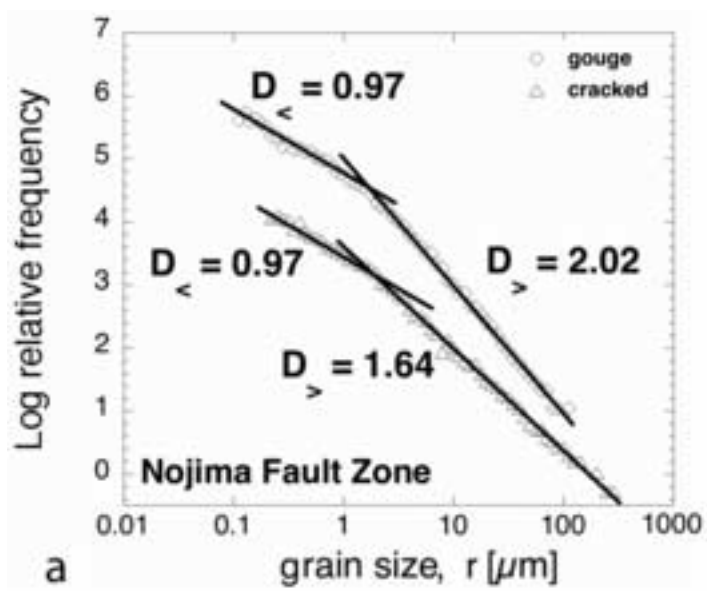


Figure09  
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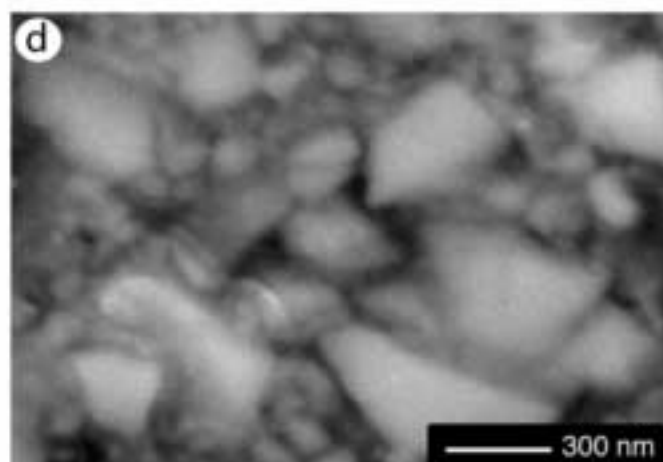
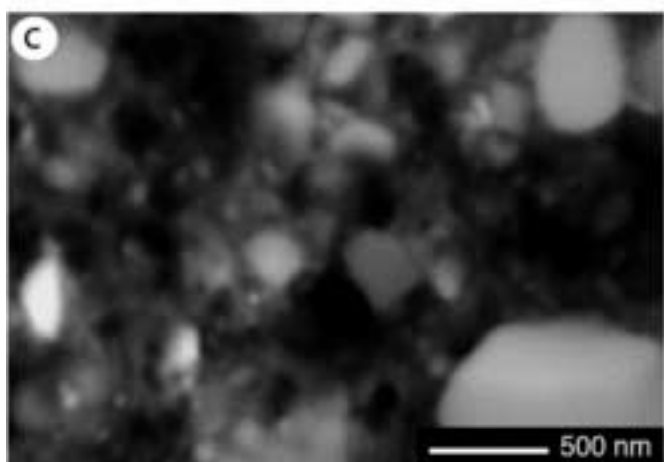
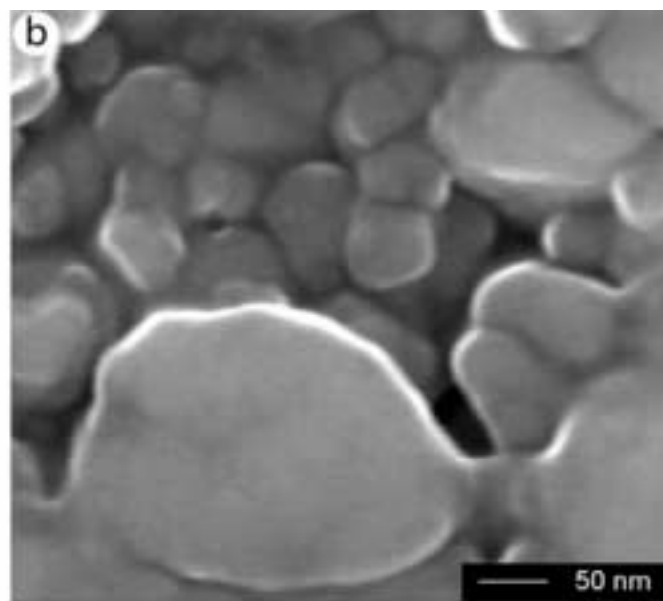
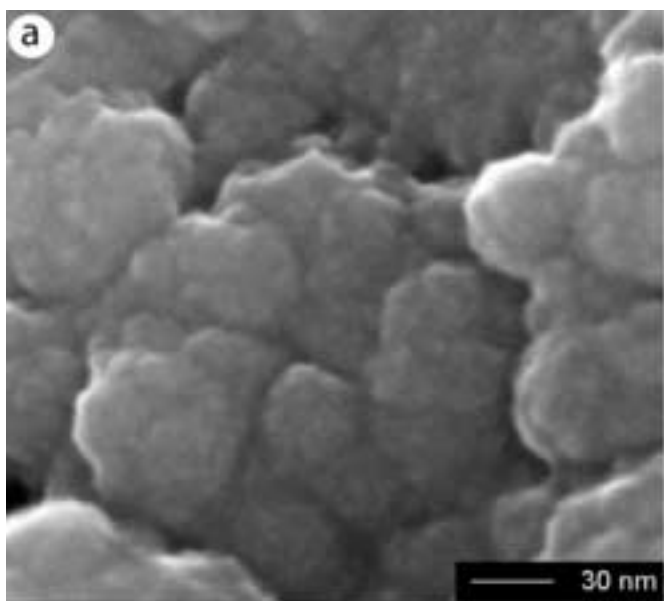
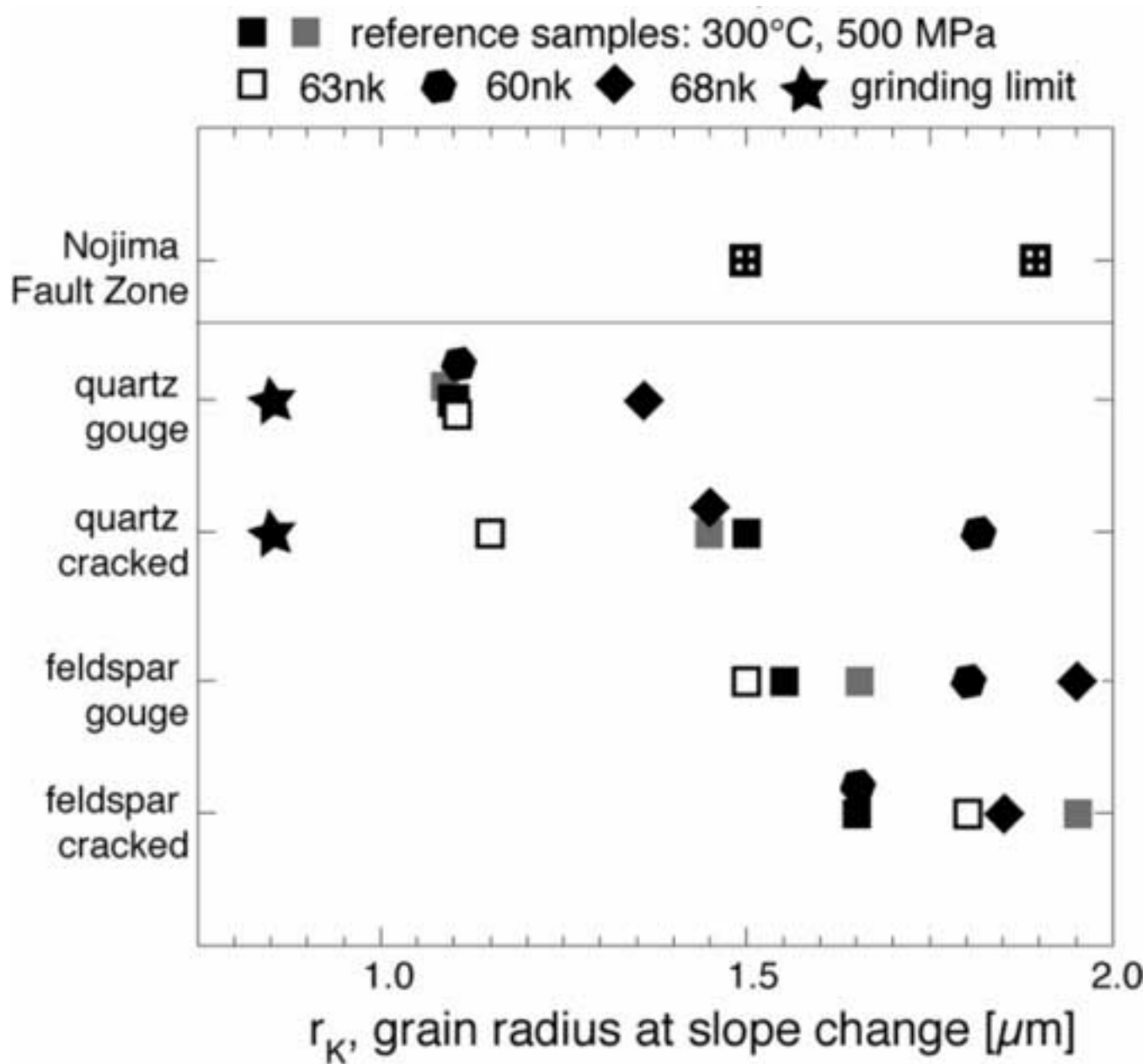


Figure10

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Authors	year	rock type	location	nat/exp	method of analysis	(2D)-fractal dimension	smallest diameter measured, $\mu m$
Sammis et al.	1987	granite	Lopez Canyon	natural	LM + SEM, manual	1.6	2
Sammis & Biegel	1989	granite		natural	manual	1.6	
Marone & Scholz	1989	quartz sand (porous)	Ottawa	experimental <sup>a</sup>	SEM, manual	1.6	1.5
Blenkinsop	1991	granite, gneiss	SAF, Cajon Pass Drillhole	natural <sup>b</sup>	LM, digitizer	0.8 - 2.1	3
An & Sammis	1994	arkose	SAF, Tejon Pass	natural <sup>b</sup>	LM, digitizer	1.6 - 2.0	5
		granite	San Gabriel	natural	sieve + Coult. Count.	1.4 - 2.6	2
		gneiss	Lopez Canyon	natural	sieve + Coult. Count.	1.6 - 1.9	2
Shao & Zou	1996	tonalite	Qinling Mountain	natural	sieve + Coult. Count.	1.5 - 1.9	2
Monzawa & Otsuki	2003	schist and gneiss	Tanakura	natural	LM + SEM, manual	1.6	6
		granite	It.-Shimotsutaki	natural	LM, manual	1.7	10
		granite	Koi	natural	LM, manual	2.1	11
		granite	Nojima	natural	LM, manual	2.1	8.1
Hadzadeh & Johnson	2003	granite	Nojima	natural	LM, manual	2.3	7.7
Wilson et al.;	2005	quartz sandstone	Masillon, Ohio	experimental <sup>c</sup>	LM, digitizer	1.9-2.5	3.5
Reches & Dewers	2005	granite	Saf, Tejon Pass	natural	Coult. Count.		0.04
Chester et al.	2005	granite	Bosman fault, R.S.A.	natural	Coult. Count.		0.04
Heilbronner & Keulen	2006	granite	SAF, Punchbowl	natural	LM + TEM, manual	2.0	0.05
This study		granitoids	Verzasca	experimental <sup>d</sup>	SEM, digital	1.4 - 2.3	0.03
		granitoids	Nojima	natural	SEM, digital	1.6 - 2.4	0.03
		granitoids	Verzasca	experimental <sup>e</sup>	SEM, digital	1.4 - 2.3	0.03

Table 1:

Symbol	Definition
$D_{<}$	slope of power law fit for small grain sizes ( $r < r_K$ )
$D_{>}$	slope of power law fit for large grain sizes ( $r > r_K$ )
$r_K$	radius of grains at slope break of GSD
$L_0$	original length of sample
$\Delta_L$	axial shortening of sample
$r_{min}$	minimum grain size used for calculation of GSD
$r_{max}$	maximum grain size used for calculation of GSD
A	cross sectional area of grain (pixel <sup>2</sup> )
P	perimeter of cross section of grain (pixel)

Table 2:

### Deformation conditions

sample nr.	$T$ °C	$P_c$ MPa	ax. $\dot{\epsilon}$ $\times 10^{-4}s^{-1}$	$L_0$ mm	$d_{chart}$ mm	$d_{(chart-1mm)}$ %wt	$H_2O$
38nk	300	510	1.5	10.44	4.01	3.0	0.2
64nk	300	500	1.3	12.33	3.18	2.2	0.2
102nk	300	510	1.3	10.17	5.78	4.8	0.2
60nk	300	1030	1.5	10.13	3.47	2.5	0.2
68nk	500	520	1.4	10.32	4.28	3.3	0.2
63nk	300	530	1.4	12.11	4.25	3.3	0.0
66nk	300	490	0.015	12.37	4.64	3.6	0.2
70nk	500	500	0.013	12.95	4.97	4.0	0.2

Table 3:

*D*-values for cracked grains and gouge

sample nr.	feldspar				quartz			
	cracked		gouge		cracked		gouge	
	$D_{<}$	$D_{>}$	$D_{<}$	$D_{>}$	$D_{<}$	$D_{>}$	$D_{<}$	$D_{>}$
38nk	1.02	1.68	1.00	2.03	1.09	1.56	1.10	2.26
64nk	0.95	1.52	1.05	2.06	0.74	1.44	1.05	2.26
102nk	0.87	1.52	1.01	2.07	0.92	1.55	1.00	2.27
60nk	0.74	1.37	0.91	1.85	0.96	1.72	0.78	1.94
68nk	0.92	1.67	1.01	1.95	0.92	1.47	1.01	2.07
63nk	0.72	1.62	1.12	2.12	0.78	1.55	0.83	2.32
66nk	0.86	1.37	1.05	2.09	0.96	1.46	0.97	2.26
Nojima Fault Zone								
	cracked		gouge					
	$D_{<}$	$D_{>}$	$D_{<}$	$D_{>}$				
	0.97	1.64	0.97	2.02				

Table 4:



$r_K$ : Grain radius at slope change [ $\mu\text{m}$ ]

sample nr.	feldspar				quartz			
	cracked		gouge		cracked		gouge	
	$r_K$	range	$r_K$	range	$r_K$	range	$r_K$	range
38nk	1.6	1.4 - 1.7	1.5	1.4 - 1.7	1.5	1.3 - 1.6	1.1	0.9 - 1.1
64nk	1.9	1.7 - 2.0	1.6	1.5 - 1.7	1.3	1.2 - 1.4	1.1	1.0 - 1.3
102nk	1.8	1.7 - 2.0	1.5	1.3 - 1.7	1.4	1.4 - 1.5	1.1	0.9 - 1.1
60nk	1.7	1.6 - 1.7	1.8	1.6 - 1.9	1.8	1.8 - 2.0	1.1	1.0 - 1.2
68nk	1.8	1.6 - 1.9	1.9	1.9 - 2.0	1.5	1.4 - 1.6	1.3	1.3 - 1.5
63nk	1.8	1.7 - 1.9	1.5	1.5 - 1.6	1.1	1.1 - 1.3	1.1	1.0 - 1.3
66nk	1.6	1.3 - 1.6	1.3	1.3 - 1.6	1.8	1.6 - 1.9	1.4	1.4 - 1.6
Nojima Fault Zone								
	cracked		gouge					
	$r_K$	range	$r_K$	range				
	1.9	1.6 - 2.0	1.5	1.2 - 1.6				

Table 5: