Deformation at the frictional-viscous transition: Evidence for cycles of fluid-assisted embrittlement and ductile deformation in the granitoid crust

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Abstract

Mid-crustal deformation is classically characterized by the transition from ductile to brittle deformation 4 5 defining the frictional-to-viscous transition (FVT). Here we investigate an exhumed continental mid-6 crustal basement section in order to envisage the relationship between ductile and brittle deformation at 7 the FVT. Our detailed study from km- to micro-scale shows that, under greenschist metamorphic 8 conditions, deformation is accommodated by a dense network of highly-localized ductile shear zones. In 9 the investigated case it is not quartz which defines the overall ductile deformation behavior but the viscous 10 granular deformation in shear zones with an ultrafine-grained polymineralic matrix consisting of quartz, 11 feldspar, sheet silicates and epidote. During viscous granular flow mass transfer processes under the 12 presence of fluids promote a chemo-mechanical mixing, resulting in grain size reduction and reaction softening. Coeval with this ductile deformation, fluid-assisted embrittlement occurs, as indicated by 13 14 biotite-coated fractures, cataclasites and injection of non-cohesive polymineralic gouge material into secondary fractures inside the host rock. The embrittlement during predominant ductile deformation 15 occurs in cycles, i.e. prolonged periods of slow viscous granular flow are interrupted by rapid brittle 16 17 deformation. We interpret this fluid-assisted cyclic embrittlement evidenced by injection of the fluidized 18 material into off-fault fractures as an alternative equivalent to pseudotachylites and as a microstructural 19 indicator for paleo-seismic activity. With exhumation and associated cooling, localized deformation 20 persists in the ultrafine-grained polymineralic shear zones but progressively transitions to cataclastic flow 21 and finally to pressure-dependent frictional flow; always showing cycles of slow interseismic flow and 22 fast seismic injection events. Overall, in the granitic crust of the Aar-massif, brittle and ductile deformation coexist up to deformation temperatures of minimum 450°C, indicating that the FVT has to be 23 placed in a rather wide range from 8 km up to > 18 - 20 km in the granitoid crust. 24

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Keywords: frictional-viscous transition, cyclic embrittlement, viscous granular flow, high temperature
 friction, fluids

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30 1 Introduction

In a simplified view, the deformation behavior can be subdivided into an upper brittle part and a ductile lower part (Fig. 1), where the frictional-viscous transition (FVT) separates the two end-member deformation behaviors (Fig. 1b). Although any deformation mechanism has to be micro-physically based, its expression in form of flow laws and their extrapolation to crustal scale resulted in a rather broad use of the term FVT in literature. Three general applications of the term FVT can be discriminated: (1) a crustalscale depth interval (2) a bulk rheological deformation behavior and (3) a microstructural deformation style.

- 38 (1)The crustal-scale depth interval expresses the transition between the pressure-dependent Byerlee's 39 frictional behavior of the upper crust and the temperature-dependent viscous flow of the middle to 40 lower crust. The resulting depth range depends on the extrapolation of the two end-member flow 41 laws of the major rock forming minerals (e.g. calcite, quartz, feldspar and olivine) deforming either 42 in the brittle (e.g., Hirth and Beeler 2015 and references therein) or ductile field (e.g. Burgmann and 43 Dresen 2008 and references therein). Depending on the rock-forming mineral, FVTs occur at 44 different depths resulting in classical Christmas tree type strength profiles (Fig. 1b). Such simplified 45 crustal rheologies generally provide the base for crustal scale geodynamic numerical modeling (e.g., Ranalli, 1995; Gerya, 2010; Duretz et al., 2015). 46
- Instead of a sharp rheological transition between frictional and viscous deformation, a gradual
 rheological change is observed in experiments promoting the generation of transitional physical
 flow laws for the FVT (e.g., Shimamoto and Noda, 2014). Extrapolations of such laboratoryderived flow laws (e.g., Kawamoto and Shimamoto 1998) allow for modifications of Christmas tree
 type strength profile (Fig. 1c).
- 52 (3) On the microscale, the contemporaneous occurrence of frictional features (microcracks, fractures, 53 cataclasites) with evidence for viscous deformation also indicates deformation at a FVT. While this 54 type of transition is known as semi-brittle behavior in the case of monomineralic aggregates (e.g. 55 Hirth et al. 1992), it is of particular importance in the case of polymineralic rocks, where often 56 frictional deformation occurs in rigid phases embedded in a mechanically weak viscous matrix (e.g., 57 Regenauer-Lieb and Yuen, 2003, 2004). Here the resulting rock strength of the composite aggregate 58 strongly depends on the amount and spatial distribution of the different mechanical phases (e.g., 59 Jordan 1987, Handy 1990, Bloomfield and Covey-Crump 1993, Kawamoto and Shimamoto 1998, 60 Handy et al. 1999, Barnhoorn et al. 2005, Margues et al. 2010, Ji and Zhao 1993, Ji et al. 2003, 61 2004).
- Moreover, by activating frictional deformation, the mechanically strong phases can undergo severe grain size reduction ending up in micron-sized polymineralic aggregates. The dramatic increase in grain surface areas enhances the efficiency of dissolution and precipitation processes (Rutter 1983),

65 allows for grain rotations and the sliding of grains along each other (Fitzgerald and Stünitz 1993, 66 Paterson 1995) and promotes mineral reactions (e.g., Menegon et al. 2013, Stünitz and Tullis 2001, 67 Stünitz 1993). The combination of all these deformation processes in ultrafine-grained 68 polymineralic aggregates is referred to as granular flow (Paterson, 1995, 2013). It depends on temperature, fluid content, grain size and the mineralogy (e.g., Wintsch and Yeh, 2013). At low and 69 enhanced temperatures, respectively, different grain sizes and strain rates will result either in 70 71 frictional granular flow (polymineralic cataclasites) or viscous granular flow (polymineralic 72 mylonites; Fig. 1).

73 The application of (1-3) in nature depends on the spatial scale of interest and on the integrated time of 74 deformation. Both, frictional and viscous deformations are based on discontinuous processes and can 75 therefore be of temporary character. For example, frictional deformation can be activated at low to 76 moderate strains in order to generate microfabrics suitable for high strain viscous deformation (Segall and 77 Simpson, 1986; Christiansen and Pollard, 1997; Guermani and Pennacchioni, 1998; Herwegh and Kunze, 78 2002; Pennacchioni et al. 2006, Mancktelow and Pennacchioni, 2005; Pennacchioni, 2005; Pennacchioni 79 and Mancktelow, 2007). In addition, building up of pore fluid pressures or strain rate weakening processes 80 will induce switches from slow viscous to fast frictional processes manifesting seismic cycles (Fig. 1c, 81 e.g., Handy et al. 2007, Fusseis and Handy 2008, Poulet et al., 2014). Hence the term FVT has a rather 82 broad application and it has to be defined specifically, which time- and space-scales are addressed when 83 dealing with it.

84 In this study we follow the arguments of Handy et al. (2007) and apply the term FVT to a depth range in 85 the granitoid crust, in which the occurrence of both brittle and ductile processes occur. We investigate the deformation in granite, which we consider as important representatives for the middle continental crust 86 (gneisses, granites). The study benefits from the relatively large volume of granitoid rocks exposed in the 87 88 Aar-massif (Central Alps, Switzerland) and considers the fault rocks as natural laboratories for several 89 reasons: (1) these rocks show simple compositions; (2) strain initially localized along steep shear zones 90 within relatively isotropic host rocks at depths of $\sim 18 - 20$ km; and (3) during exhumation of the Aar-91 massif, strain further localized in these steep shear zones and progressively narrowed in width (Wehrens et 92 al., in review). The latter allows a telescoped view of deformation behavior at different depths. We link 93 field investigations and microstructural observations from these different depths and search for evidence 94 for simultaneous activity of both frictional and viscous deformation processes as well as the depth 95 limitations of the FVT. Special emphasis will be paid to the effect of fluids, given their importance for 96 hydrofracturing, thermal-pressurization as well as mass transfer processes; all processes playing a major 97 role in the FVT (e.g., Wintsch and Yeh, 2013). In contrast to the general assumption of quartz as the 98 dominant rheology-defining mineral in the middle crust, we will demonstrate that the formation of 99 ultrafine-grained polymineralic fault rocks (ultramylonites and fault gouges) is much more significant, 100 given their high abundance and ability to reduce fault strengths from the surface down to depths of 20 km.

101 **2** Geological setting

The Aar massif is located in central Switzerland and represents one of the external massifs of the Alps. It consists of polymetamorphic (pre-)Variscan basement rocks, which have been intruded by post-Variscan granitoid rocks around 298 Ma (Abrecht, 1994; Schaltegger, 1994). The present study focuses on a transect through the Southern Central Aar massif (Haslital transect), where post-Variscan plutons are bounded by an aplitic boundary facies at the contact to the Paleozoic gneisses of the Grimsel Zone (Fig. 2; Stalder, 1964; Schaltegger, 1990).

108 Although deformation in the Paleozoic basement is long lasting (Proterozoic, Ordovician to Variscan and 109 Alpine; Stalder 1964; Steck 1966, 1968, 1984; Schaltegger 1993; Schaltegger et al. 2003), the Post-110 Variscan plutonic bodies only underwent Alpine deformation. Temperatures at the Northern rim of the 111 Aar massif never exceeded ~250°C (e.g., Bambauer et al., 2009), whereas at the Southern boundary 112 (Grimsel Pass), peak Alpine conditions were in the range of 450°C and 600 MPa (Challandes et al., 2008; 113 Goncalves et al., 2012). This suggests that the Southern rim of the Aar massif, i.e. our study area, was at a 114 depth of about 18-20 km at its maximum. A multiphase Alpine deformation history, consisting of a phase 115 of compressional deformation under peak metamorphic conditions, followed by transpressional 116 deformation associated with retrogression and finally exhumation, affected the study area (Steck, 1968; 117 Challandes et al., 2008; Rolland et al., 2009; Wehrens, 2015). Strain is distributed in the Haslital along 118 many major shear zones (up to tens of meters wide and several km long) and a multitude of small discrete 119 shear zones (<10 cm thickness).

120 **3 Methods**

In the studied area, zones of localized deformation can be subdivided into brittle fault zones (fractured rock, cataclasite, breccia, fault gouge) and ductile shear zones. Given their both ductile and brittle nature, we refer to them in general as fault zones following the nomenclature of Schmid and Handy (1991).

124 A field-based study was conducted in order to define the km-scale strain distribution of both brittle and 125 ductile deformation zones. Detailed mapping of individual shear zones along and across their strain gradients provided information on their evolution and strain localization history. We focus on granitoid 126 127 rocks, since they exclusively underwent Alpine deformation and their petrology is well known (Keusen et 128 al., 1989; Schaltegger, 1989). Samples were taken across transects of ductile shear zones and brittle faults. 129 In addition, drill core samples from fault gouges in the NAGRA Grimsel Test Site (GTS, Blechschmidt 130 and Vomvoris, 2010) were analyzed. The structures in these samples were impregnated with a fluorescein 131 doped resin prior to drilling in order to provide cohesion during the drilling process: preserving the in-situ 132 microstructure and porosity of the fault gouges (Tanaka et al., 2014).

Microstructural analyses were carried out using optical light and scanning electron microscopy. The dimensions and heterogeneity of the structures were identified on the micro-scale, which provided the basis for further microstructural and geochemical analyses. Fault gouge samples from the GTS, stabilized by the aforementioned approach, have been analyzed in under UV-microscope highlighting thefluorescence of the infiltrated resin.

- Phase and element distribution along brittle and ductile structures were investigated at the thin section scale. For this purpose, electron backscatter imaging and X-ray mapping with energy dispersive
- 140 spectrometry (EDS) was carried out using a ZEISS EVO 50 scanning electron microscope with an
- 141 acceleration voltage of 20 kV and a beam current of 3-6 nA. Element maps were generated by EDAX
- 142 TEAMTM software (Nylese and Anderhalt, 2014), and element quantification in oxide wt.% was carried
- 143 out. The created element maps and backscatter images were combined to determine the phases and were
- 144 analyzed with ImageJ (http://imagej.nih.gov/ij/) to calculate their phase area percentages. Multiple areas 145 were analyzed along each structure to control the quality of phase- and element quantification. Due to the
- 146 heterogeneous nature of the structures, great care was taken in the choice of the location and size of the
- 147 area, in order to represent the structures best. Adding the results for all areas along one individual structure
- 148 defines a representative average. The total variation of the element wt.% results is indicated with bars in
- 149 the graphs. The total analyzed area for each structure ranged from 1.5 to 22 mm^2 . Each total analyzed area
- 150 consists of multiple individual areas with sizes of 0.5-4.5 mm². Each area contained in the order of 200-
- 151 1000 grains. For each hydrous phase, the water content was calculated on the basis of the obtained phase
- 152 volume percentage and mineral stoichiometry (Deer et al., 1992).

Electron backscatter diffraction (EBSD) under low-vacuum conditions (10-15 Pa) was carried out on a whole suite of oriented and polished thin sections. Here, additional polishing with colloidal silica was applied in order to reduce damage of the sample surface. Representative crystal orientation maps and corresponding pole figure diagrams were obtained from Euler angle measurements, using the TSL/Ametek OIM version 7.0 software package.

158 4 Results

159 4.1 Spatial and temporal relationship of the investigated structures

160 The units described above are cut by numerous shear zones, which can be well grouped from overprinting 161 relationships and orientations (Wehrens, 2015, see section 2). The maximum formation depth of the 162 exhumed shear zones is around 18-20 km and estimated maximum metamorphic temperatures are around 163 450°C (e.g., Goncalves et al., 2012). These shear zones are steeply inclined and show shear zone 164 narrowing during the exhumation (retrograde) path. Over the past years, we have generated a 165 comprehensive inventory of fault zones and associated fault rocks in the study area (Baumberger, 2015; 166 Wehrens, 2015; Belgrano et al., 2016). Based on (1) large-scale shear zones and their overprinting 167 relationships as well as (2) microstructures of frictional and viscous deformation, we will demonstrate that 168 brittle deformation is not restricted to retrograde "young" deformation in shallow crustal levels (Fig. 1), 169 but it also occurred at maximum metamorphic conditions. In order to envisage the occurrence of brittle deformation under peak metamorphic conditions, i.e. at high temperature (HT), we follow the following strategy. Firstly the host rock as starting material of deformation will be introduced. We then present a characterization of the aforementioned low temperature (LT) brittle structures. Subsequently, ductile shear zones (formed under HT) will be shown before addressing the criteria and characteristics of high strain-

- 174 rate brittle faulting under HT peak Alpine metamorphic conditions.
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176 **4.1.1** Host rocks

The host rocks are ranging from leucocratic granite (Central Aar granite) to granodiorite (Grimsel 177 178 granodiorite) and polymetamorphic granitoids (Grimsel zone; Stalder, 1964; Niggli, 1965; Keusen et al., 179 1989; Schaltegger, 1989). The granitoids show a primary compositional increase in biotite from 5 to 180 12 vol%, which is a key for variation in the mechanical behavior (Wehrens et al., in review). The weakly 181 deformed granitoid rocks still show magmatic feldspars with their primary sizes (Table 1). Interstitial 182 quartz and biotite nests exist between the large magmatic feldspar grains. The quartz is mostly 183 dynamically recrystallized by subgrain rotation recrystallization and few remaining hosts grains show 184 undulose extinction. The average recrystallized grain size of the subgrains is around 180 µm (Table 1). A 185 weak foliation may be present where biotite is aligned and connected between magmatic feldspar grains. 186 Epidote, most notably allanite, occurs as single crystals with a relative large grain size (Table 1).

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188 **4.1.2** Low temperature brittle structures

189 The low temperature cataclastic zones are clearly younger than the ductile shear zones (see below) as they 190 overprint their ductile precursors. Given the common reactivation of the ductile shear planes, they must 191 have formed under retrograde conditions. In the following, they are named low temperature cataclasites 192 (LT-cataclasite, Table 2). The non-cohesive gouge material and cataclasites is often washed out at the 193 surface by recent erosion (Fig. 3c). To enable the study of deformation microstructures and associated 194 deformation processes, these structures were stabilized in a complex approach in the underground GTS 195 laboratory (Grimsel Test Site) by drilling, resin injection and over-coring allowing the in-situ sampling. 196 The host granitoid shows magmatic feldspars being successively overprinted by ductile deformation as 197 manifest by grain size reduction of feldspars as well as elongation and dynamic recrystallization of quartz 198 aggregates (Fig. 4e). The magmatic and mylonitic microstructures are dissected by individual fractures as 199 well as a polymineralic cataclasite (Fig. 4a-c). Clasts within the 1-2 mm wide cataclasites are angular and 200 consist of feldspar, quartz, mica and some epidote. The grain size in the cataclasite is much smaller than 201 that of the host rock (Table 1, 2). This polymineralic cataclasite occurs even within fractures cutting 202 through monomineralic coarse-grained parts of the host rock (Fig. 4c). Furthermore, no overgrowth of 203 mineral phases in the cataclasite is observed. UV-fluorescence light microcopy shows the resin-accessible 204 porosity within the cataclastic matrix (Fig. 4b). Resin can also be found along the straight to slightly 205 curved grain boundaries of the dynamically recrystallized quartz grains (Fig. 4d). Note that in both cases

206 the occurrence of the fluorescent resin indicates open pore space is present in the nowadays subsurface 207 non-cohesive fault gouges within brittle fault zones.

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209 4.1.3 Localized ductile deformation

210 Ductile shear zones range from mm to several meters width. Laterally, they can extend over several 211 kilometers. Inside the mylonites, feldspar clasts are reduced in size by brittle processes (Fig. 5a, b; Table 212 1, 2), and interstitial quartz domains become elongated by crystal plastic deformation (Fig. 5b), while 213 biotite is aligned parallel to the shear plane (Fig. 5a, b). With increasing strain, the size of feldspar grains 214 is further reduced. Quartz domains are more elongate and a polymineralic matrix of very fine-grained 215 recrystallized quartz, biotite, white mica and minor epidote develops (Fig. 5a-c). The vast majority of 216 ductile shear zones consist of ultramylonites with polymineralic composition, while monomineralic quartz 217 aggregates occur sparsely in the form of lenses or bands. The grain sizes within the polymineralic matrix 218 of such mylonites are order of magnitudes smaller as those of the surrounding gneisses (Table 1). Within 219 monomineralic quartz domains, few remnant (primary) quartz grains reveal undulose extinction patterns 220 and the presence of subgrains. Progressive subgrain formation leads to newly recrystallized quartz grains 221 (subgrain rotation recrystallization; Guillope and Poirier, 1979; Drury and Urai, 1990). The grain size in 222 these aggregates is around 100 µm (area-weighted mean grain size; Figs. 5b, 6a). In most samples, 223 dynamic recrystallization of monomineralic quartz aggregates is nearly complete throughout the fabric 224 (Fig. 6a). All these observations infer subgrain rotation recrystallization as the dominant dynamic 225 recrystallization process with a minor component of grain boundary migration being present as well. In 226 order to define the deformation mechanisms in quartz, a characteristic, monomineralic quartz aggregate 227 and a polymineralic layer were quantitatively analyzed (EBSD; Fig. 6a-c). The data are presented in a 228 grain orientation map (Fig. 6a), which was calculated from Euler angles obtained from the EBSD 229 measurement, and a pole figure (lower hemisphere, equal-area plot; linear intensity contouring, Fig. 6b, c). 230 These monomineralic quartz aggregates typically reveal a moderate crystallographic preferred orientation 231 (CPO; Fig. 6c). The pole figure of the [c]-axis (0001) indicates a strong peripheral maximum (cluster), 232 rotated around 60° with respect to the foliation (f) following the sense of shear. In this aggregate, a [c]-233 axis single girdle is developed. The $\langle a \rangle$ axes (11-20) consistently align subparallel to the stretching 234 lineation (L). Rhomb $\langle a \rangle$ glide can also be identified, as some of the [c]-axes is concentrated between the 235 Y- and Z-axes (Fig. 6c). The fine-grained polymineralic layers are intercalated with biotite bands (Figs. 236 5d, 6d). In the case of the polymineralic layers, grain sizes of quartz vary between 10-80 µm depending on 237 the degree of pinning by the other minerals (Fig. 6e, f). Both the quartz [c]-axes and $\langle a \rangle$ axes show 238 random distributions (Fig. 6g, h), clearly contrasting the CPOs of the monomineralic layers.

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Cyclic embrittlement

240 4.1.4 High temperature brittle structures

As will be demonstrated in the following, brittle structures exist at high temperatures, i.e. at conditions similar to those described for the aforementioned ductile deformation. The relative ages are inferred from field crosscutting relations (Fig. 7). This section presents the microstructural characteristics of these hightemperature (HT) brittle structures.

The host rock fabric consists of magmatic feldspar and dynamically recrystallized quartz grains (see section 4.1.1). Both of these features are cut by micro-fractures (Fig. 7b). They also cut through aggregates and individual grains and are filled with a polymineralic assemblage consisting of biotite, albite, white mica and epidote (Fig. 7b, c). Transgranular and intergranular fractures can be observed within the dynamically recrystallized quartz aggregates as indicated by biotite precipitates (Fig. 7c, d). Hence, they are younger than dynamic recrystallization of quartz in these microstructural domains, but the conditions of this brittle deformation were still within the mineral stability field of biotite.

252 Additionally, epidote-filled veins are observed parallel to the shear plane cutting through both the 253 magmatic feldspar clasts and recrystallized quartz aggregates. Ductile micro-shear zones developed along 254 epidote-filled veins, which are defined by quartz and biotite recrystallization. In addition, cataclasis is 255 observed at the margin of a decameter-scale shear zone (Fig. 8a). A cataclasite (HT-cataclasite) occurs 256 within the least ductilely deformed part of the decameter-sized strain gradient of a shear zone (Fig. 8a). At 257 the weakly deformed margin, the host rock is fractured in a similar fashion as seen in the LT-cataclasites. 258 The fractures in the host are composed of clasts surrounded by a fine-grained polymineralic matrix (Table 259 1). At several locations, smaller secondary fractures, which branch from the main cataclasite, cut into the 260 weakly deformed host rock (Fig. 7a, b). The fine-grained polymineralic matrix is also observed in fractures passing through monomineralic host rock domains (Fig. 7b). The polymineralic matrix shows 261 262 very small grain sizes (Table 1; Fig. 8c, f). It is important to note, in contrast to the LT cataclasite 263 example, that biotite (50-150µm) is overgrowing the polymineralic matrix. This observation places the 264 conditions of brittle cataclastic deformation into the mineral stability field of biotite (Fig. 8c).

Occasionally, micro-epidote veins cut this ultrafine polymineralic matrix (most obvious as white lines in Fig 7c), again recording brittle deformation. At some localities within the ultrafine polymineralic matrix these veins are even folded (Fig. 7d, f). This folding illustrates that viscous processes affected the finegrained matrix and veins, which both have a brittle origin.

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4.2 Chemical and mineralogical compositions of different tectonites

We investigate LT-cataclasites, ductile shear zones and HT-cataclasites in detail. The selected microstructural domains reflect a representative selection which is based on the inspection of a large series of thin sections (Table 2). The detailed microstructural analyses revealed differences between the described structures and their mineralogical compositions. A LT-cataclasite and a ductile mylonite have been studied within the Central Aar granite. The LT-cataclasite is compositionally similar to the granitic

- host rock (Figs. 9, 10, 11). The K_2O to SiO₂ ratio is higher for the cataclasite (Fig. 11). The water content within the LT-cataclasite (0.4 wt%) is similar to that of the undeformed granitoid, which ignores potential
- 278 water content within the pore space (Fig. 10b).
- 279 In contrast to the LT-cataclasite, the mylonite shows significant variations compared to the host granitic
- 280 composition (Fig. 10a). An increase in mica within the mylonite relative to the weakly deformed Central
- Aar granite has been measured. Furthermore, the albite content decreased within the mylonite compared to
- the host. The mylonite shows a loss in Na₂O and a gain in K₂O, MgO and to a lesser degree Fe_2O_3 compared to the host (Fig. 11). The calculated water content for the mylonites is ~1 wt% and is therefore higher than in the undeformed Central Aar granite (Fig. 10b). Similar observations have been published
- for shear zones within the Grimsel granodiorite (Marquer et al., 1985; Goncalves et al., 2012).
- Four samples were analyzed along a strain gradient inside the Grimsel Zone (Fig. 9). The gradient shows increasing strain over a distance of 40 meters from weakly deformed granitoid to an ultramylonite (Fig. 9a). The structures along the gradient consist of (i) an epidote-rich HT-cataclasite from the weakly deformed granitoid part. (ii) weakly sheared epidote white mica-rich bands and (iii) schistose granitoid band from the moderately deformed schistose granitoid with highly sheared epidote-albite-mica rich layers, and finally (iv) an ultramylonitic core (Fig. 9).
- 292 The HT-cataclasite clearly differs from the Grimsel Zone granitoid composition and is dominated by (i) 293 albite and epidote (Figs. 10, 11). The HT-cataclasite consists of almost 50 vol% epidote compared to 294 a few vol% in the host rock. Furthermore, the HT-cataclasite sample has a low quartz content 295 (6 vol%) and almost no K-feldspar. Also the element wt.% ratios are clearly different from the granitoid composition (Fig. 11). The HT-cataclasite shows higher values for Na₂O₃ Al₂O₃ and most 296 297 clearly for CaO. The K₂O to SiO₂ ratio is lower than for the granitoid composition. Furthermore, the sample is poor in SiO_2 (Fig. 9c). The water content within the HT-cataclasite (1.2 wt%) is higher than 298 299 in the undeformed granitoid and mylonites (Fig. 10b).
- (ii) The schistose granitoid band contains less mica and slightly more epidote although is generally
 similar to the granitoid host (Fig. 10). The element wt.% shows a strong similarity with the Grimsel
 Zone host rock and no significant changes in elemental abundance can be noted (Fig. 11). The water
 content of 0.26 wt.% is lower than in the host and comparable to that of the Central Aar granite (Fig.
 10).
- (iii) The epidote white mica-rich band shows a markedly different composition to the Grimsel Zone granitoid (Figs. 9, 10). This band contains a large amount of epidote (23 vol%) and white mica (17 vol%). Furthermore, the quartz content is extremely low (4 wt.%). The element wt.% of Na₂O, Al₂O₃, K₂O, CaO for this area plots in between the host rock and the HT-cataclasite (Fig. 11). The water content of this second schistose area (1.27 wt.%) is comparable to that of the HT-cataclasite (Fig. 11).
- 310 10b).
- (iv) At the most strongly deformed part within the strain gradient, the ultramylonite displays an increase
 in white mica compared to the host rock and a marked decrease in K-feldspar (Figs. 9, 10). The

element wt.% of the various elements indicates no large differences between the ultramylonite and the host rock, although a slight loss of K_2O can be noted (Fig. 11). The water content within the ultramylonite is around 1 wt.%, therefore higher than in the undeformed granitoid rocks and comparable to the water content of the mylonite (Fig. 10b).

317 Overall, higher water content for deformation structures compared to the undeformed rock is noted (Fig.

318 10b). The ductile mylonite and ultramylonite show an increase in mica content compared to the host rocks

and a decrease in feldspar (Fig. 10a). The LT-cataclasite shows a similar composition to the host and no

320 major change in major element chemistry. The HT-cataclasite is not comparable to the composition of the

- 321 surrounding Grimsel Zone granitoid.
- 322

323 **5 Discussion**

This section discusses the interplay of grain size reduction, viscous and frictional deformation mechanisms and cyclic fast-brittle and slow-ductile deformation. We suggest that the presence of fluids particularly contributes to the associated mechanical and chemical deformation behavior. In the following discussion it will be crucial to discriminate between deformation occurring during the main Alpine HT metamorphic conditions and subsequent deformational overprint during retrograde cooling.

329

330 5.1 Deformation mechanisms

331 The main characteristics of the investigated fault rocks under both brittle and ductile deformation are 332 attributed to a grain size reduction and a mechano-chemical phase mixing (e.g., Kruse and Stünitz, 1999; 333 Linckens et al., 2015; Figs. 4, 5a-c, 8). During deformation, the grain-size of originally coarse grained (0.2 334 -2 mm) magmatic mineral phases like feldspar and quartz is dramatically reduced by up to three orders of magnitude as observed in all described tectonites (mylonite, ultramylonite, HT-cataclasite and LT-335 336 cataclasite; Table 1). The associated progressive evolution into compositional layers in the mylonites/ultramylonites implies an important role of chemical processes during these deformation 337 338 processes. In the following, we discuss the brittle and ductile deformation mechanisms and afterwards 339 their spatial and temporal interplay.

340 **5.1.1** Low temperature cataclasites

Within the LT-cataclasites, the original grain size is reduced to angular clasts of various sizes (Table 1). Although being partly affected by an earlier grain size reduction during preceding ductile deformation (e.g. subgrain rotation of dynamically recrystallized quartz, see below), deformation in these tectonites is controlled by purely brittle grain refinement and frictional flow as already described for many cataclasites (e.g., Rutter, 1986, Marone, 1998; Snoke et al., 1998; Keulen et al. 2007). Angular clasts (μ m – tens of µm) are surrounded by an ultrafine-grained polymineralic matrix, which accommodated the cataclastic 347 flow. In addition to this probably aseismic cataclastic flow, the gouge also experienced stages of fluidized 348 granular flow as made evident by the observed (i) granular texture, (ii) sharp contacts between gouge and wall rock, (iii) sorting of grain size classes of clasts in fault-plane-parallel bands and (iv) the injection of 349 350 non-cohesive gouge into off-fault cracks in the host rock (see also Monzawa and Otsuki, 2003; Boullier et 351 al., 2009; Fondriest et al., 2013; Rowe et al. 2012; Rowe and Griffith, 2015). Particularly the injection 352 structures, the non-cohesive character of the gouge as well as the potentially laminar fluidized granular 353 flow point to fast processes under the presence of water and elevated pore fluid pressure at seismic rates at 354 shallow crustal levels. Note that the observed LT-microstructures and inferred deformation processes will 355 provide an important base for the interpretation of the HT frictional microstructures (see below).

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357 5.1.2 Mylonites and ultramylonites

358 Initial fracturing or the concentration of stress along pre-existing mechanical anisotropies forced strain to 359 localize in the ductile field, as already proposed in previous studies (Fig. 7; Mancktelow and Penniachioni 360 2005; Mancktelow and Penniachioni 2013, Wehrens et al., in review). Besides these strain nucleation 361 features, in the samples studied, additional strain-softening processes were activated. Quartz deforms in a 362 ductile manner by means of dislocation creep. The dominance of this deformation mechanism within 363 monomineralic quartz layers is identified based on the active slip systems combined with a 364 crystallographic preferred orientation (CPO) as well as typical subgrain rotation recrystallization 365 microstructures (Fig. 6a-c; e.g., Hirth and Tullis, 1992; Herwegh and Handy, 1996; Kilian et al., 2011). 366 The fabric is dominantly characterized by active basal and rhomb glide along the $\langle a \rangle$ axis (Schmid and 367 Casey, 1986), which is an indicator for deformation of guartz under greenschist facies conditions (e.g., 368 Schmid and Casey, 1986; Stipp et al., 2002).

369 Within the polymineralic domains of the ultramylonites, the weak CPO of quartz, the equiaxial small grain 370 sizes and the homogeneous mixing of the different phases within mylonitic layers suggest deformation by 371 viscous granular flow (diffusion creep) involving grain boundary sliding, dissolution and precipitation and diffusive mass transfer processes (Fig. 6d-h, see also Fliervoet et al., 1997; Paterson, 1995). In our 372 samples, neo/re-crystallization of biotite, white mica, quartz and feldspar takes place. A volumetric 373 374 increase in the mica content as well as viscous granular flow in combination with mineral reactions and 375 mass transfer processes (e.g., Stünitz and Fitzgerald, 1993; Paterson, 1995; Menegon et al., 2013) allowed 376 for an enrichment of such mechanically weak phases in these high strain zones. A combination of grain 377 size softening and reaction softening therefore controls strain localization in these polymineralic tectonites 378 (Stünitz, 1998; Goncalves et al., 2012; Marsh et al., 2009).

379

380 **5.1.3** High temperature cataclasites

381 Similar to the LT-cataclasites, HT-cataclasites show reduced grain size and angular clasts of various sizes, 382 which are embedded in an extremely fine-grained matrix (down to a few µm; Table 1; Figs. 4, 8). Sharp 383 contacts to wall rocks and grain size sorting in bands parallel to the fault plane are also present (Fig. 8a). 384 Deformation is therefore controlled by brittle grain refinement and frictional granular flow (cataclastic 385 flow). As already mentioned for the polymineralic ultramylonites, dilation in combination with 386 dissolution-precipitation and mass transfer processes represents important strain accommodation processes 387 during granular flow in these ultrafine-grained deformation products. Fluid-assisted injection of low 388 viscosity gouge material into secondary fractures implies a complete embrittlement of the polymineralic 389 aggregates and a loss in cohesion between grain contacts. In contrast to the non-cohesive LT-cataclasites, 390 however, the HT-cataclasites were initially cohesive, as made evident by the intergrowth microstructures. 391 Probably the presence of fluids in combination with earthquake shaking allowed the aggregates to 392 fragment along the grain boundaries, promoting the injection of the at-that-time non-cohesive 393 polymineralic gouge into the off-fault fractures (van der Elst et al., 2012; Rowe and Griffith, 2015). This 394 brittle deformation must have occurred under HT-conditions, as indicated by the overgrowth of ultrafine-395 grained cataclastic matrix by biotite and the ductile folding of veinlets (Fig. 8). Given these high 396 temperatures, non-cohesive grain boundaries have a low survival potential, requiring both fast 397 embrittlement and injection processes.

398

5.2 The role of fluids during brittle and ductile deformation

400 Fluids play a crucial role in the development of the brittle and ductile fault rocks observed in this study 401 (e.g., Etheridge et al., 1984; Marquer et al., 1985). On the one hand, fluids serve as transport media for 402 mass transfer processes, on the other hand, the building up of pore fluid pressure increases the potential 403 for hydrofracturing (e.g., Sibson, 1989; Cox, 2010) or ductile shear fracturing (Weinberg and Regenauer-404 Lieb, 2009). Principally three important questions need to be discussed with respect to our natural 405 examples: (1) the potential presence or absence and amount of fluid at different stages of deformation, (2) 406 the role of fluids for deformation mechanisms in the frictional and viscous field and (3) their implications 407 for the resulting rheology. Despite numerous studies on the effect of dissolution-precipitation processes in 408 carbonates (Rutter 1986), gypsum (de Meer et al. 1997) and phyllosilicates (Bos and Spiers 2001), to our 409 knowledge yet no experimental data for granitic systems in general and for polymineralic ultramylonites 410 in particular exist. Such knowledge would be crucial for absolute estimates on parameters in any flow law. 411 In the following, we therefore have to restrict on a discussion of (1) and (2).

412 A variety of evidence exists for the presence of fluids during the entire evolution of the tectonites in the 413 Aar massif (e.g., Gonclaves et al. 2012). In the following we treat the effects of fluids as a function of 414 increasing depth. Chemical analyses have shown a similar mineralogical composition of both the LT-

415 cataclasite and their host rock, which implies a purely mechanical grain size reduction and phase mixing

416 without the necessity of chemical reactions or mass transfer via fluids (Figs. 4a-c, 9c). Contrastingly, the 417 fine-grained polymineralic matrix within fractures in monomineralic host rock domains, points towards a 418 fluid-assisted injection of non-cohesive polymineralic material into the opening fractures (see Fig. 4c and 419 above). Both the injection structures of fault gouge material and the existence of the open fractures

420 illustrate the presence of fluids under conditions of LT- deformation.

421 Fluids as transport media play a peculiar role during mineral precipitation and mineral growth at elevated 422 deformation temperatures. For example, the quartz or epidote-albite veins, as well as the inter- and 423 transgranular fractures in the ductilely recrystallized quartz aggregates, filled with fine-grained biotite or 424 polymineralic precipitates, indicate fluid-assisted fracturing and mass transfer at HT-conditions (Fig. 7b-425 d). Also the significant chemical difference between HT-cataclasite and the granitoid host rock requires 426 fluids as mass exchange media (Figs. 10, 11). During stages of low to intermediate pore fluid pressures, 427 i.e. during stages of slow aseismic creep, these fluids promote the build up of cohesion in the 428 polymineralic fine-grained matrices of the HT-cataclasites. Owing to their high surface areas, surface-429 controlled dissolution and precipitation processes become efficient under elevated temperatures. These 430 chemico-physical conditions allow an enhanced coupled grain coarsening in the polymineralic domains 431 (Herwegh et al. 2011 and references therein), leading to the build-up of renewed cohesion after each 432 brittle HT-event. In this way, the mechanical character of the polymineralic matrices alternatingly changes 433 between non-cohesive and cohesive. Last but not least, the injection structures of fine-grained 434 polymineralic matrix into off-fault fractures, combined with the overgrowth of biotite, again prove the 435 presence of fluid during cataclastic deformation under upper greenschist facies conditions (Fig. 8).

436 Besides the cataclasites, also the mylonites show evidence for deformation-assisted fluid flow and element 437 transport. Element transport is illustrated by the higher K_2O to SiO₂ ratio of the mylonite compared to the host (Fig. 11; see also Marquer et al., 1985; Goncalves et al., 2012). During deformation by viscous 438 439 granular flow (i.e. precipitation and dissolution, grain boundary sliding, creep cavitation) dynamic 440 porosity is created at the micro-scale promoting fluid transport along grain boundary voids and 441 intergranular cavities (Etheridge et al., 1984; Ree, 1991; Herwegh and Jenni, 2001; Fusseis et al. 2009; 442 Menegon et al. 2015). Such advective fluid flow during viscous granular allows element transport and 443 growth of new phases (Herwegh and Jenni, 2001; Goncalves et al., 2012; Menegon et al., 2015).

Altogether, these observations clearly indicate fluid activity plays a major role ranging from shallow and low to mid-crustal high temperature deformation. In the latter case, mass transfer processes become increasingly important with rising temperature. For fluid flux in the studied fault system two end-members have to be discriminated: (i) Stages of continuous and (i) discontinuous fluid in- and exfiltration. (i) The enrichment of K₂O, SiO₂ and hydrous phases indicates an open system with infiltration of external fluids. Such infiltration may have occurred during continuous grain-scale processes (e.g. granular fluid pump of

such minimution may have occurred during continuous grain scale processes (e.g. grandal nard pump of

450 Fusseis et al. 2009). Fluid sources probably were dehydration reactions at lower crustal levels. (ii)

451 Contrastingly rapid and discontinuous fluid flux and exfiltration is indicated by the presence of veins (e.g.

452 quartz, epidote) and more importantly by the fluidized injections of ultrafine-grained gouge material

453 eventually being initiated at sites of locally enhanced pore fluid pressures. Such fast processes are an 454 efficient way to transport significant amounts of fluids through the highly localized fault zone channel.

455

456 **5.3** Frictional and viscous deformation and its cyclic behavior

457 **5.3.1** Incipient brittle deformation as nucleus for strain localization

458 Incipient brittle fracturing provoking subsequent dynamic recrystallization has been described in previous 459 publications (e.g., van Daalen et al., 1999; Mancktelow, 2006; Vernooij et al., 2006; Fusseis and Handy, 460 2008; Diamond and Tarantola, 2015). In nature, recognition of these incipient precursor brittle structures 461 is often difficult, since they usually vanish due to overprinting during ductile shear zone widening (Fig. 462 12). Low strain domains with limited ductile overprint (Fig. 12b see 3-6) therefore represent the most 463 promising sites for the detection of brittle precursors on both the micro-scale (van Daalen et al., 1999; 464 Vernooij et al., 2006; Diamond and Tarantola, 2015) as well as the m- to km-scale (e.g., Mancktelow and 465 Pennacchioni, 2005; Fusseis et al., 2006). In the case of the shear zones of this study, several pieces of 466 evidence reveal a brittle origin for many of the km-scale shear zones in the Aar massif: (i) straight and 467 planar lateral appearance over kilometers, (ii) occurrence of biotite-coated fracture patterns on the 468 millimeter to decameter scale as well as overprinting of brittle horse-tail structures by ductile shear zones 469 at shear zone tips on the decameter scale (see also Wehrens et al., in review; their figure 5a-c). Besides the 470 structural crosscutting/overprinting relation between incipient brittle and succeeding ductile structures, 471 incipient fracturing is of great importance from a mechanical point of view, since a significant stress drop 472 is expected between the stage of incipient non-localized to post-fracture deformation of the host rock (Fig. 473 13). Such early rate- and pressure-dependent fracturing therefore provides the base for subsequent ductile-474 brittle cycles acting with smaller stress variations (see below). A major new aspect of this study is the 475 occurrence of the non-cohesive HT-cataclasites. In contrast to aforementioned HT-fractures with their 476 biotite-coating, severe displacements in combination with fast energy releases are necessary to supply a 477 considerable amount of non-cohesive polymineralic material as fluid-supported lubricant during strain-478 rate weakening and the injection of the fluidized material into the off-fault fractures. Such events may 479 have already occurred during the initialization stage but must be more prominent during succeeding stages 480 of brittle-ductile cycles.

481 **5.3.2 Cyclical deformation behavior**

The change between frictional and viscous deformation at the FVT has often been discussed in earthquake geology (see introduction and Figs. 1, 12, 13 and 14). The role of rate dependency on deformation mechanisms and/or switches between different deformation mechanisms is incorporated in established FVT-models. Depending on the time-scale of interest, effective viscosity can be either expressed by a time-integrated approach (e.g., Noda and Shimamoto 2010; Shimamoto and Noda 2014 and references therein) or by the temporal resolution of the activity of different microstructural processes (e.g., Handy et

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488 al., 2007; Poulet et al., 2014). It is the latter time-scale, which actually links to switches in deformation 489 processes resulting in changes from aseismic to seismic creep. Aforementioned structures and related deformation mechanisms demonstrate the alternating occurrences of mylonites dominated by viscous-490 491 granular-flow/dislocation creep and HT-fracturing. This can described as switching between frictional and 492 viscous deformation (Figs. 13a). Transferred to crustal rheology such switches require modifications from 493 the classical Christmas tree strength profiles (Fig. 1c, 13). In addition brittle and ductile end-member flow 494 laws, cyclic behavior of brittle and ductile deformation as well as dominant deformation in ultrafine-495 grained polymineralic mylonites/cataclasites have to be integrated. The cyclic behavior is evident from 496 two types of observations: (1) ductile overprinting relationships inside the HT-cataclasites and, vice versa, 497 brittle overprinting in ultramylonites; as well as (2) the coexistence of repeated ductile and brittle 498 deformation within the same pressure/temperature interval.

- 499 (1) The formation of microscopic epidote veins within the HT cataclasite matrix and their subsequent 500 folding suggests that brittle and ductile deformation occur cyclically (Figs. 7c-d, 12 and 13). 501 Prolonged periods of relatively slow viscous granular deformation are interrupted by events of rapid 502 embrittlement. Folding of micro-scale epidote veins within the HT-cataclasite indicates the renewed 503 onset of viscous granular flow after such fast brittle events (Fig. 8c, d). Also, the HT-ultramylonites 504 contain former epidote veins, but additionally show highly elongated and deformed epidote-rich 505 lenses indicating first an event of brittle vein formation followed by boudinage during subsequent 506 viscous granular flow of the polymineralic ultramylonite (Figs. 9a, 12 and 14).
- 507 (2) Biotite-filled micro-fractures cutting through dynamically recrystallized quartz grains are one of the
 508 brittle structures overprinting ductile deformation (Fig. 7b-d). This requires primary viscous flow in
 509 quartz (mainly dislocation creep), which is overprinted by frictional processes at elevated biotite 510 stable temperatures.
- 511 In summary, prolonged periods of slow viscous granular processes are interrupted by rapid brittle 512 deformation. The positive feedback between localization and fluid-assisted chemical mass transfer 513 (granular fluid pump) results in major strain softening (Figs. 1 and 13). Porosity destruction by mineral 514 crystallization and fracture sealing, as seen for example by biotite overgrowths of the coupled grain 515 coarsening of the matrix of the HT-cataclasite, induce a reduction in permeability. This reduction in permeability enables a recovery of the pore fluid pressure during stages of viscous flow, which results 516 517 again in fracturing once the yield strength is reached. The subsequent post-fracturing stress- and pore fluid 518 pressure drop causes again a switch, back into the viscous granular flow regime (Figs. 1, 12, 13 and 14).
- 519 Under physical conditions of homogeneous flow (constant strain rate, constant mineral dissolution and 520 precipitation, constant cavity formation) theoretically a steady state granular flow fabric may evolve. As 521 shown in our deformation structures, however, microfabrics and deformation style are rather 522 heterogeneous (see above). The simultaneous occurrence of synkinematic quartz-veins within 523 ultramylonites suggests an interplay between the granular fluid pump (Etheridge et al. 1984, Fusseis et al., 524 2009) and a fault valve behavior in the sense of Sibson (1992). Combining the two models requires stages

of slow but pervasive fluid flux in the fault zones during granular flow but also stages where locally pore fluid pressures enhance to induce hydrofracturing. The question by which processes and at which time scales the building up of elevated pore pressures occur in the granitoid crust (Figs. 13 and 14) and when the system starts to fail catastrophically in form of earthquake activity remains open and requires further research.

530

531 **5.4 Seismic versus aseismic behavior**

532 From surveillance of recent seismic activity in N-Switzerland it is known that earthquakes occur down to 533 crustal depths of up to 20 km (e.g., Deichmann et al., 2000, 2011; Diehl et al. 2015), i.e. in a depth range 534 structurally documented by the exhumed granitoid fault rocks of our study. In many studies, 535 pseudotachylites are classically seen in such rocks as indicators for paleo-seismicity (Cowan, 1999; 536 summary in Rowe & Griffith 2015). Commonly, pseudotachylite formation is interpreted to occur under dry deformation conditions (Pennacchioni and Cesare, 1997; Yardley and Valley, 1997; Mancktelow and 537 538 Pennacchioni, 2004) but evidence for pseudotachylite generation under hydrated conditions also exists 539 (Magloughlin, 1992; Price et al. 2012). In the case of the hundreds of investigated shear zones of the Aar 540 massif, we did not encounter any pseudotachylites. One may relate this paucity to the low preservation 541 potential of these delicate deformation features (Sibson and Toy, 2006; Kirkpatrick et al., 2009; 542 Kirkpatrick and Rowe, 2013). The lack of pseudotachylites in the Aar granitoids may either mean that (i) 543 seismic strain rates were too slow to induce friction-induced melting, (ii) the friction between grains was 544 not high enough to allow for significant temperature rises at the aggregate's scale, (iii) friction-induced 545 heat was dissipated fast enough via advective fluid flow. The aforementioned recent seismic activity 546 demonstrates that the studied granitoid rocks should have experienced episodes of fast seismic deformation during the exhumation history of the Aar massif and we therefore exclude (i). The presence 547 548 of fluids strongly affects (ii) and (iii). In particular, the lubrication of grain boundaries by fluids, the 549 fragmentation of rocks by seismic shaking at elevated pore fluid pressures and the advective heat transport 550 during injection of the non-cohesive gouge into the host rock all counteract significant shearing induced-551 temperature rises, which would be required for frictional melting. If true, a potential temperature rise by 552 frictional heating during seismic faulting must therefore have been lower than the solvus of the wet 553 granitoid crust. For these reasons, we speculate that the presence of fluids prevents pseudotachylite 554 formation in the case of the granitoid rocks of the study area providing insights into alternative processes 555 for energy dissipation during seismic events.

Particularly fluidization during embrittlement has been associated with seismicity in previous studies (Smith et al., 2008; Brodsky et al., 2009; Bjørnerud, 2010, Rowe & Griffith 2015), where prolonged periods of slow viscous granular flow are interpreted to be interrupted by rapid seismic embrittlement. The observation of cyclic occurrences of frictional events, as shown in our study area, provides additional evidence for seismic activity (Scholz, 1998; Gratier et al., 2002; Handy and Brun 2004, Matysiak and 561 Trepmann, 2012; Wintsch and Yeh, 2013). In this sense we conclude that the aforementioned HT-562 cataclasite injection structures are indicators of paleo-seismic activity. Whereas the presence of substantial amounts of fluids prevents the formation of pseudotachylites due to strain weakening and cooling effects, 563 564 representing an alternative explanation to shear melting as a focal mechanism of mid-crustal earthquakes. 565 In light of the preservation potential of such injection structures, best changes to find such remnants in 566 high strain zones are the low strain rims (left parts of Fig. 12) or domains within pressure shadows. In the 567 ductile high strain domains, however, ductile overprint of the short-term embrittlement structures will be 568 severely overprinted if not even obliterated. This is particularly problematic in the case of isochemical compositions of host granite, injection structures and final mylonite. As demonstrated above, the injection 569 570 structures often contain substantial amounts of epidote. We therefore suggest that the epidote-rich bands in 571 Figures 9a and 12 represent older injections cycles, which experienced stages of high strain ductile 572 overprint. Hence, the occurrence (distribution, spacing) of such epidote-rich bands within ultramylonites 573 might represent indicators of paleoseismic activity.

574

575 **5.5** Linking the rock record to crustal-scale rheology profiles

576 In this section, we link the observed microstructures and associated deformation mechanisms with 577 potential crustal-scale rheology profile(s) incorporating their evolution in space and time. First, the simpler and well-explored monomineralic quartz system (e.g., Hirth et al. 2001, Hirth and Beeler 2015 578 579 and references therein) is treated before dealing with the more complex, but volumetrically far more 580 dominant polymineralic granitoid system. We follow the considerations of Simamoto and Noda (2014) 581 and link them with our field observations. Although few flow laws for viscous granular flow have been 582 suggested (e.g. Paterson 1995, 2013, Platt 2015) their degree of uncertainty is still too high to satisfactorily describe the rheology of granitoid ultramylonites deformed in the range of 300 – 450°C. We 583 584 are therefore restricted at this stage to the conceptual level.

585 At elevated temperature and enhanced pore fluid pressure, i.e. at conditions with reduced effective 586 pressure, both coarse-grained vein quartz and dynamically recrystallized quartz aggregates can 587 hydrofracture at seismic rates resulting in a drop in shear stress (steps 1 & 2 in Fig. 13a). Subsequent grain 588 growth and dynamic recrystallization during the post-seismic deformation induces an adaptation of the 589 now mylonitic microstructures to the slow interseismic deformation rates, first by strain hardening 590 followed by steady-state or at least near steady-sate creep (step 3 in Fig. 13a). Simultaneously, the pore 591 fluid pressure builds up again, before kicking off a renewed seismic event by ductile shear fracturing, 592 where steps 2 and 3 repeat (Fig. 13a). Dilation accompanying the fracturing generates pore space enabling 593 the precipitation of the biotite or the fine-grained polymineralic trails observed in Figure 7. Particularly 594 the isometric quartz grains, with straight grain boundaries and 120° triple junctions in the Aar massif 595 microstructures closely resemble the experimental ones developed under coseismic conditions (Trepmann 596 et al., 2007).

597 In the case of the polymineralic granitoid host rock, similar initial hydrofracturing together with grain 598 refinement has to occur (either seismic or aseismic) in order to provide an incipient mechanical 599 discontinuity at both low and high temperatures (Fig. 13b, steps 4 and 6). For the entire range in 600 temperatures, this initial localization of deformation is fundamental for providing fine-grained 601 polymineralic aggregates controlling subsequent deformation by frictional granular flow and by cycles of 602 frictional/viscous granular flow at low and high temperatures, respectively (Figs 5, 8 and 9). In this way, a 603 first significant stress drop occurs (Fig. 13b, steps 4 and 6), ending with a polymineralic aggregate either 604 staying non-cohesive in the case of LT-interseismic creep or gaining strength owing to coupled-grain coarsening at the HT-equivalent (Fig. 13b, steps 7 and 8), though the latter is limited in grain size increase 605 606 because of the typically slow coarsening kinetics in polymineralic aggregates (e.g., Herwegh et al., 2011 607 and references therein). In both cases, velocity strengthening will take place during interseismic slip but 608 owing to the fine-grained polymineralic aggregates, grain-size dependent granular flow guarantees shear 609 stresses much lower than those of the coarse-grained host rocks. Under the presence of fluids, build up of 610 pore fluid pressures will continue until the next seismic failure occurs (Fig. 13b steps 4 and 5 for LT as 611 well as of steps 9 and 8 for HT). In the case of high temperature deformation, thermal pore fluid 612 pressurization together with seismic shaking may help to generate instantaneous loss of intergranular 613 cohesion in the ultramylonties, allowing fast slip by velocity weakening as well as the injection of the now 614 non-cohesive polymineralic matrix into newly created off-fault fractures (Fig. 8b). In this way, the 615 generation of the fine-grained polymineralic matrices have three important consequences: (i) they reduce 616 the flow stresses during interseismic creep owing to grain-size-sensitive viscous granular flow rheology, 617 (ii) their deformation results in a build up of pore fluid pressure in cavities, which is necessary to induce a 618 new seismic event and last but not least (iii), once formed, they force strain to stay within the first order 619 mechanical discontinuities for the remaining deformation and exhumation history.

Note that besides rises in pore fluid pressures, the occurrence of brittle seismic events at dominantly ductile crustal depths can also be explained by the friction to flow behavior (Shimamoto and Noda 2014) or ductile shear fracturing (Weinberg and Regenauer-Lieb, 2009). We cannot rule out these processes but our data clearly indicate the presence of fluids during seismic events in the area studied.

624 6 Conclusions

Under upper greenschist metamorphic conditions the investigated fault zones dominantly deform by ductile deformation, evidenced by the occurrence of quartz mylonites and, more importantly, by mylonites and ultramylonites with granitic composition. In addition to viscous granular flow, the presence of biotitecoated fractures, cataclasites and injection structures of non-cohesive gouge material into weakly deformed host rocks indicate periods of brittle deformation at temperatures up to 450°C and depths of about 18 - 20 km. Non-cohesive gouges must have been generated by fast deformation and considerable energy releases at such HT-conditions but they are subsequently overprinted by ductile deformation resulting in repeated brittle-ductile cycles (see Figs. 12, 13 and 14). With progressive exhumation and cooling, deformation further localizes along these existing mechanical discontinuities and viscous granular flow becomes steadily replaced by cataclastic flow inside the fine-grained polymineralic fault rocks, where cycles of slow cataclastic flow and fast velocity weakening during elevated pore fluid pressures characterize the cycles of aseismic and seismic deformation.

- 637 In the past, the FVT in the Earth's crust has often been defined by means of a simplified quartz rheology at a depth interval of around 10-12 km (e.g., Fagereng and Toy, 2011). In the case of the granitoid crust 638 639 (granites and gneisses) however, our study clearly indicates that quartz deformation plays a subordinate 640 role and ultra-fine grained polymineralic fault rocks prevail the deformation structures encountered in the 641 field. Hence at the crustal scale, they represent the most important rheology controlling material for which 642 (under the presence of fluids) the lower limit of the FVT has to be expanded down to depths of up to 643 20 km. The FVT is therefore not as sharp as suggested in previous studies. In this sense, our provided field 644 evidence supports the concept of Shimamoto and Nuda (2014), suggesting a rather gradual transition 645 between brittle and ductile flow over a considerable depth range. Despite the importance of major rock 646 forming minerals in terms of crustal strength profiles (e.g., Ebert et al., 2008; Herwegh et al. 2011), recent 647 studies increasingly point to the importance of an improved understanding of fine-grained polymineralic 648 systems not only in the crust, but also in the upper mantle (e.g., Linckens et al. 2011, 2015). In this sense, 649 the rheological evaluation of such polymineralic fault rocks is crucial in advancing understanding of 650 deformation behavior at various crustal levels. In this light, future experiments on polymineralic granitoids 651 are of great importance to improve our rheological understanding of deformation in the middle crust.
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Fig. 1: (a) Conceptual fault-zone model (see Fagereng and Toy, 2011 and reference therein); (b) Schematic crustal strength profiles which are based on experimental data of monomineralic rocks (i.e. quartz and feldspar); (c) Schematic crustal strength profiles for different time intervals. The long-term behavior averages over time intervals of 10³-10⁵ years (Handy et al. 2007).



Fig. 2: Geological map of the upper Haslital area with ductile shear zones (grey lines) that show a width larger than 10 cm. Indicated are the two sample locations mentioned specifically in the text.





Fig. 3: (a) Graph showing changes in deformation density for each lithological unit, indicated as number of both cataclasites and mylonites per km. Note that the width of cataclasites is mostly restricted to few mm to cm, whereas mylonites may have widths up to several meters. CAGR: Central Aar granite, GrGr: Grimsel granodiorite, GZ: Grimsel Zone, SWAGR: Southwestern Aar granite. (b) Outcrop photo of a small ductile shear zone in the CAGr. An asymmetric strain gradient is manifest by a transition from schistose (left part of photo) to ultramylonitic fabric (central part of photo) and the contact to weakly deformed rock (left). (c) Subsurface outcrop photo of a brittle fault zone characterized by a fault gouge.







Fig. 4: Micrographs of Central Aar granite (CAGr) host with LT-brittle deformation structures (samples 11.004-URc and 11.003-ULc). (a) Overview of a LT-cataclasite cutting through the weakly deformed host granite. (b) Detail of thin section of the LT-cataclasite with plane polarized light (b1) and inserted UV filter (b2). The blue fluorescing resin clearly mimics the recent in-situ fluid pathways in the cataclastic matrix but also in the fractured host rock. (c) Detail, where host granite is cut by a cataclasite. Furthermore, secondary fractures, filled with a fine-grained

polymineralic gouge merge out of the cataclasite and cut into the host rock. (d) Dynamically recrystallized quartz aggregate with grain boundaries that are filled with UV fluorescent resin indicate in-situ fluid permeability at the grain boundary scale within the host rock. (e) low temperature fracturing overprinting viscuous mylonites. Note the parallel overprinting.



Fig. 5: Micrographs of a ductile shear zone, which dissects the weakly deformed Central Aar granite. (a) Overview of the asymmetric strain gradient from host rock (bottom of image) to a schistose and a mylonitic fabric with a sharp contact to the host (top of image). (b) Detail of the deformation fabric of (a). Clearly visible are elongated polycrystalline quartz domains and a very fine-grained polymineralic matrix as well as a white mica band (indicated with the red arrow). (c) Backscatter electron image of the polymineralic matrix and a feldspar clast. (d) Detail of a monomineralic recrystallized quartz domain (indicated with the red arrow).



Fig. 6: Detailed study of the characteristic microstructures and textures of (a)-(c) a completely dynamically recrystallized monomineralic quartz domain and (d)-(h) of polymineralic layers deformed by viscous granular flow (Sample Gr10). (a) Grain orientation map (inverse pole figure map) obtained from EBSD measurements, (b) related pole figure diagrams with [c]-axis raw data; (d) contoured [c]- and <a> axis data (lower hemisphere, equal-area, linear contouring) of a dynamically recrystallized pure quartz aggregate undergoing dislocation creep. (d) BSE image showing mica-rich bands (bright grey) with fine-grained polymineralic layers. (e) Image quality map and (f) grain orientation map of finely dispersed quartz in a polymineralic aggregate (compare (e) and (f)). Stereographic pole figures with (g) [c]-axis raw data and (h) contoured [c]- and <a> axis data. Note the smaller quartz grain sizes and the absence of a crystallographic preferred orientation in case of (a),(h) compared to (f),(c). Step size: 5µm. Mean = area-weighted mean grain size; n = number of measurements, max = maximum intensity of [c] pole figure. L = stretching lineation; fn = foliation normal.



Fig. 7: (a) Field photo (left) and sketch (right) from a ductile shear zone cutting discrete fractures in the weakly deformed Central Aare-granite (CAGr). (b) Micrograph of brittle fractures cutting a monomineralic dynamically recrystallized ductile quartz (Qtz) fabric. Bourder is shown by K-feldspar crystals (Kfs). Some fractures show a polymineralic infill. (c,d) Detail of a transgranular (c) and intergranular fracture (d), where biotite (Bt) has precipitated, within a dynamically recrystallized monomineralic quartz domain (samples 12.004-H-UR and gts1; see Table 2).



Fig. 8: (a-d) Micrographs of a HT-cataclasite from the Grimsel Zone (Sample GR47; Table 2). (a) Overview, where cataclasites cut through the host granitoid. Rectangles in (a) indicate locations of figures (b-f). (b) Detail showing secondary fractures with injections of the extremely fine-grained polymineralic matrix. (c) Detail of the extremely fine-grained polymineralic matrix with overgrowing biotite and epidote veins. (d) Micro-epidote veins are folded within the polymineralic matrix. (e,f) Back-scatter electron images of the contact (stippled) between the polymineralic matrix and the weakly deformed host (e) and an ultrafine-grained, polymineralic matrix (f). (Wm: white mica, Qtz: quartz, Ab: Albite, Ep: Epidote, Bt: biotite)



Fig. 9: (a) Sketch of the strain gradient in a fault zone at the contact of the Grimsel Zone to the aplitic body (top row), photographs of the analyzed samples (central row) and corresponding spatial distributions of isocompositional domains (bottom row; for color code see legend in (b)). Indicated are the locations of the samples along the strain gradient and the analysis areas on the samples (Sample Serie from Swiss-coord. 668890/157144 to 668942/157227, see also Table 2). (b) Harker diagram for CaO to SiO2 wt.% data from this study (solid symbols) in combination with literature data (transparent symbols) from Goncalves et al. (2012) and Keusen et al. (1989). Data has been separated on the basis of structures. Additionally, three groups of similar CaO content exist. Note that the mylonites and host rock fall within a domain of similar composition although mica content may vary as observed in (a). The transparent area in the graph represents the total range of compositions, which in case of the HT-cataclasite is given by the spread within own measurements. Noteworthy is also the epidotized GrGr sample from Keusen et al. (1989), which shows some similarities with our sample from the Grimsel Zone.



Fig. 10: Area percentages of different phases (a) and the resulting water content (b) for each structure and a reference granitoid host within both Central Aar granite and Grimsel Zone.



Fig. 11: Harker diagrams for each structure and the host granitoid for (a) Na_2O_2 , (b) Al_2O_3 , (c) K_2O and (d) CaO. Data points are representative averages for each structure. The bars in the Harker diagram show the total spread in analysed data with respect to the representative average.



primary localization incipient fractures

Fig. 12: Structural evolution between frictional and viscuous deformation and its cyclcic behaviour. Compare figures 4, 5 and 8 for the observed microstructures.



Fig. 13: Linking microstructures with rheological profiles (a) the theoretical evolution for quartz: Upper diagram shear stress verses depth; lower diagram shear stress and fluid pore pressure versus time. (b) the theoretical evolution for granitoid bulk rock: same type of diagrams as in (a). (c) a conceptual model for granitoid rheology and the FVT



Fig. 14: Overview of the FVT transition and the role of chemical differences in a continuous process. (a) sketch showing the relationship between strain rate, temperature and fluid pressure and the deformation processes. (b) column of microphotographs in corssed polarized light, (c) BSE overview images, (d) Zoom-in from column (c) showing detail structures (BSE-images). First row show the HT-catalcaslites with fracturing producing injection veins rich in Epidote (see also Fig. 8). Second row show intermediate stage developing bands out of the injected Ep/Ab bands. Third row show the mylonitic structure with some relics of such Ep-rich bands. Compare also figure 9b.

Table 1:Summary of grain sizes. Measurements are done on
representative 20-50 grains per group. Data are given
in μ m

Fault rocks	feldspar	quartz	mica *	
Host rocks	500-20000	180		
low temp.				
cataclasite:				
clasts	50-500	35-350	35-500	
matrix	1-5	1-5	1-5	
high temp.				
cataclasite:				
clasts	100-300	60-150	-	
matrix	*	1-5	1-5	
Mylonite:				
clasts	150-300	-	3-150	
matrix	10-80	10-80	~50	

- not existing; *: not measured

Table 2: Location of key samples

sample	Туре	location	Swiss Koord	Chemical	Grain	EBS
				data in Figs.	sizes	
				9, 10 and 11	Table 1	
11.004VURc	LT catacl.	GTS	667584 159912	Х	х	
GTS1100	mylonite	GTS	667584 159912	Х	х	
GR47	HT-catacl.	Grimsel pass	668890 157144	Х	х	
Gr45	schistose band	Grimsel pass	668921 157178	Х		
Gr43.1	ultramylonte	Grimsel pass	668942 157227	Х		
Gr10	mylonite	Bächli area	664650 159825			х