Microstructures and their implications for faulting processes – insights from DGLab core samples from the Gulf of Corinth

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Abstract

We have examined microstructures, mineralogical composition, geochemical alteration, and texture of four selected fault rock samples from the Deep Geodynamical Laboratory“ (DGLab) Gulf of Corinth project using optical microscopy, cathodoluminescence microscopy (CL), scanning electron microscopy (SEM), transmission electron microscopy (TEM), and synchrotron X-ray diffraction measurements. The fault core is composed of red and gray clayey gouge material and surrounded by a damage zone of brecciated limestones. Pressure solution features, calcite veins and calcite clasts in the breccia and gouge material attest the presence of paleo-fluids and fluid-driven mass transfer during deformation. Differences in CL-colors between the matrix and calcite vein cement and inside the vein cement suggest repeated infiltration of fluids with different composition from various sources (formation water and meteoric water). Twin lamellae densities estimated in calcite veins are used as paleo-piezometer. The deduced differential stress is $\sim 140\pm 70$ MPa for the older vein generation and appears to be higher than stress for the youngest veins (45 ± 23 MPa). In spite of the relatively small clay content in both samples, newly formed clay minerals have been observed in gray as well as red clayey gouge material. Differences between gray and red clay gouge material are found in fault rock

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composition, porosity and clay fabric. The proportion of chlorite in the red gouge is significantly less than that in the gray gouge whereas the initial porosity is significantly higher than in the gray gouge material. The detection of a well-oriented clay fabric in red clay gouge samples is unique in comparison to other major fault zones.

1. Introduction

The Gulf of Corinth, in western Greece, is one of the most seismically active regions in Europe and renowned for high extension rates with up to 1.5 cm/year (Moretti et al., 2003; Cornet 2007; McNeill et al., 2014). This fast opening is associated with an active seismogenic low angle detachment zone localized at depths between 6 and 12 km dipping in northern direction (Bernard et al., 1997; 2006). Five earthquakes of magnitude larger than 5.8 have been observed in this region within the last forty years (Cornet et al., 1997). Among many possible drilling targets in the Gulf of Corinth, the active Aigion fault was selected as an ideal site to investigate *in situ* fault related deformation processes by drilling into the fault zone (Cornet 2007) because this area is a place where one may expect a moderate to large earthquake ($M > 6$) to occur in the coming decades (Bernard et al., 2006).

The Deep Geodynamical Laboratory“ (DGLab) Gulf of Corinth project was part of a set of projects clustered under the generic name “Corinth Rift Laboratory“ to test fundamental questions regarding earthquake and fault mechanics (Cornet et al., 2004). One of the primary objectives of the DGLab project was to understand the relationships between faults, fluid flow and strain in a seismic zone. In addition, fault-healing processes that might affect the fault were also part of the investigation (Moretti et al., 2002). The borehole (AG10) was drilled in July and August 2002 (Cornet et al., 2004). Previous investigations on core samples provided a macroscopic description of fault rocks (Daniel et al., 2004; Micarelli et al., 2006). In addition,

50 fault gouge material was used for the experimental characterization of the thermo-poro-
51 mechanical properties (Sulem et al., 2004; 2005).

52 Although the AG10 was already drilled in 2002, a microstructural analysis of fault core
53 material has only recently been conducted. Core samples from the DGLab borehole provide a
54 unique opportunity to analyze microstructures of fresh fault rocks representing the state of the
55 fault before it ruptures. In this paper, we provide an overview of dominant microstructures at
56 micro-to-nano scale (e.g. brittle fracturing, dissolution precipitation processes, twinning,
57 intracrystalline plasticity, nano-porosity) in four selected samples from damage zone and fault
58 gouge. The observations were made using optical microscopy, scanning electron microscopy
59 (SEM), high-resolution transmission electron microscopy (HRTEM) techniques with focused ion
60 beam preparation and synchrotron X-ray diffraction measurements. Based on these
61 investigations we discuss our results with regard to faulting processes and compare our findings
62 with results from the Taiwan Chelungpu fault Drilling project (TCDP), the San Andreas Fault
63 Observatory at Depth (SAFOD) and the Japan Trench Fast Drilling project (JFAST), which
64 represent co-seismic weakening and creeping processes respectively, thereby facilitating a
65 comparison of faults from different tectonic regimes.

66

67 **2. Geological setting**

68 The Gulf of Corinth is a 120 km long, 30 km wide asymmetric marine basin (graben) that lies
69 within the Aegean extensional province (Fig. 1). The graben separates the Peloponnesus south of
70 the Gulf from the rest of mainland Greece (Roberts and Stewart 1994). The basin is
71 characterized by active N-S directed extensional tectonics (10 mm/y) accommodated by WNW-
72 ESE and ENE-WSW striking normal faults (e.g. Helike, Aigion; Roberts and Koulouvelas 1996).
73 The faults generally dip to the north with maximum cumulative fault throws of around 3 km
74 (Roberts 1996). They have produced destructive earthquakes in the past centuries (Chery 2001).

75 Earthquake clusters and their related fault plane solutions, located between 6 and 11 km depth,
76 suggest the existence of a detachment zone beneath the Gulf of Corinth. All the large active
77 normal faults appear to connect downwards with the proposed northward dipping detachment
78 fault at around 10 km (Rigo et al., 1996). The Gulf of Corinth therefore presents a remarkable
79 example of an association between high-angle and low-angle faulting during the same tectonic
80 phase (Chery 2001). Most of the normal fault zones are developed within Mesozoic carbonates
81 although a small number of faults affect Mesozoic ophiolites and siliciclastic rocks. Existing
82 fault slip data indicate that the orientation of the inferred stress field varies along faults and
83 earthquake ruptures (Roberts and Ganas 2000).

84 The Aigion normal fault is one of the youngest normal faults in the Gulf of Corinth (McNeill
85 2014) and belongs to a system of WNW-ESE and ENE-WSW striking active faults that are
86 related to the opening phase of the Gulf of Corinth, which started during Pliocene (Micarelli et
87 al., 2003). At the surface, the 12 km long fault forms a fault escarpment from about 40-100 m
88 with an average slope angle of 37° (Koukouvelas and Doutsos 1996). It was slightly reactivated
89 during the magnitude 6.2 Aigion offshore earthquake that occurred in 1995 (Cornet et al., 2004).
90 The current slip rate is about 2-5 mm/year and the vertical offset is about 150 m (Moretti et al.,
91 2003). The upper portion of the Aigion fault transects a sequence of Cretaceous carbonate rocks
92 and Plio-quaternary clastic sediments (Rettenmaier et al., 2004). Below the fault, brecciated
93 limestones constitute the damaged zone. The stratigraphic sequence is shown in Fig. 1 and 2.

94 The DGLab-project is centered on the south shore of the Corinth rift near the city of Aigion
95 (Fig.1). Four wells, 500 to 1200 meters deep, intersecting the active Aigion-fault have been
96 proposed, but only the 1000-m deep well was realized. The cored section between 710 and 791
97 m is a sequence of Cretaceous carbonate rocks (Olonos-Pindos platy limestone formation)
98 intercalated with several cataclastic bands, marls and thin shaly layers (Daniel et al., 2004).
99 Pressure solution features and pressure shadows in calcite crystals of the cataclastic bands are

100 interpreted as indicators for dip-slip displacement (Cornet et al., 2003). At 760 m the Aigion
101 fault was crossed by the borehole. The dip of the fault contact (55-60° N) between the hanging
102 wall (limestone) and the footwall (radiolarite) is coherent with the dip of the Aigion fault plane
103 on the surface (Rettenmaier et al., 2004). The fault offset constitutes a hydraulic barrier that
104 sustained a 0.5 MPa differential fluid pressure (Cornet et al., 2004).

105 The fault core is about 1 m thick and is composed of clay-rich material surrounded by a damage
106 zone of brecciated limestones that are affected by pressure solution processes (Sulem et al.,
107 2004). The thickness of the damage zone is 3 m above the fault and 9 m below (Daniel et al.,
108 2004). The clayey core is divided by a shear plane into two distinct parts recognizable by the
109 gray color of the upper part, hereafter referred to as “gray clayey gouge” and the red brown color
110 of the lower one, hereafter referred to as “red clayey gouge” (Fig. 2, Sulem et al., 2005). The
111 origin of clay is a matter of controversy. Daniel et al. (2004) assumed that the clay is derived
112 from radiolarites, whereas Koukouvelas and Papoulis (2009) suggested the underlying Pindos
113 Unit flysch as the probable source of clay minerals.

114

115 **3. Methods**

116 First, microstructures in the core samples were inspected optically in thin sections. If possible,
117 we quantified density of calcite twins within the calcite grains to arrive at an estimate of the
118 paleo-stresses governing deformation after vein formation. The density of twins was measured
119 using an optical microscope (Leica DM RX) with an attached high-resolution digital camera
120 (Leica DFC 420). Measurements were performed on polished thin sections. To determine the
121 twin density, the number of twins perpendicular to the twin boundaries of individual grains was
122 counted and normalized to a unit length of 1 mm (compare Rowe and Rutter 1990). For grains
123 containing multiple twin sets we applied this procedure on each orientation and calculated the

average density. We restricted the measurement of twin densities to twins visible by optical microscopy ($\geq 1 \mu\text{m}$). In order to estimate the peak differential stress we use the twin density piezometer proposed by Rybacki et al. (2013):

$$\sigma = (19,5 \pm 9,8) \cdot \sqrt{N_L}; N_L = \text{twin density } (N_L = \# \text{ twins/mm})$$

Cathodoluminescence (CL) microscopy was used to identify different cement zonations. Variations in the cathodoluminescent properties of carbonates are usually attributed to a different trace element composition of the original fluids from which the calcite precipitated.

A Zeiss FESEM Ultra 55 Plus (Schottky-type field emission scanning electron microscope) complemented with an energy-dispersive system by Thermo Fisher Scientific (UltraDry silicon-drift detector) and a FEI Tecnai G2 F20 X-Twin transmission electron microscope (TEM/AEM) equipped with a Gatan Tridiem energy filter, a Fishione high-angle annular dark field detector (HAADF), and an energy dispersive X-ray analyzer (EDS) were used to analyze microstructures at micro- and nano-scale and to determine chemical composition. To avoid preparation-induced damage, the samples for TEM studies were prepared with a focused ion beam (FIB) device (FEI FIB200TEM) at GeoForschungsZentrum (GFZ; for more details see Wirth 2004 and 2009).

To determine the crystallographic orientation of clay minerals, synchrotron X-ray diffraction measurements were conducted at the high-energy beam line ID-11C of the Advanced Photon Source (APS) at Argonne National Laboratory. A monochromatic X-ray beam with a wavelength of $0.10803 \text{ \AA}$ and $0.7 \times 0.7 \text{ mm}$ in size was used. Diffraction images were recorded with a Perkin Elmer amorphous silicon large area detector with 2048×2048 pixels and mounted about 2110 mm from the sample. Cylindrical samples, 2 mm in diameter, were mounted on a goniometer with the axis perpendicular to the incoming X-ray. During the 30 s exposure the sample was translated $1\text{--}2 \text{ mm}$ parallel to the cylinder axis to provide a better volume average. The sample was rotated in increments of 15° from -90° to $+90^\circ$ producing 12 images to derive the preferred orientation pattern. For calibration purposes an image of a CeO_2 standard was

149 recorded at identical conditions. Details of sample preparation and diffraction measurements are
150 described in Wenk et al. (2010). The images were then analyzed with the Rietveld method to
151 quantify orientation distributions and phase volume fractions. The program MAUD (Materials
152 Analysis Using Diffraction) was used because it provides the best options for texture analysis
153 (Lutterotti et al., 1997). For texture analysis the EWIMV method was employed. Step-by-step
154 tutorials have been developed for synchrotron images (Lutterotti et al., 2014; Wenk et al., 2014).

155 The crystallographic orientation of calcite occurring in veins was determined in a colloidal-
156 silica polished thin-section by electron backscatter diffraction (EBSD, compare Adams et al.,
157 1993). We have used the FEI Quanta 3D FEG SEM/FIB equipped with an EDAX–TSL Digiview
158 IV EBSD detector and TSL–OIM system (v. 5.31) also operated at GFZ. The operating
159 parameters include an accelerating voltage of 20 kV, beam current of 8 nA, working distance of
160 12 mm and a step size of 10 μm . Post-indexation filtering included the grain confidence index
161 (CI) standardization using a grain tolerance angle of 5° and the neighbor CI correlation,
162 assuming a CI of 0.1. Only data with confidence index >0.2 was considered in further
163 calculations. Orientation distribution functions and pole figures were calculated with the
164 harmonic method, expanding the series to 24 and Gaussian half-width of 8° .

165 4. Results

166 4.1. Sample composition and macroscopic description

167 Four fault rock samples obtained from the fault gouge and the adjacent limestone breccia (Fig.
168 2) of DGLab cores were used to analyze the microstructural record (referred to as **98**, **99**, **100**
169 and **101**). The mineralogical composition of the selected fault gouge samples estimated by XRD
170 analyses has already been reported by Sulem et al. (2004 and 2005) and is summarized in Table
171 1 (the mineral composition is normalized to 100%).

172 **Sample 98** was taken from a strongly brecciated band at 759,70 m (Fig. 2), which belongs to
173 the Olonos-Pindos platy limestone sequence. The inspected pure limestone shows no

174 macroscopic fossils. The fault breccia sample is completely healed by calcite cement and several
175 veins of different thickness with crosscutting relations can be distinguished (Fig. 3a, see below).
176 The matrix is composed of a fine-grained micritic carbonate mixed with greenish-gray angular to
177 subrounded carbonate fragments. The fragments show little if any evidence of frictional attrition.
178 The size of fragments ranges from a few millimeters to several centimeters. The transition
179 between brecciated limestone and the gray clayey fault-gouge zone is sharp and distinct.

180 **Sample 99** was collected from the gray clayey gouge zone at 760 m. The dark gray sample
181 with many fragments of different size and thin veins contains 55% quartz and 35% chlorite and
182 amorphous material (not included in Table 1a; Figs. 2 and 3b). In addition, calcite (7%) and
183 albite (2%) are further mineral constituents of the sample (Table 1a). Minor chrysotile was
184 recognized only by High-Resolution Electron Microscopy (HREM) imaging and TEM-EDS
185 analyses. Polished slip surfaces with striation on the slip plane are macroscopically discernable.

186 **Sample 100 and 101** were taken from the red clayey gouge (Fig. 2 and 3c; Table 1b). The
187 contact between gray and red gouge is marked by a shear plane. Both red samples are
188 characterized by a high amount of quartz (71%) due to the presence of silicified biological
189 fragments (radiolaria; Fig. 3f,g). Further mineral constituents of the red gouge samples are illite
190 (11%), albite (10%), and hematite (3%) responsible for the red coloration. The proportion of
191 chlorite (2,3%) is significantly smaller compared to the gray clayey gouge. SEM analyses
192 allowed the detection of chamosite (Fig. 4). Calcite was not detected by XRD analyses but was
193 visible under the cathodoluminescence microscope. Minor amount of chrysotile and newly
194 formed apatite were detected by TEM-EDS analyses.

195

196 4.2. Microstructural analysis

197 In contrast to the mesostructural deformation patterns, microstructures reveal considerable
198 differences in deformation intensity between fault gouge samples and brecciated limestone

199 sample. Microscopic observations have been predominantly performed on the brecciated
200 limestone sample because calcite grains have the required size to characterize the different types
201 of cement and to observe microstructures. The gray and red clayey gouge samples are usually
202 too fine-grained and SEM and TEM were mainly used for the microstructural characterization of
203 clayey gouge samples.

204

205 4.2.1. Limestone breccia

206 At microscopic scale, the brecciated limestone **sample 98** contains a micritic matrix, veins,
207 pores filled with sparry cement (Fig. 3a,d) and non-carbonated detrital grains (mostly quartz).
208 CL microscopic observations show conformities and differences in CL colors between vein
209 cement and host rock and also inside the vein cement (Fig. 5a-b, right columns). Variations in
210 the CL-colors of calcite cements are attributed to the initial trace element composition in the
211 fluid and/or to changes in redox conditions at the site of precipitation (Meyers 1974). For
212 example, the outside edge of the matrix calcite crystals has a brighter CL color (Fig 5b; right
213 column), indicating more luminescence activators (Mn) in the rim. At least three vein
214 generations are distinguishable (Fig. 5a-b). The older generation (CV₁) is made of dull calcite
215 cement with coarse grain sizes (Fig. 3a). It luminesces in the same dark orange color as the fine-
216 grained micritic matrix (Fig. 5a, right column). The CV₁ generation is cross cut by veins (CV₂)
217 with red/orange CL colors, slightly darker than the host rock color (Fig. 5a, right column). The
218 youngest veins (CV₃) are filled with euhedral blocky calcite and calcite with columnar habit at
219 the margin of the vein (Figs. 3d and 5b, right column). Columnar calcites show growth zones
220 with different CL-colors. The central zone luminesces brightly orange whereas the luminescence
221 color of the outer zone is dull orange (Fig 5b; right column).

222 Calcite grains of the older vein generation (CV₁) are intensely twinned implying that they
223 suffered a strong (fault-related?) deformation. The twinned crystals display one or two sets of

224 straight or bent twins. The twin densities measured in the sample vary between 10 and 94
225 twins/mm with a mean density of 54 twins/mm. Using twin density piezometry, the peak
226 differential stress is about 140 ± 70 MPa (Fig. 6). The high uncertainty may results from local
227 stress heterogeneity or from different deformation events recorded in the microstructure (for the
228 methodological constraints compare Rybacki et al., 2011; 2013). Twin densities measured in
229 CV₂ calcite grains vary between 9 and 79 twins/mm with a mean density of 35 twins/mm and the
230 peak differential stress is about 115 ± 60 MPa (Fig. 6). Vein filling of the youngest generation
231 (CV₃; fibrous and blocky calcite) is less twinned or not twinned, suggesting that these calcite
232 crystals underwent a weaker deformation than the older calcite cement generations. The mean
233 twin density is 5 twins/mm and the peak differential stress is about 45 ± 23 MPa (Fig. 6).
234 Compared to the in-situ state of stress data, the estimated peak differential stress values for CV₁
235 and CV₂ are unusually high (see discussion).

236 Pressure solution seams in the matrix and dissolved calcite grain boundaries suggest
237 considerable activity of dissolution-precipitation processes (inset in Fig. 7a). In SEM images
238 pores represent unconnected voids in 2D, mainly located between calcite grains and around
239 larger quartz grains or grain fragments (Fig. 7a). The grains are commonly sub-angular and do
240 not display a preferred alignment.

241 We also have measured the crystallographic orientation of calcite occurring in veins cross-
242 cutting the main foliation in our samples (Fig. 8a). The predominantly coarse-grained (CV₁) vein
243 (bottom SW and top NE), with grain sizes varying between 0.4 mm – 4 mm, contains crystals
244 with well-developed subgrain boundaries (Fig. 8a). This vein is then crosscut irregularly by a
245 second vein generation (CV₃), with grains varying from 70 to 300 μ m (central part of the
246 orientation map, Fig. 8a). The smaller crystals in this portion of the map also present subgrain
247 boundaries, as indicated by the low angle grain boundaries drawn on the map (Fig. 8a). The
248 calcite crystallographic preferred orientation (CPO) in the coarse-grained vein (CV₁) is weak and

249 is characterized by three maxima of [0001] distributed symmetrically along a great circle, with
250 the <2-1-10> maxima parallel to the center of the net and two secondary (weaker) maxima (Fig.
251 8b). Although the multiples of random distribution indicates the presence of a weak CPO in this
252 sample, the maxima of 3.5 likely result from the insufficient number of grains on the scanned
253 area for this calcite vein and the crystallographic orientation is probably close to random. The
254 [0001] axes of calcite of fine-grained veins (CV₃) is characterized by a incomplete, broad single
255 girdle along NW-SW in the pole figure (Fig. 8c), while the <2-1-10> are distributed along broad,
256 great circle girdles, normal to one of the [0001] maxima.

257

258 4.2.2. *Gray clay gouge*

259 The strongly fractured gray clayey gouge **sample 99** is composed of angular to subrounded
260 large quartz (cryptocrystalline chert/radiolarite) and calcite fragments embedded in a dark-brown,
261 fine-grained clayey gouge matrix (grain size <1 μm; Fig. 3b and 3e, Fig. 5c, 7b). Some fragments
262 and matrix patches have a preferred aligned orientation exhibiting a weak fabric (Fig. 3b). TEM
263 images of this sample reveal a densely packed fabric composed of clay particles and quartz
264 patches (Fig. 9a), which consist of small recrystallized quartz grains (Fig. 9c). Few patches are
265 wrapped by thin sheet silicates (Fig. 9c), resembling clay-clast aggregates (CCAs; Boutareaud et
266 al., 2008). At some places, dislocation cores ending at the surface of the quartz grains are
267 dissolved and newly formed clay minerals are growing into the open spaces (Fig. 9e). The few
268 calcite grains are characterized by highly sutured grain boundaries indicating pressure solution
269 (Fig. 7b). They are intensely twinned with a similar twin density (49 twins/mm) as in the older
270 vein generation of sample 98. Accordingly, twinning paleo-piezometry yields a peak differential
271 stress magnitude of 137±68 MPa, comparable with calcite grains of the older vein cement (CV₁)
272 in sample 98 (Fig. 6). However, the calcite grains contain no or only few dislocations, whereas
273 the twin lamella contain dislocation clusters (Fig. 9f). Under CL, the calcite fragments

luminesces yellow-orange and the outer growth rim has a dark-orange color (Fig. 5c), suggesting a precipitation from formation water under burial conditions (Mayer 1978). The gray gouge sample is too heterogeneous to analyze the crystallographic preferred orientation of clay minerals.

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4.2.3. Red clay gouge

Similar to sample 99, **samples 100 and 101** consist of angular to subrounded fragments and a red-brown, fine-grained clayey gouge matrix (Fig. 3c,f and g; 5d; 7c). Microscopically, some patches of the gouge matrix show a weak foliation defined by preferred orientation of phyllosilicates (Fig. 3c). Besides chert and matrix fragments (Fig. 7c), a high amount of round radiolarian chert is visible in thin sections and SEM images (Fig. 3g and 7d). CL – microphotographs show orange luminescent calcite particles, even though XRD analyses revealed no calcite. Quartz grains in chert fragments luminesce blue (Fig. 5d), indicating a hydrothermal origin (Götze et al., 2001). In contrast to sample 99, the clay matrix of the red gouge is characterized by stacked clay flakes arranged in a card-house fabric with open pore spaces between the flakes (Figs. 9b and 9d). The initial porosity of the material is about 19.7% and therefore significantly higher than in the gray clayey gouge sample (Cornet et al., 2003). The stacked phyllosilicates domains are locally aligned parallel to each other. They are not or only slightly bent suggesting that they were newly formed (Fig. 7f and 9d). At other places, SEM observations reveal a fabric with bent and folded clay flakes resembling poorly developed *S-C* fabrics (Fig. 7e). Quantitative crystallographic preferred orientation analyses were performed on two samples of the red gouge of two adjacent regions. Phase fractions from Rietveld analysis are similar to XRD analysis with dominating quartz (49-65 wt%), illite-mica (15-48 wt%), and chlorite (3-19 wt%). Pole figures are displayed for illite-mica and chlorite (Fig. 10). The orientation patterns for both phyllosilicate minerals are very similar. All pole figures show

299 relatively strong fabrics (with maxima up to 3.7 m.r.d.), but especially in sample 100 with
300 considerable local heterogeneity (spot 1 versus spot 2). In sample 101 the selected areas were
301 more homogeneous. The preferred orientation of both minerals in the selected areas is significant.
302
303

304 **5. Discussion and conclusion**

305 The microstructural and mineralogical evidence presented above reveal different deformation
306 mechanisms and preferred orientation patterns in the three analyzed fault components (limestone
307 breccia with veins/damage zone; gray gouge/fault core; red gouge fault core). Published data
308 from neighboring faults and international drilling projects are used to draw general conclusions
309 for faulting processes.
310

311 *5.1. Limestone breccia*

312 In the limestone breccia above the fault contact, dissolution-precipitation processes act as
313 dominating deformation mechanism as is demonstrated by pressure solution seams, dissolution
314 along grain boundaries and healed (cemented) fractures. Taking into account the small number
315 of grains, we still observed a weak preferred orientation of calcite in the coarse-grained veins.
316 Previous texture analyses of calcite fabric patterns in SAFOD samples reveal that aseismic creep
317 promotes the development of a weak to moderate calcite fabric (Janssen et al., 2014). In contrast,
318 Verberne et al., 2013 described nanocrystalline slip zones in calcite fault gouge that are
319 characterized by grain size reduction and an intense crystallographic preferred orientation.

320 In SEM images, many pores between the vein calcite grains appear to be unconnected in 2D.
321 They were possibly filled with fluids but we found no evidence to suggest that elevated fluid
322 pressure existed. Such a possible high pore fluid pressure could reduce fault strength very
323 effectively (Hubbert and Rubey 1959). Similar CL-colors in matrix and calcite veins and

324 differences in CL-colors inside the vein cement suggest repeated infiltration of fluids with
325 similar and/or different composition from various sources. For instance, the correspondence in
326 Cl-color between host rock and CV₁ vein suggests that the vein calcite was locally derived from
327 the matrix (Götze 2000). The dull-to-bright luminescent color of columnar calcites of CV₃
328 indicates that this cement did not precipitate from near-surface meteoric water under oxidizing
329 conditions (Verhaert et al., 2004). Therefore, it is more likely that the younger veins precipitated
330 under reducing conditions typical for a deep-burial environment (Boggs and Krinsly, 2010).
331 Labaume et al. (2004) documented similar variations in vein calcites (and related fluid sources)
332 of the neighboring Pirgaki fault. However, they suggest that the younger vein calcite is related to
333 circulation of external (meteoric) water during both seismic and interseismic periods, suggesting
334 a hydraulic connection to the surface and hydrostatic fluid pressure. In contrast, first generation
335 vein calcite precipitated from fluids with a formation water composition (Labaume et. al. 2004).
336 Bussolotto et al. (2015) suggest that all calcite cements in fault rocks from Corinth rift normal
337 faults precipitated from meteoric fluids in a close or open circulation system depending on depth.
338 Also, stable isotope and thermometry data indicate a development from a closed system with
339 fluid-rock equilibrium during brecciation to a more open one with influx of meteoric water
340 (Benedicto et al., 2008). Helium isotopic ratios of fossil fluids trapped in both calcite generations
341 indicate an absence of a mantle-He signal in the Corinth rift fluids suggesting that the fault
342 system is rooted in the upper crust (Pik and Marty 2009).

343 In summary, fluid-assisted weakening mechanisms and fluid-assisted healing processes
344 (cementation) are the dominant microstructures in the limestone breccia. Especially mass
345 transfer by pressure solution is widespread and favor aseismic creep. Such fluid-assisted
346 weakening mechanisms are common in many fault zones (e.g. Hadizadeh et al., 2012; Richard et
347 al., 2013; Janssen et al., 2014). Bussolotto et al. (2015) suggest, that the becciation process was
348 probably responsible for a mechanical weakening of the fault zone and favored further faulting

349 deformation.

350 The unusual high stress values measured in the older veins (140 MPa; CV₁ and CV₂) may
351 indicate episodes of high local stress on the grain scale that could be attributed to former seismic
352 events. These stress values are similar to those estimated in SAFOD samples (92-251 MPa,
353 Rybacki et al., 211). The authors suggest that twinning represents the maximum local (grain-)
354 scale paleo-stresses possibly produced by localized high transient stress during seismic events.
355 The orientation of the principal stresses is locally almost parallel and perpendicular to the fault
356 axis (Sulum 2007). The considerable number of untwinned grains in the youngest calcite veins
357 (VC₃) suggests that healing processes outlasted the period of brittle faulting.

358

359 5.2. *Clay gouge*

360 Fault rock composition of the selected gray and red gouge samples reveal some similarities
361 with the mineralogy of fault rock samples from the neighboring Helike Fault, which is also
362 characterized by the presence of quartz, plagioclase, illite, calcite and Fe-oxides (Koukouvelas
363 and Papoulis 2009). However, unlike DGLab samples, no chlorite was detected. Koukouvelas
364 and Papoulis (2009) suggest for the Helike as well as Aigion and Pargaki faults a weak
365 hydrothermal alteration by meteoric water at shallow levels. Calcite (vein) fragments embedded
366 in a dark-brown, fine-grained clayey gouge matrix indicate that the gouge matrix formed at least
367 in part after the vein formation (compare Labaume et al., 2004).

368 In comparison to fault gouge samples from other drilling projects (SAFOD, TCDP, JFAST), a
369 significant difference is a relatively low clay content in general and a complete absence of
370 smectite in particular. For instance, SAFOD gouge material from the creeping section
371 contains ~70% saponite/smectite (Moore and Rymer 2012; Janssen et al., 2014). Smectite is
372 abundant (up to ~ 80%) in the principal slip zone (PSZ) of the Chelungpu fault that ruptured
373 during the 1999-Chi-chi earthquake (Kuo et al., 2009) and in samples from the plate-

374 boundary decollement zone between the Pacific and North American plate (JFAST; Janssen et
375 al., 2015; Kameda et al., 2014). Similar to SAFOD, TCDP and JFAST fault gouge samples, in
376 DGLab samples, too, clay minerals are in part newly formed, likely during inter- and post-
377 seismic phases of the earthquake cycle.

378 Based on the assumption that the clay mineral content contributes significantly to fault
379 weakening (Solum et al., 2004; Di Toro et al., 2011) the question arises to what extent the
380 relative low clay content in DGLab fault gouge samples may have influenced the mechanical
381 behavior of the Aigion fault. Friction experiments from Collettini et al. (2009) and de Paola
382 (2014) demonstrate that fault weakness can occur in cases where weak mineral phases
383 (phyllosilicates) constitute only a small percentage of the total fault rock assemblage.
384 Deformation experiments on DGLab samples show that the small clay fraction has a significant
385 influence on the thermal-mechanical properties of the fault material in terms of fluid
386 pressurization inside the Aigion fault (Sulem et al., 2004; 2005). Permeability estimates for fault
387 gouge samples vary with stress and ranges from 0.1 to $2 \times 10^{-18} \text{ m}^2$ (Cornet et al., 2004). Taking
388 into account the strong clay fabric, we suggest that the small amount of clay minerals was
389 sufficient enough to reduce the fluid permeability.

390 CCAs are regarded as possible indicators of large slip at co-seismic velocity with thermal
391 pressurization (Boutareaud et al., 2008). They are described for many clay-bearing fault gouges
392 including TCDP and SAFOD samples (Boullier et al., 2009; Janssen et al., 2014). However, their
393 presence in SAFOD samples within and outside of the deforming zones rule out that CCAs were
394 exclusively formed during seismic slip (Janssen et al., 2014); the authors suggest that the
395 observed CCAs are rather formed by rotation of clast (rolling process) over a wide range of slip
396 rates as suggested by Han and Hirose (2012).

397 The detection of a preferred clay fabric orientation of red clay gouge samples by synchrotron
398 X-ray diffraction measurements is different compared to other major fault zones. Recent

399 analyses of gouge samples conducted on several other fault zones (e.g. Alpine Fault, Warr and
400 Cox 2001; Moab Fault, Solum et al., 2005; Carbonera Fault, Solum and van der Pluijm 2009;
401 Bogd fault, Buatier et al., 2012; San Andreas Fault, Wenk et al., 2010; Janssen et al., 2014;
402 Chelungpu fault, Janssen et al., 2014) showed only very weak preferred orientation of sheet
403 silicates, regardless of their origin. It is supposed that the lack of a strong fabric in clay-rich
404 samples may be attributed to newly formed phyllosilicates, that grew in many orientations, and
405 to fault-related kinking and rotation of clay particles (Janssen et al., 2014). For the red clay
406 gouge samples, we assume that patches of oriented clay particles may be tectonically emplaced
407 into the Aigion fault.

408 Interestingly Sulem et al. (2005) found a high amount of amorphous material detected by
409 XRD analyses (20% for the red gouge and 40% for the gray gouge; not included in Tab. 1).
410 However, TEM and SEM images show no melt textures such as vesicles or amygdules, so
411 we assume that amorphous material is a product from radiolaria (the skeletal material consists
412 of amorphous silica) and was not produced by frictional melting, as described in TCDP
413 samples (e.g. Kuo et al., 2009; Otsuki et al., 2009; Janssen et al. 2014). Although the exact
414 origin of the amorphous material remains unclear, it is expected to significantly contribute
415 to fault weakening (e.g. Goldsby and Tullis 2002; Di Toro et al., 2004; Janssen et al., 2010).

416 The deformation features most frequently found in DGLab samples are pressure solution and
417 mineral alteration, produced by aseismic slip. Dynamic weakening mechanisms that occur
418 during coseismic rates (e.g. frictional melt, trace element mobility; see Rowe and Griffith,
419 2015) have not been observed. Our results are in line with observations made by Labaume
420 et al., (2004) who emphasized that aseismic deformation during the interseismic period must
421 be taken into account when describing the deformation in the recent fault zones of the
422 Corinth rift. When comparing our microstructural record with microstructures observed in
423 TCDP, SAFOD and JFAST samples it appeared that the strong clay fabric and the low clay

424 content in the gouge material is a unique feature of the DGLab samples, whereas
425 dissolution-precipitation processes occur in all drill core samples.

426

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436

437

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636
637

638 **Figures**

639 **Fig. 1.** Simplified S-N geological cross section through the Helike and Aigion fault modified
640 after Doutsos, T. & Poulimenos, G., 1992 and Giurgea et al., 2004. The insert shows the
641 geographical position of the survey site.

642 **Fig. 2.** Generalized litho-log of the AIG 10 borehole from Cornel et al., 2003 and the Aigion
643 fault core at depth 759,70m with sample position (modified after Sulem et al., 2004)

644 **Fig. 3.** Photomicrographs of thin sections (standard size of 24 x 46 mm) and photomicrographs
645 of selected sections thin sections. The sample number is in the upper left corner **(a)** Thin section
646 from the limestone breccia. **(b)** Thin section from the gray clayey gouge **(c)** Thin section from
647 the red clayey gouge. **(d)** Micritic matrix (left) and blocky and fibrous calcite. **(e)** Fissured dark-

648 brown clayey gouge matrix with embedded angular fragments. (f) Red-brown gouge matrix with
649 embedded radiolarites. (g) Accumulation of round radiolarites in the red clayey gouge matrix.

650 **Fig. 4.** Element distribution from scanning electron microscopy (SEM) at grain scale. The
651 localized concentration of Fe (green color) and the more distributed concentration of Mg (red
652 color) indicate the presence of chamosite.

653 **Fig. 5.** Photomicrographs of main microstructural features observed in thin sections under
654 parallel polarizers (left), crossed polarizers (middle) and Cathodoluminescence (CL) images of
655 the same region. Note the difference in CL-colors between the gray and red clayey gouge. (a)
656 Micritic matrix (M) with three vein generations (CV_1 - CV_3). (b) Micritic matrix with bright rim
657 and the youngest vein generation (CV_3). (c) Densely packed matrix with calcite fragments. (d)
658 Matrix with quartz grains.

659 **Fig. 6.** Estimated differential stress based on paleo-piezometry of measured twin densities in
660 calcite veins of samples 98 (CV_1 - CV_3) and sample 99.

661 **Fig. 7.** SEM images of main microstructural features. (a) Unconnected voids between calcite and
662 quartz grains. The inset shows pressure solution seams. (b) Chert and calcite fragments in the
663 gray clayey gouge matrix. (c) Chert fragments in the red clayey gouge matrix. (d) Radiolaria
664 skeletons wrapped by clay minerals. (e) Quartz grains and bent phyllosilicates. (f) Newly formed
665 phyllosilicates.

666 **Fig. 8.** Pole figures of calcite. (a-c) Crystallographic preferred orientation of older (CV_1) and
667 younger vein (CV_3) calcite. Equal area projection. RD and TD refer to electron microscope
668 coordinates.

669 **Fig. 9.** Brightfield TEM images of main microstructural features. (a) Many quartz grains and few
670 clay particles in the gray clayey gouge matrix. (b) Red clayey gouge matrix with voids between
671 clay flakes. (c) Patches of small recrystallized quartz grains, wrapped by randomly oriented clay
672 minerals. (d) Quartz, albite and stacked clay particles in the red clayey gouge matrix. Clay stacks

673 arranged in a card house fabric (e) Parallel and well oriented clay particles (bottom). The arrows
674 indicate the growth of authigenic clay mineral on a quartz grain. (f) Calcite twin lamellae in a
675 grain with many dislocations. Abbreviations: q=quartz, i=illite

676 **Fig. 10.** (001) and (100) pole figures for illite/mica and chlorite of red clayey gouge samples 100
677 and 101 determined on 2 spots. Equal area projection, contours in multiples of a random
678 distribution (m.r.d.).

679

680 **Tables**

681 **Tab. 1.** Composition of the gray and red clayey fault gouge

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Neogene-Quaternary

- Holocene: beach and river mouth deposits
- Quaternary: alluvial fan deposits, talus cones, mainly limestone-gravel
- Quaternary: fan delta deposits
- Lower Pleistocene: lacustrine-lagoonal deposits
- Upper Pliocene-Holocene: filling of the Corinth Graben, marine fine-to coarse sediments, partially fluvial deposits, high percentage of limestone gravel

Triassic-Paleogene -Olonos-Pindos Zone

- Paleogene-Upper Eocene: flysch upper part-wild flysch
- Upper Cretaceous: half pelagical deposits, shales with embedded limestone strata transition to flysch
- Upper Cretaceous: platy limestones (deep sea limestones) highly fissured and karstified
- Upper Cretaceous- Upper Jurassic: radiolarites, cherts, marly schists, siliceous limestones, first Pindos-Flysch
- Jurassic, partially Lower Cretaceous siliceous limestones, Upper Triassic: schists-cherts-platy limestones group
- Middle Triassic: flysch-like, sandy-clayey sediment

completed and modified after T. Doutsos & G. Pulimenes
by Dr. V. Giurgea, Applied Geology Karlsruhe (AGK)

depth (m)

*Figure 6

Illite-mica
(001)

Illite-mica
(100)

Chlorite
(001)

Chlorite
(100)

100-1

100-2

101-1

101-2

Table 1a: Mineral composition of gray fault clayey gouge (sample 99)

Sample	Quartz (wt.%)	Chlorite (wt.%)	Calcite (wt.%)	Albite (wt.%)	
Clay particles <80µm	49	47.4	2.5	1	
Clay matrix <400µm	56	34	6.5	2.5	
Fragments and gravels	61	24	12	2	

Table 1b: Mineral composition of red fault clayey gouge (samples 100 and 101)

Sample	Quartz (wt.%)	Chlorite (wt.%)	Illite (wt.%)	Albite (wt.%)	Hematite (wt.%)
Clay particles <80µm	74	3	10	9	3
Clay matrix <400µm	66	2	15	13	3.5
Fragments and gravels	73	2	9	8	2.5