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4 Quantifying the partition between seismic and aseismic deformation along
5 creeping and locked sections of the North Anatolian Fault, Turkey
6

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18

19 **Abstract**

20 Shallow aseismic creep is a key deformation component along plate boundaries that
21 contributes to the energy budget during the seismic cycle. Several major active
22 continental faults show spatial alternation of creeping and locked sections. The
23 present study focuses on the evaluation of the aseismic part of the total displacement
24 along the North Anatolian Fault in Turkey. Detailed microstructural analyses of finite
25 strain were performed using various methods, based on change of length or angle, on
26 six representative samples collected over 32 outcrops along locked and creeping
27 sections of the fault. Chemical analyses were used to map mineral composition of
28 fault rocks and to calculate relative volume changes associated with creep.
29 Relationship between finite strain and volume change allowed quantifying the
30 evolution of the penetrative pressure solution cleavage mechanism of creep. In
31 volcanic and analogous creeping rocks, finite strain measurements revealed two
32 spatial scales of strain that correspond to the alternation of two types of shear zones,
33 with cleavages either oblique or sub-parallel to the fault displacement. Using geodetic
34 and geological data, cumulative aseismic displacement was calculated in the range 9
35 to 49% of the total 80 km displacement in the creeping sections, and was negligible in
36 locked sections. The large uncertainty in the kilometer-width creeping sections was
37 related to the difficulty of quantifying high strain values that are associated with high
38 shear displacement and for which measurement uncertainties are large. A promising

39 way to improve such quantification would be to develop reliable statistical analysis of
40 cleavage orientation in the field.

41

42 **1. Introduction**

43 Aseismic creep has been extensively documented along several active faults
44 worldwide (Chen and Bürgmann, 2017; Harris, 2017; Bürgmann, 2018). Active fault
45 creep processes develop either as transient (mostly post-seismic) sliding or as
46 permanent sliding. Post-seismic creep rate can evolve through time with various
47 patterns: with an exponential or a power law decay, down to zero or to a residual
48 constant creep velocity (Çakir et al., 2005). In some cases, shallow creep can
49 accommodate the whole tectonic loading as, for example, along the permanent
50 creeping section of the San Andreas fault (Savage and Burford, 1973). In cases where
51 shallow creep only partly releases the tectonic loading, as is the case along the North
52 Anatolian, the Hayward or the Longitudinal Valley faults in Turkey, California and
53 Taiwan, respectively (Çakir et al., 2005, Graymer et al. 2005, Thomas et al., 2014),
54 major earthquakes may still occur at depth and propagate toward the surface.

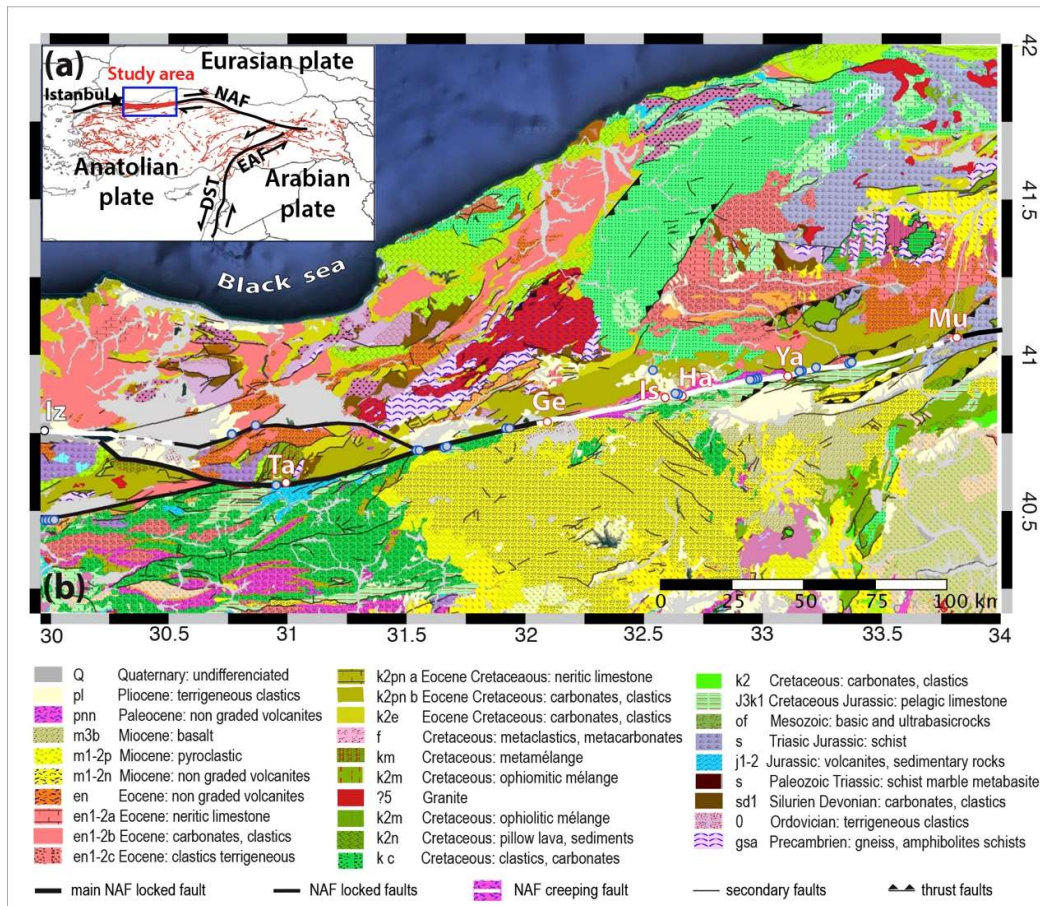
55 These contrasting creep behavior patterns may have different implications for seismic
56 hazard assessment. It is therefore crucial to understand the spatio-temporal
57 characteristics of creep and its mechanisms (Bürgmann, 2018). Rock types can partly
58 control fault creep behavior. For example, observations have shown that surface creep
59 rates along the North Anatolian Fault are almost null along fault sections that are rich
60 in massive limestones, but become significant along sections that host various kinds
61 of volcanic rocks that were softened by progressive deformation processes (Kaduri et
62 al., 2017). However, other important characteristics of creep, such as creep
63 deformation duration during a single seismic cycle and over several seismic cycles, as
64 well as the aseismic part of the total displacement during geological fault history,
65 remain to be determined more accurately.

66 To make progress in understanding creep mechanisms, the displacement associated
67 with creep processes must be evaluated. On one hand, the total displacement over the
68 lifetime of a fault accommodated by both creep and earthquakes can be evaluated by
69 measuring the offset of a number of geological markers (Emre et al., 2013). On the
70 other hand, the displacement associated with creep processes can be deduced only
71 from geodetic or geological strain measurements, quantifying the short-term
72 interseismic strain and the long-term finite strain, respectively. However, surface

73 geodetic measurements include both the irreversible (ductile creep) and reversible
74 (elastic) parts of the deformation. The distinction between creep and tectonic loading
75 is generally made through simple modeling by considering an elastic half-space
76 surrounding a fault plane (Savage and Burford, 1973; Okada, 1985). In addition,
77 geodetic measurements are representative of slip only in the recent past, covering a
78 few years to several decades. Extending geodetic displacement measurements to the
79 long term (geological times of several millions years) must include an assumption on
80 the variations of the displacement rate with time, which is almost impossible to
81 estimate, especially in areas associating seismic and aseismic processes.
82 Consequently, in order to evaluate only the irreversible part of the large finite strain
83 associated with the creep processes, complementary detailed finite strain
84 measurements in the gouge and damage zone rocks of fault zones are required.
85 The aim of the present study is to quantify the displacement related to aseismic
86 irreversible creep processes in both creeping and apparently, presently locked sections
87 along the North Anatolian Fault in Turkey. To reach this aim, one needs to measure
88 the associated geological finite strain in both the gouge and the damage zone rocks
89 **(Fig. 1)**. In order to calculate this displacement, two types of data are required: the
90 finite strain values and the width of the zone of associated shear deformation
91 (Ramsay, 1980; Ramsay and Graham, 1970). However, the measurement and the
92 interpretation of such data are rather complex because strain and fault width values
93 evolve with time and along the fault strike during deformation, since the beginning of
94 the formation of the fault to the present day. Strike-slip creeping faults accumulate
95 strain in the upper crust by plastic and viscoelastic mechanisms involving both shear
96 deformation and mass transfer, developing what is known as compaction or dilatant
97 shear zones (Ramsay, 1980; Ramsay and Graham, 1970). In such zones, the mineral
98 composition and rock fabric change due to fracturing, mass transfer, fluid circulation,
99 chemical alteration and metamorphism (Gratier et al., 2013, 2011; Steward et al.,
100 2000; Imber et al., 2001; Jefferies et al., 2006; Collettini, et al., 2009). Moreover, clay
101 gouge comprising low friction minerals such as montmorillonite and saponite also
102 develop (Carpenter et al., 2016; Kaduri et al., 2017; Lockner et al., 2011; Samuelson
103 and Spiers, 2012). In addition, fault growth is not linear with displacement. Power law
104 relationships between maximum displacement and fault width are often reported
105 (Pennacchioni, 2005; Scholz, 2002). At the outcrop scale, irreversible shear strain
106 gradients generally decrease to zero away from the fault over distances of several

107 centimeters to several kilometers, and sometimes over distances of up to 25 km
108 (Mavko, 1981). The lateral evolution of such strain profiles is often non-linear
109 (Pennacchioni, 2005). The change in shear zone width in time and space thus depends
110 on the deformation mechanisms that combine simple shear, pure shear and volume
111 change and that determine the displacement behavior (Bos and Spiers, 2002; van der
112 Pluijm and Marshak, 2010; Fossen and Cavalcante, 2017). Such heterogeneous shear
113 zones can be classified into three categories based on their width variations:
114 increasing, decreasing or constant in time (Hull, 1988). The shape of displacement
115 profiles across faults depends on whether the deformation process is strain hardening
116 (e.g. deformation diffuses into the host rock) or strain softening (e. g. deformation is
117 localized in narrow zones) (Vitale and Mazzoli, 2008). In particular, strain-softening
118 processes involving simple shear with volume loss are consistent with clay gouge
119 formation (Kaduri et al., 2017).

120 Measuring and interpreting finite strain and the width of the associated creeping shear
121 from geodetic and geological observations are challenging tasks, especially when
122 deformation involves very high strain values. Despite considerable uncertainties
123 associated with both types of observations, the data presented in this study of the
124 North Anatolian fault can be used to characterize: (i) the spatial correlation between
125 the creep-related strain measured by geodetic methods and that measured by
126 geological methods, (ii) the control of rock lithology on aseismic creep mechanisms
127 and temporal evolution, (iii) the aseismic part of the total displacement on the fault.



128

129 **Figure 1.** Structural and geological map of the study area along the North Anatolian Fault
 130 (NAF), modified from Kaduri et al. (2017). (a) Tectonic setting of study area, located east of
 131 Istanbul (blue rectangle with NAF in red), adapted from Emre et al., (2013). EAF: East
 132 Anatolian Fault, DST: Dead Sea Transform fault. (b) Geological map of the study area with
 133 outcrop locations indicated as blue or red circles for the 32 investigated outcrops. Red circles
 134 with names locate the six representative outcrops with strain markers described in the present
 135 study: Ta=Taskesti (40°34'54.00"N/ 31°1'60.00"E), Ge=Gerede (40°47'3.00"N/
 136 32°6'30.00"E), Is=Ismetpasa (40°51'55.00"N/ 32°35'41.00"E), Ha=Hamamli (40°52'12.19"N/
 137 32°39'8.78"E) Ya= Yazioreni (40°56'1.59"N/ 33°6'11.09"E), Mu=Mülayim (41°3'5.28"N/
 138 33°48'17.61"E). The presently creeping segments of Izmit=Iz (west) and Ismetpasa (east) are
 139 marked in white (Çakir et al., 2005; Cetin et al. 2014). The locked segments are marked in
 140 black. The regional geological map is extracted from an interactive map of the Turkish
 141 geological survey (Akbas et al., 2016) available at
 142 <http://yerbilimleri.mta.gov.tr/anasayfa.aspx>. The legend gives the lithology of the formations
 143 that are cut by the studied North Anatolian Fault section.

144

145 **2. Measuring creep-related deformation from fault zone outcrops: setting and**
 146 **approach**

147

148 **2.1 Seismotectonic setting**

149 The North Anatolian Fault (NAF) in Turkey is a right-lateral strike-slip plate
150 boundary. This major continental fault accommodates the relative motion between the
151 Anatolian plate, to the south, and the Eurasian plate, to the north. The NAF formed
152 approximately 13 to 11 Ma ago in the east and slowly propagated westward (Şengör,
153 et al. 2004). The long-term geological rate is 20 ± 8 mm/yr (Şengör, et al. 2004),
154 consistent with short-term geodetic rates of 24 ± 2 mm/yr measured from GPS data
155 (Reilinger et al., 2006) and 25 ± 1.5 mm/yr measured from InSAR data (Cetin et al.,
156 2014). The total cumulative displacement along the studied NAF segment (**Fig. 1**)
157 since its initiation is of the order of 80 km (Janssen, et al., 1997; Armijo et al., 1999;
158 Emre et al., 2013; Akbas et al., 2016). A sequence of large earthquakes propagated
159 from east to west and ruptured over ~ 1000 km of the fault during the 20th century (e.g.
160 Emre et al., 2016), providing evidence of the potential of the fault to accumulate large
161 seismic displacements. However, a striking feature of the NAF is the existence of two
162 slip modes along the fault. Some sections are prone to major earthquakes (Stein et al.,
163 1997) and remain locked in between earthquakes while other sections, well-identified
164 by geodesy measurements, display aseismic creep that possibly initiated as post-
165 seismic slip (Kaneko et al., 2013; Çakir et al., 2014; Cetin et al., 2014) and may
166 behave as a succession of transient creep episodes (Bilham et al., 2016; Rousset et al.,
167 2016). The relative contribution of seismic and aseismic displacement in the 80 km
168 cumulated displacement remains to be evaluated, which is the aim of the present
169 study.

170 At least two sections of the inland part of the North Anatolian Fault have been
171 identified in recent years as aseismic creeping fault sections (Ismetpasa and Izmit
172 sections, **Fig. 1**). Based on an InSAR study of Envisat satellite data along descending
173 orbits, it has been shown by Cetin et al. (2014) that the Ismetpasa creeping section
174 runs from longitude 32.2° to 34.0° , with a maximum horizontal surface creep rate of
175 ~ 20 mm/yr, located 30 km east of Ismetpasa, and a creeping depth of ~ 5 km. These
176 results were revised from a new Envisat InSAR analysis including both descending
177 and ascending data, which helped to obtain a better separation of the vertical and
178 horizontal creep components. The Ismetpasa creeping section was found to be longer
179 (running from 31.0° to 35.0°), with two peaks of maximum horizontal creep rate up to
180 ~ 14 mm/yr, and creep extending down to 9 km depth (Hussain et al., 2016). A clear
181 correlation between shallow creep rate and near-surface fault lithology has been
182 demonstrated (Kaduri et al., 2017): the locked section west of the Ismetpasa creeping

183 section is mostly composed of massive limestone and dolomite whereas the creeping
184 section is composed of volcanic and ophiolite rocks including clay-rich gouges (**Fig.**
185 **1**).

186

187 **2.2 Fault sampling and strategy analysis**

188 We calculated the aseismic part of the total displacement along both locked and
189 creeping zones of the North Anatolian Fault. The finite strain and the creeping width
190 for various lithologies were evaluated by investigating 32 outcrops, over a 400 km-
191 long section between the city of Izmit to the west and the eastern end of the Ismetpasa
192 creeping section to the east (**Fig. 1**). Satellite images and geological maps (Herece and
193 Akay, 2003, Emre et al., 2013, 2016, Akbaş, et al 2016) were used to identify all
194 accessible outcrops following the procedure described in Kaduri et al., (2017). No
195 continuous outcrop was found that would cover the whole kilometer-width of the
196 NAF shear zone. The links and the common features between these dispersed
197 outcrops is that they all show markers of creep deformation associated with the NAF
198 shear zone, either as penetrative cleavage (in volcanic or analogous rocks) or as
199 stylolites (in limestones). So they all attest of the finite aseismic strain, which was
200 associated with the large strike-slip 80 km-displacement of the NAF shear zone
201 during geological times. It must be noted that geological evidence of such creep
202 deformation was found away from the present-day creeping fault, sometimes at more
203 than one-kilometer distance. This observation indicates that the NAF was not a single
204 localized fault during geological times but a kilometer-width shear zone that include
205 several seismic faults and large associated zones of aseismic deformation, an
206 observation consistent with geological maps (Herece and Akay, 2003, Emre et al.,
207 2013, 2016, Akbaş, et al 2016). Building on our first study on the implication of fault
208 rock transformation on aseismic creep (Kaduri et al., 2017), we focus in this
209 following study on the calculation of the total cumulative aseismic displacement from
210 microstructural observations. The main challenge was not only to find outcrops but
211 also to find samples that allow calculating strain values. From the 32 visited outcrops
212 (7 outcrops in the locked segments, 25 outcrops in the creeping segment), 90 oriented
213 hand samples were collected and analyzed in the laboratory and 130 thin polished
214 sections of rock samples were prepared for mineralogical, geochemical and
215 microstructural analyses. All outcrops and samples were chosen to be representative
216 of the wide NAF shear zone and of the geological deformation associated with the

217 NAF displacement. However, estimating strain values was not possible for all these
218 outcrops and samples. Six representative outcrops with strain markers are presented
219 here, one from the locked section and five from the creeping section (**Fig. 1**). They
220 represent the various aspects of the strain measurement methods and give a
221 representative focus on the evolution from initial rock to damaged rock and gouges, in
222 order to characterize strain related to aseismic creep.

223 Strain was measured at various scales from thin section to outcrop scales. Micro- to
224 meso-structural strain measurements were performed on thin sections from millimeter
225 to decimeter scales. Chemical maps were acquired using either X-ray fluorescence
226 spectrometer (XRF, Eagle III) (mm-cm) or Wavelength Dispersive Spectroscopy on
227 an Electron Probe Micro Analyzer (WDS-EPMA) (microns-mm) (Kaduri et al.,
228 2017). Image processing techniques based on MATLAB codes were implemented to
229 calculate internal finite strain from particle distributions or deformation of geological
230 objects. Meso- to macro-structural strain measurements at meter to kilometer scales
231 were derived from field observations and measurements in quarries with outcrops
232 more than 100 m wide, located within the NAF shear zone, of kilometer-width. Due
233 to different deformation markers, two distinct approaches for creeping (volcanic and
234 analogous rocks) and for locked (limestone) sections were used to evaluate the strain
235 values. The techniques are briefly described below including a discussion on the
236 uncertainty of the measurements. These techniques rely on the classical strain analysis
237 procedure well described, for example, in Ramsay (1967, 1980).

238

239 **3. Strain measurements methodology**

240

241 **3.1 Strain measurements along the creeping section (volcanic and analogous 242 rocks)**

243 The main parameters of the shear zone are defined in **Fig.2a**: the kinematic axes a
244 (direction of shear), ab (shear plane) and ac (displacement plane) and the
245 corresponding coordinate axes x , y , and z .

246

247 *Boudinage and folding analysis*

248 Strain values were extracted from XRF chemical maps relying on the difference of
249 composition of the boudins and the veins relative to the matrix. The quadratic

250 extension/contraction λ is defined as $\sqrt{\lambda} = l_f/l_i$ where l_f is the final length of a
 251 deformed object, and l_i is its initial length. The quadratic extension is calculated from
 252 boudinage or fractured structures: i) the initial length (l_i) is measured by adding the
 253 length of each boudin, and ii) the final length (l_f) is the length from the first to the last
 254 boudin (**Fig. 2b**). The quadratic contractional strain is estimated from folded veins
 255 with constant length during deformation by measuring i) the final length l_f of the vein
 256 and ii) its initial length l_i as the length of an arc along a deformed object using $l_i =$
 257 $\sum_{i=1}^n ds_i = \int_0^{l_0} dx \sqrt{1 + (dy/dx)^2}$, where ds_i is the length of the section between two
 258 successive points and can be written as $ds_i = \sqrt{(x_i - x_{i-1})^2 + (y_i - y_{i-1})^2}$ (**Fig.**
 259 **2b**).

260 Uncertainty is linked to the ratio between the resolution of the image and the
 261 measured lengths. For the boudinage, the total error is the sum of the errors on the
 262 measurements of each boudin element that gives the initial length plus the error on the
 263 final length. For the folding, it is the sum of the error on initial and final length. In the
 264 following, uncertainties are given with the results and are shown as error bars in the
 265 figures.

266

267 *Volume change due to mass transfer*

268 The relative mass change was estimated by comparing the composition of insoluble
 269 and soluble minerals between protected (as undeformed as possible) and exposed
 270 (deformed) zones at the micro- to meso-scales within a given shear zone (**Fig. 2c**).
 271 The relative mass change is $dm/m = I_p/I_e - 1$, where I_p and I_e are the
 272 concentration of all insoluble minerals in the protected and exposed zones,
 273 respectively (Gratier et al., 2013). If the rock density does not change, which is the
 274 case in the studied samples, the mass change is equivalent to the volume change. In
 275 plane strain, the product of the two principal strain values is related to the volumetric
 276 change as follows:

$$277 \sqrt{\lambda_1 \lambda_2} = 1 + \Delta, \quad (1)$$

278 where Δ is the change in volume (or surface area) (Ramsay, 1967). Because this
 279 measurement is based on mineral maps, the accuracy depends on the dimension and
 280 on the resolution of these maps. Then, the relative mass change depends on the
 281 amount of the insoluble minerals in the protected and exposed zones. An error,

282 defined as a percentage of the calculated volume change or strain, is given in the
283 following. In general, it is smaller than 2%.

284

285 *Grain geometry and orientation analysis*

286 Mineral phases were identified on the chemical maps and the following geometrical
287 properties of individual grains were measured: the minor and major axes of each grain
288 (r_{\min} , r_{\max}), the grain orientation φ_g defined clockwise relative to the North, and the
289 coordinates of the center of each grain (x_c , y_c) (**Fig. 2d**). In such cases, the standard
290 deviation of the data is calculated and used to estimate the uncertainty.

291

292 *Fry method*

293 This method is also known as center-to-center method (Fry, 1979, Genier and Eparad,
294 2007). The center of every strain marker (e.g. oxide grains) is plotted with respect to
295 the position of all other markers. A plot is built by locating one marker at the origin
296 and by plotting as dots the positions of all other markers. Then another marker is
297 located at the origin and the positions of all other markers are plotted, until all
298 markers have been considered. The result is a cloud of points that contains an empty
299 space (i.e. void) at the origin and which represents the strain ellipse. Strain is
300 quantified using i) the aspect ratio R_{Fry} between the elongation axis (a) and the
301 contraction axis (b) of the strain ellipse; and ii) the angle between the x-axis and the
302 elongation axis also defined as φ_{Fry} . This method is similar to an auto correlation
303 function, which can be used for the same type of object (Heilbronner, 2002;
304 Heilbronner and Barrett, 2014). The normalized center-to-center method (Erslev,
305 1988; Erslev and Ge, 1990) improves the Fry method by taking the grain size into
306 account and was used in the present study. The strain ellipse parameters (R_{Fry} , φ_{Fry})
307 were measured by superimposing all minerals together (**Fig. 2e**). The actual values of
308 the principal strain axes were then obtained using $R_{Fry} = \sqrt{\lambda_1}/\sqrt{\lambda_2}$ with $\sqrt{\lambda_1} =$
309 $\sqrt{R_{Fry}(1 + \Delta)}$ and $\sqrt{\lambda_2} = \sqrt{(1 + \Delta)/R_{Fry}}$. Uncertainty derives from fitting an ellipse
310 to the points along the rim of the void in order to extract (R_{Fry} , φ_{Fry}) values. For best
311 fitting higher weights are given to points that are close to the rim while none of the
312 clustered points remain out of the calculation as suggested by Mulchrone (2013). The
313 angle of the ellipse was fixed using the main orientation of the grains. We evaluate
314 the accuracy of this fitting procedure to be $\pm 5\%$.

315

316 *Cleavage angle evolution within the shear zone*

317 Evolution of the cleavage orientation within shear zone was used to evaluate the
318 displacement (Ramsay, 1980; Ramsay and Graham, 1970) when integrating the
319 volume change (Fossen, & Tikoff, 1993) at millimeter to decimeter scales on thin
320 sections, and up to macro-scale at meter to hectometer scales. In a shear plane,
321 passive line markers (e.g. cleavage or vein) originally making an angle α with the
322 shear zone walls (**Fig. 2f**) are sheared to make a new angle α' (Ramsay, 1980) such
323 that

$$324 \cot(\alpha') = \frac{\cot(\alpha) + \gamma}{1 + \Delta} \quad (2)$$

325 High γ values are very sensitive to α' angles field measurements: α' values of 1° , 0.5°
326 or 0.1° lead to γ values of 57, 114, or 572, respectively (**Eq. 2**). Consequently,
327 estimating γ from α' is extremely challenging for high strain values, due to the
328 difficulty of measuring α' with sufficient accuracy when it is smaller than 1° . In the
329 following, the uncertainty of this technique is given as the standard deviation of the
330 calculated strains.

331

332 **3.2 Strain measurements along the locked sections (limestone)**

333 The strain contraction component was calculated by two techniques. The first one
334 involves identifying veins shifted by stylolites (Gratier et al., 2013) and using $\Delta l =$
335 $l_{v2v} / \tan(\beta)$, where Δl is the thickness of the layer dissolved by the stylolite, l_{v2v} is
336 the distance between the vein walls, and β is the angle between the direction of the
337 vein and the stylolite peaks (**Fig. 2g**). The initial length of the vein is then obtained by
338 adding the shifted length to the final length of the shifted vein $l_i = l_f + \Delta l$, and the
339 quadratic contraction can be calculated. The second technique involves using the
340 maximum amplitude of the stylolite peaks (Toussaint et al., 2018) and
341 applying: $\sqrt{\lambda_{2_{stylolite}}(\varphi)} = \frac{L_{Tot}(\varphi)}{L_{Tot}(\varphi) + N_{Stylolite}(\varphi) \cdot \langle Amp_{max}(\varphi) \rangle}$, where $\sqrt{\lambda_{2_{stylolite}}}$ is the
342 quadratic contraction due to stylolite in their peak direction φ , $N_{Stylolite}$ is the number
343 of stylolites in direction φ , $\langle Amp_{max} \rangle$ is the average maximum peak height in
344 direction φ , and L_{Tot} is the total length of the sample in direction φ . The strain
345 extension component is measured perpendicular to the vein direction. The final length
346 is defined from side to side and the initial length is the spacing between the veins.

347 Finally, the contraction and extension components are combined and plotted in polar
 348 coordinates to define the strain ellipse. Uncertainty is linked to the ratio between
 349 resolution of the image and measured lengths. The total error is the sum of the errors
 350 on the measurements of shifted veins or stylolites size and on the final length and is
 351 given with the results.

352

353 **3.3 Relation between shear displacement and finite strain values**

354 For simple shear followed by volume change (**Fig. 2f**), the deformation matrix
 355 transforms the undeformed vector (x, z) (initial state) to a new position after
 356 deformation (deformed state) (x', z') : $\begin{pmatrix} x' \\ z' \end{pmatrix} = \begin{pmatrix} 1 & \gamma_{s,\Delta} \\ 0 & 1 + \Delta \end{pmatrix} \cdot \begin{pmatrix} x \\ z \end{pmatrix}$ where $\gamma_{s,\Delta}$ is the
 357 shear strain due to shear and dilation (Fossen and Tikoff, 1993), and the principal
 358 strains are:

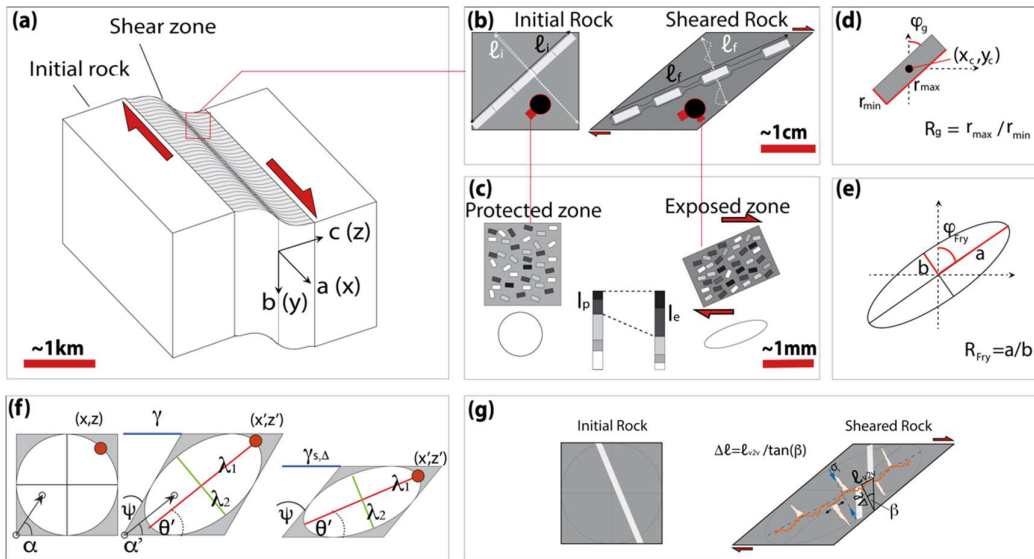
$$359 \lambda_{1,2} = \frac{1}{2} (1 + \gamma^2 + (1 + \Delta)^2 \pm \sqrt{[1 + \gamma^2 + (1 + \Delta)^2]^2 - 4(1 + \Delta)^2}) \quad (3)$$

360 where Δ is the volume change. Both the aspect ratio between the principal strain axes
 361 $R(\gamma, \Delta) = \sqrt{\lambda_1/\lambda_2}$, and $\theta'(\gamma, \Delta)$ are nonlinear functions of γ and Δ (i.e. surfaces in a
 362 three-dimensional coordinate system). Such functions are known as $R - \theta'$ diagrams
 363 when plotted together (Fossen and Tikoff, 1993). θ' is the angle between the
 364 quadratic extension λ_1 and shear direction. The total shear displacement $d_{\hat{a}}$ is
 365 obtained by integrating the shear strain along the z-axis (Ramsay, 1980), (**Fig. 2**)

$$366 d_{\hat{a}} = \int_0^w \gamma \cdot dz \quad (4)$$

367 where w is the width of the shear zone, and γ is the shear strain.

368



369

370 **Figure 2.** Strain measurements in shear zones. (a) Kinematic axes with a the direction of
 371 shear, ab the shear plane, and ac the displacement plane, and corresponding coordinates axes
 372 x,y,z . (b) Example of a boudinage strain marker at meso-scale. (c) Magnification of protected
 373 and exposed zones with their mineral composition and strain ellipse. (d) Definition of
 374 measured parameters of a single grain: major and minor axes r_{min} , r_{max} the aspect ratio R_g and
 375 the angle ϕ_g , and the location of the center of the grain (x_c, y_c) . (e) Definition of a strain
 376 ellipse using the Fry method with major and minor axes a , b and the aspect ratio between
 377 them $R_{Fry}=a/b$ with the angle ϕ_{Fry} . (f) Schematic deformation in simple shear coupled to
 378 volume change, with distortion of a unit circle into an ellipse, adapted from (Ramsay, 1967,
 379 1980). (g) Contraction evaluated from shifted veins along tectonic stylolite.

380

381 4. Results of finite strain measurements at selected sites

382 The structural strain markers differed from one outcrop to another although all were
 383 consistent with dextral slip on the North Anatolian Fault. Therefore, the strain values
 384 were extracted using different methods, depending on the particular type of strain
 385 markers along the shear zones. For five selected sites (Taşkesti, Hamamli, Ismetpasa,
 386 Yazıoren, Mülâyim, **Fig. 1**), strain measurements at thin section scale are shown in
 387 **Figures 3, 4, 5, 6, and 7**. For each outcrop, the mineral maps, mineral content
 388 histograms, Fry and rose diagrams have a similar color code. For the last selected site,
 389 Gerede (**Fig. 1**), strain measurements were integrated at the outcrop and regional
 390 scales (**Fig. 8**). Finally, all strain measurements on the six outcrops are integrated in
 391 **Figure 9**.

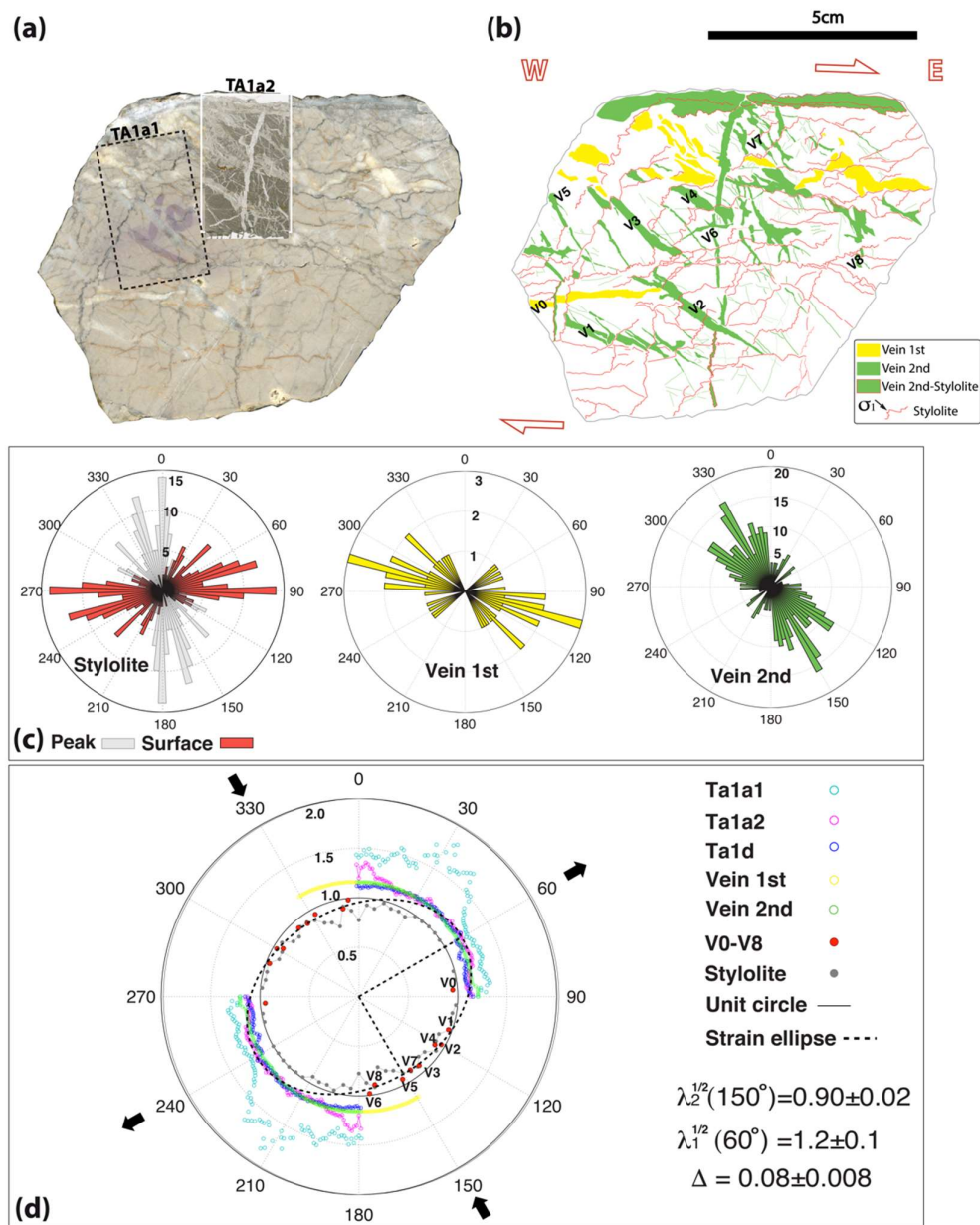
392

393 4.1 Taşkesti outcrop (locked section)

394 The Taşkesti outcrop (**Fig. 1**) exposes the fault zone in massive limestone formations,
 395 which are deformed by tectonic stylolites associated with veins as seen in a hand

396 sample (**Fig. 3**). This sample is well representative of the seven visited outcrops in
397 limestones, which all show several relatively narrow, meter-wide, shear zones with
398 numerous stylolites. The cumulated width of all these narrow shear zones is a few
399 meters at maximum for the whole NAF width. Six thin sections in three orthogonal
400 planes were cut from the Taskesti sample. Two types of tectonic stylolites can be
401 distinguished from the different color of their insoluble residues (either red or black)
402 (**Fig. 3a**). Several generations of veins can be seen that cross cut each other. Some are
403 sub-parallel to the shear fault plane; others are oblique to it. In places, some of the
404 veins are dissolved and shifted perpendicularly to the stylolite plane, indicating the
405 shortening direction. In other places, some of the horizontal veins are stylolitized,
406 while some vertical stylolites are filled with calcite (**Fig. 3b**). **Figure 3c** shows the
407 orientation of maximum amplitude peak directions and stylolite surfaces that are
408 oriented in various directions, with the highest density in the range N70 – N90. It also
409 shows the various orientations of the oblique-to-the fault veins, the most frequent one
410 in the direction N110, and other maximum directions range between N120 and N150
411 (**Fig. 3c**). This observation is well indicative of the successive generations of
412 stylolites and veins in non-coaxial deformation with alternatively seismic and
413 aseismic deformation.

414 The total strain was measured from stylolites and veins as follows. Firstly, the
415 extension was measured perpendicular to veins from both the hand sample and thin
416 sections (**Fig. 3a-b**). The strain values obtained from the thin sections were slightly
417 higher compared to the hand sample (**Fig. 3d**). Secondly, the contraction was
418 quantified by two independent measurements (see **section 3.2**) using: i) shifted veins
419 and ii) stylolite average amplitude. Both methods gave very similar results with a fit
420 of the strain ellipse with $\sqrt{\lambda_1} = 1.2 \pm 0.1$, $\sqrt{\lambda_2} = 0.9 \pm 0.02$. The main results from
421 these analyses were that, at the decimeter scale, the calculated mass change was $\Delta =$
422 0.08 ± 0.008 , and the shear strain values calculated from the stylolite values (as only
423 the stylolites are representative of the creep process) led to $\gamma = 0.20 \pm 0.05$ (**Eq. 3**).
424 It must be noted that, assuming that the shear fault corresponds to the shear direction,
425 the angle of the ellipse with the fault plane, which is about 30°, could indicate higher
426 shear value of about $\gamma = 1.15$. However, in such semi-brittle deformation, it is
427 difficult to evaluate the orientation of the shear direction, which is not always parallel
428 to the fault.



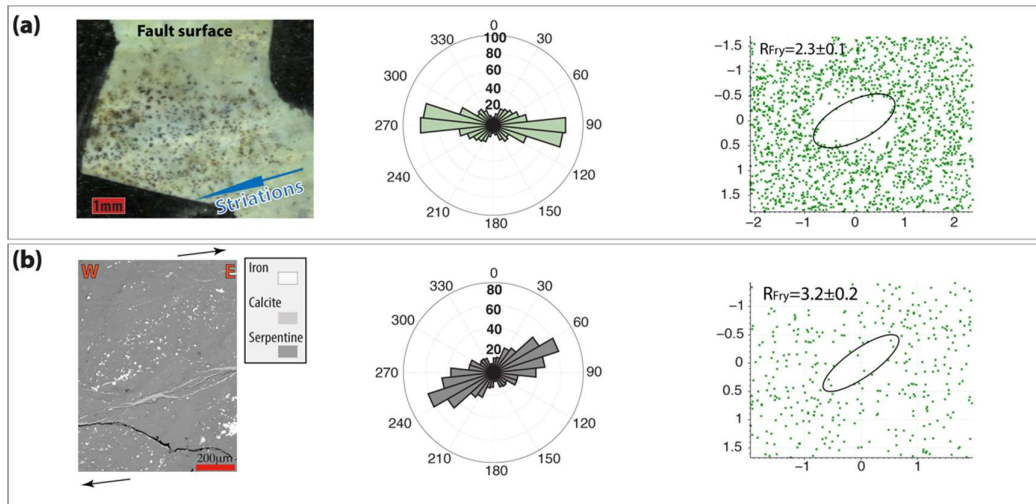
429

430 **Figure 3.** Taskesti outcrop. (a) Limestone hand sample in a horizontal displacement plane
 431 (ac, **Fig. 2a**) in the Taşkesti outcrop deformed by tectonic stylolites associated with two
 432 families of veins. Locations of the two thin sections Ta1a1 and Ta1a2 are indicated. A third
 433 thin-section, Ta1d, is not shown here. (b) Digitized tectonic stylolites and veins. (c) Rose
 434 diagrams representing the orientations of the tectonic stylolite surfaces (red) and peaks (gray),
 435 and the first and second generations of veins. (d) Strain ellipse based on stylolite and vein
 436 data giving contraction and extension values (black arrows), respectively. The locations of
 437 individual veins v1 to v8 are given in (b).

438

439 **4.2 Hamamli outcrop (creeping section)**

440 The Hamamli outcrop (**Fig. 1**) exposes serpentinite rocks deformed by dissolution
 441 along cleavage surfaces associated with precipitation in veins (Kaduri et al., 2017). In
 442 places, the strain was measured by the alignment of iron oxide minerals that
 443 agglomerate sub-parallel to the cleavages. The strain was measured along two planes:
 444 a shear plane on the fault (ab) and a displacement plane (ac) (**Fig. 2a**). The results are
 445 presented in **Fig. 4**. The strain values are $R_{Fry} = 2.3 \pm 0.1$ and $R_{Fry} = 3.2 \pm 0.2$ in
 446 the shear fault plane (ab) (**Fig. 4a**) and in the displacement plane (ac) (**Fig. 4b**),
 447 respectively. Rose diagrams of the iron oxides showed alignments in the strain
 448 elongation direction in the displacement plane (**Fig. 4b**) but were slightly oblique to
 449 the strain elongation in the shear plane (**Fig. 4a**). This result may be due to the effect
 450 of episodic oblique displacements along the fault surface. Because it was not possible
 451 to find an area that was less deformed than the studied samples, the mass change
 452 could not be calculated. A maximum shear strain value in the displacement plane $\gamma =$
 453 1.3 ± 0.1 was derived from **Eq. 3** (**Fig. 2f**).
 454

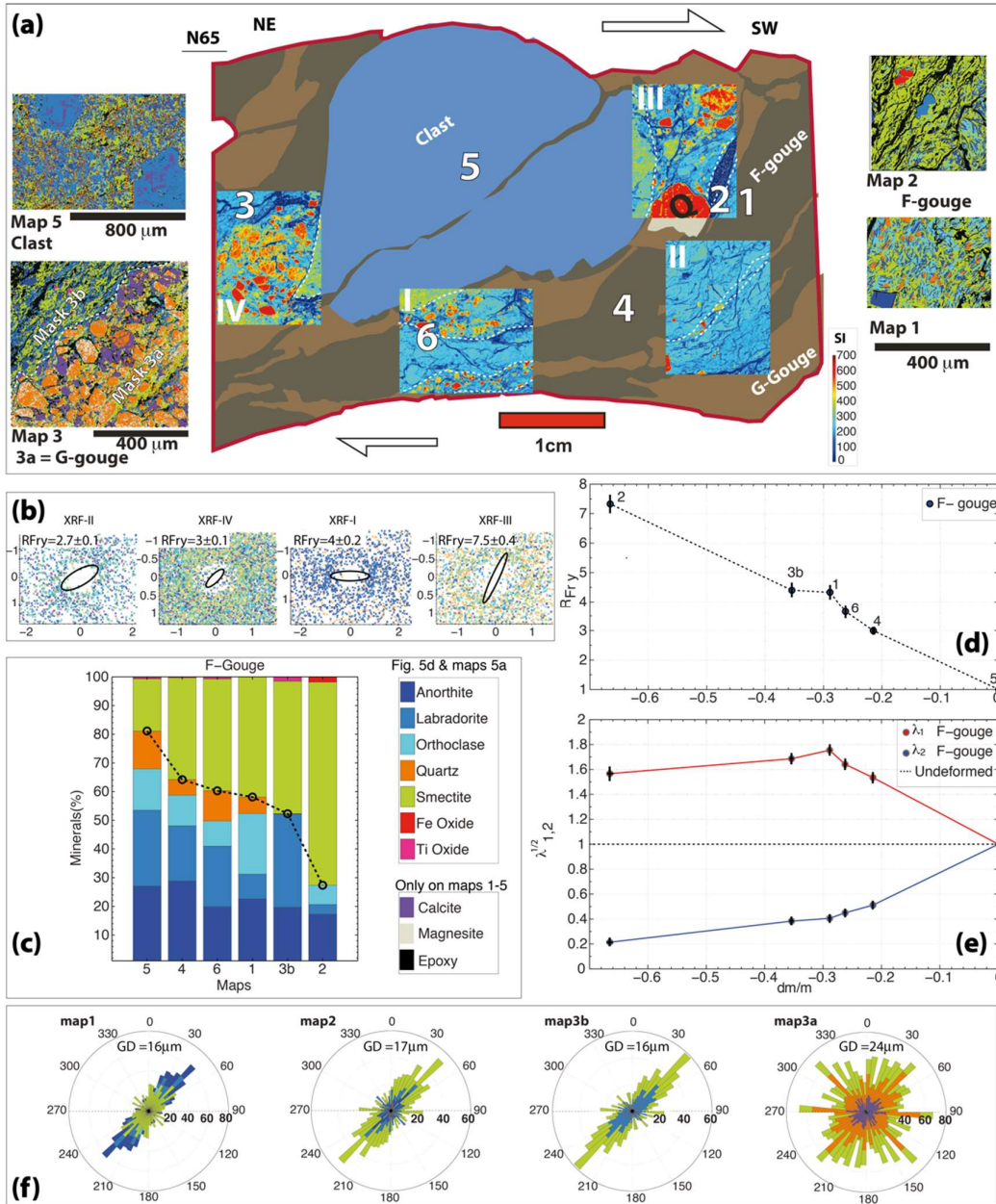


455
 456 **Figure 4.** Hamamli outcrop. Serpentinite rocks show cleavage-foliation and anisotropic
 457 clustering of iron oxide minerals in the cleavage. These microstructures were used to evaluate
 458 the strain values in two directions. (a) Strain analysis of a fault surface in vertical shear plane
 459 (ab), with the rose diagram of Fe oxides, and plots of the normalized Fry using Fe oxides. (b)
 460 Strain analysis with the same two methods in a horizontal displacement plane (ac). The shear
 461 arrows are indicative of the sense of shear and do not show an accurate orientation of the
 462 shear direction, which is not available on the outcrop where the sample was collected.
 463

464 **4.3 Ismetpasa outcrop (creeping section)**

465 The Ismetpasa outcrop (**Fig. 1**) exposes the fault within volcanic rocks near their
466 boundary with sandstone-shale units. At this site, the deformation in the gouge is
467 associated with tectonic layering leading to more or less fine-grained foliated to
468 granular rocks (F-gouge to G-gouge), **Fig. 5a**. The foliations are deformed around a
469 rigid clast of residual domain of the initial (undeformed) rock. **Figure 5a** shows a
470 digitized horizontal thin section, parallel to the displacement plane (ac) of the shear
471 zone, with four XRF maps of silicon on which the Fry-method analysis gave four
472 R_{Fry} values ranging from 2.7 ± 0.1 to 7.5 ± 0.4 at centimeter scale (**Fig. 5b**). It was
473 also possible to calculate the strain distribution at millimeter scale within this thin
474 section from six EPMA maps of mineral content acquired in exposed (the most
475 deformed areas) and protected zones (the less deformed area which is the clast of
476 undeformed initial rock), **Fig. 5a**. Some of the maps were masked in order to obtain a
477 more accurate evaluation of localized strain and mineral composition in zones of
478 disaggregated but relatively undeformed large quartzite clasts (map 3, **Fig. 5a**). The
479 relative mass change was calculated excluding magnesite and calcite from the
480 calculation considering that these minerals sealed the porosity of the rock more
481 recently in the deformation process and at least after the process of massive
482 deformation considered here (Kaduri et al., 2017). There was a clear decrease in
483 soluble mineral content from the protected (undeformed) zone to the exposed zones
484 (**Fig. 5c**). Then, in parallel, the strain was measured on these EPMA maps using the
485 Fry method and R_{Fry} values for few representative minerals (albite, anorthite,
486 orthoclase, quartz, smectite). These values were plotted as a function of the relative
487 mass change and showed a near-linear trend (**Fig. 5d**). The values of R_{Fry} converted
488 to quadratic contraction/extension $\sqrt{\lambda_{1,2}}$ using the Δ values (**Eq. 1**), showed the actual
489 successive change in deformation as a function of the relative mass change (**Fig. 5e**).
490 Maximum shear strain values are given by maps 2 and 1 with $\gamma = 1.2 \pm$
491 0.1 and 1.3 ± 0.1 with $R_{Fry} = 7.3$ and 4.5 and with $\Delta = -0.66$ and -0.3 , respectively.
492 The orientation of the grains in the exposed zones was most often aligned with the
493 local foliation. On the contrary, the protected (near undeformed) zones showed
494 random grain distribution with negligible preferred orientation (**Fig. 5f**). It must be
495 noted that the estimated grain orientations and the Fry plots are not parallel, because
496 the calculations are not performed at the same scale and because the deformation is
497 heterogeneous around the clast. The Fry plots are obtained from large XRD maps that

498 represent a sampling of the whole thin section (area XRF I, II, III, IV) whereas the
 499 grain orientation are obtained from local EPMA chemical maps and are located in
 500 zones where the orientation of the deformation is pretty the same (area map 1, 2, 3b).
 501



502 **Figure 5.** Ismetpasa outcrop. (a) Thin section in horizontal displacement plane (ac) with two
 503 types of gouge layers: foliated and granular gouges (F-gouge and G-gouge), around an
 504 undeformed clast of initial rock, with four XRF maps of silicon (I-VI, centimeter-scale) and
 505 six detailed EPMA maps of minerals located by numbers (1-6, millimeter scale), each
 506 corresponds to a different mineral, given in the legend in (c). (b) Fry method based on the
 507 four XRF large maps (I-IV) with their location above and using the minerals in the legend in
 508 (c). (c) Mineral content of the mineral maps. (d) Strain measurements using the Fry method
 509

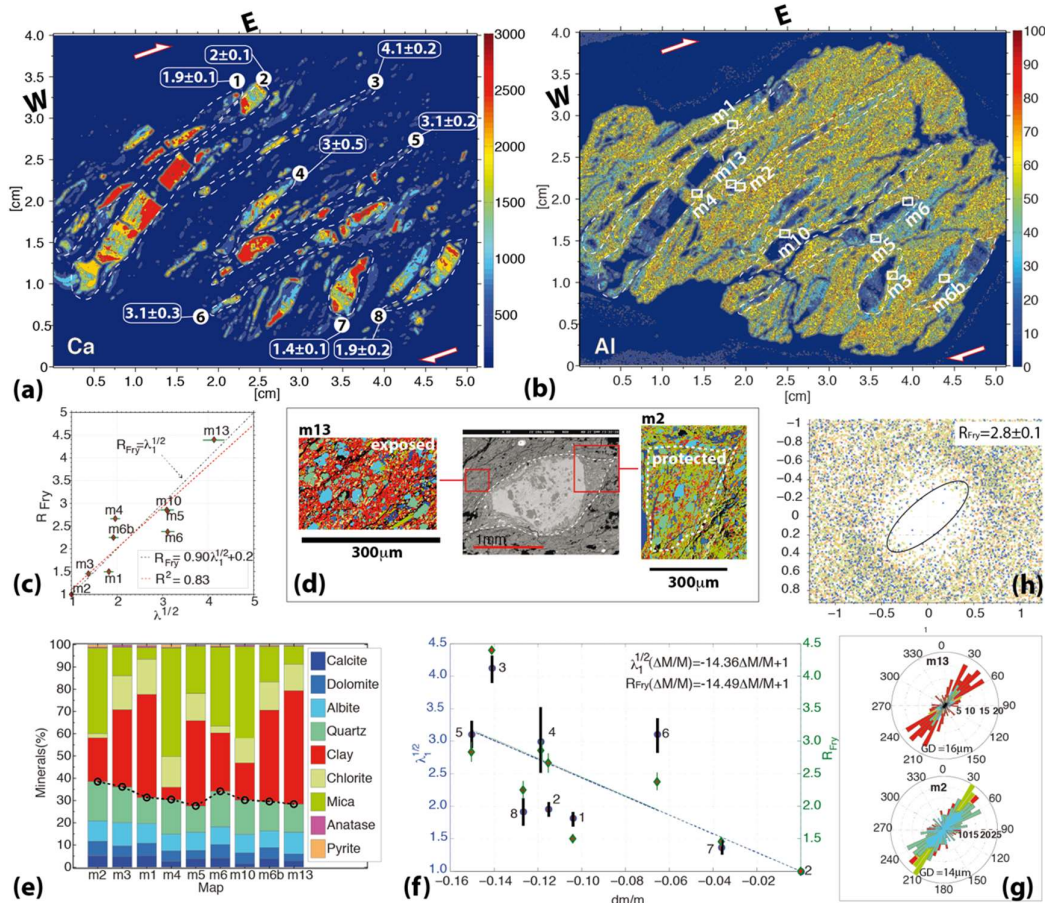
510 and R_{Fry} on these detailed maps plotted as a function of the relative mass change (uncertainty
511 is lower than the symbol size). The reference (protected zones) is map 5 (clast). (e) Plot of the
512 calculated quadratic contraction/extension as a function of the relative mass change. (f) Rose
513 diagrams of grain orientations from mineral maps with mean grain diameters (GD). The color
514 code is the same as in (c). The shear arrows are indicative of the sense of shear and do not
515 show an accurate orientation of the shear direction, which is not available on the outcrop
516 where the sample was collected.

517

518 **4.4 Yazioeren outcrop (creeping section)**

519 The Yazioeren outcrop (**Fig. 1**) exposes a shear zone in a mélange of carbonate and
520 volcanic rocks with tectonic layering (rich in alumino-silicates) and carbonate
521 boudinage. A horizontal thin section parallel to the displacement plane (ac) of the
522 shear zone was imaged by XRF in order to segment and measure strain markers.
523 **Figure 6a** shows the calcium XRF map, which emphasizes carbonate boudinage. The
524 strain measurement included two steps. The first step was to measure the quadratic
525 extension $\sqrt{\lambda_1}$ along the preserved boudinage (strain markers 1 to 8 in **Fig. 6a**), which
526 varied in the range $1.4 \pm 0.1 - 4.1 \pm 0.2$. The second step was to measure the
527 deformation recorded in the deformed matrix located near the boudins, using the Fry
528 method on maps of elements acquired from EPMA measurements (location m1-m13,
529 in the Al map **Fig. 6b**). The R_{Fry} values varied from 1.0 ± 0.05 to 4.5 ± 0.2 . The
530 plot of these two types of measurement showed a good correlation with a fitted linear
531 regression coefficient $R^2=0.89$ (**Fig. 6c**). The relative mass change was calculated
532 using the EPMA maps by comparing exposed and protected zones. **Figure 6d** shows
533 an example of these two types of zones: an exposed zone (map13) and a protected
534 zone (pressure shadow contoured with a dashed line in map2) around a calcite clast.
535 From the protected to the more exposed zones, the mineral assemblage showed a
536 reduction in soluble mineral content and a relative increase in insoluble mineral
537 content (**Fig. 6e**). It is thus possible to calculate the mass change relative to the
538 protected zone (map2) (**Fig. 6f**). There is a near-linear trend between these strain
539 values evaluated by two different methods, $\sqrt{\lambda_1}$ and R_{Fry} , and the mass change (**Fig.**
540 **6f**). Note that such mass change values were only minimum values since the protected
541 zone was not an initial state but rather a less deformed state as attested by the
542 preferred orientation of the grains (**Fig. 6g**). Based on these measurements, boudinage
543 values associated with mean $\Delta = -0.1$ lead to strain ratios ranging from 2 to 19 (**Eq.**
544 **1**) and gave a shear strain in the range $\gamma = 1.3 \pm 0.1$ to 4.0 ± 0.2 while the Fry

545 method gave a range $\gamma = 0.5 \pm 0.05$ to 1.5 ± 0.1 (Eq. 3). These values may be
 546 compared with the value of $\gamma = 1.3 \pm 0.1$ obtained by the Fry method at the
 547 decimeter scale with $R_{Fry} = 2.8 \pm 0.1$ (Fig. 6h) and considering a mean value $\Delta =$
 548 0.1. The variation of γ values at millimetric scale indicated the development of
 549 parallel-to-the-cleavage differentiated layers with various associated shear values.
 550



551
 552 **Figure 6.** Yazioren outcrop. (a) Calcium map measured by XRF showing boudinage with
 553 variation in extensional strain in the horizontal displacement plane (ac). (b) Aluminum map
 554 measured by XRF showing the foliation with the locations of the EPMA mineral maps. In (a)
 555 and (b) color bars indicate relative content. (c) Correlation between $\sqrt{\lambda_1}$ extension of the
 556 boudinage and R_{Fry} of the nearby matrix (with map-numbers). (d) SEM-BSE image and
 557 EPMA mineral maps, showing protected (right) and exposed (left) zones. The color-coding
 558 for the minerals is given in the legend of (e). (e) Comparative mineral compositions, the
 559 dashed line separates soluble (mobile) and insoluble (non-mobile) minerals. (f) Extension
 560 values $\sqrt{\lambda_1}$ (from boudinage with numbers) and R_{Fry} (from EPMA mineral maps) versus the
 561 relative mass change (uncertainty is lower than the symbol size). (g) Rose diagrams of grain
 562 orientations from EPMA mineral maps with mean grain diameters (GD) and orientation with
 563 respect to the North. (h) Fry diagram at centimeter scale on the whole thin section from XRF
 564 data using the minerals in the legend in (e). The shear arrows are indicative of the sense of

565 shear and do not show an accurate orientation of the shear direction, which is not available on
566 the outcrop where the sample was collected.

567

568 **4.5 Mülayim outcrop (creeping section)**

569 In this outcrop (see **Fig. 1**), the fault crosscuts a block of ophiolitic mélange
570 embedded into schist, marble and metabasite units. Measurements were performed in
571 the schist that constitutes the main fraction of the rock body. Deformation was
572 measured based on the Fry method at different scales with also the use of folded
573 carbonate veins and the statistical analysis of grain clusters. A horizontal thin section
574 parallel to the displacement plane (ac) revealed a shear zone with cleavage - foliation
575 sub-parallel to the fault and crosscut by carbonate veins in all directions (**Fig. 7a**).

576 The cleavage can be seen on the XRF aluminum map (**Fig. 7b**). Parallel-to-the
577 cleavage and oblique-to-the-cleavage veins registered only a small part of the finite
578 extension $\sqrt{\lambda_1}(0^\circ) = 1.1 \pm 0.02$, and $\sqrt{\lambda_1}(90^\circ) = 1.2 \pm 0.02$, respectively.

579 Contraction perpendicular to the foliation was measured from folded veins giving

580 $\sqrt{\lambda_2} = 0.45 \pm 0.01$ from the decimeter-scale thin section (**Fig. 7a**) to millimeter size

581 EPMA map (**Fig. 7d**) indicating a minimum shear strain $\gamma = 1.8 \pm 0.1$. The
582 deformation evaluated from the Fry method using XRF aluminum maps had a
583 relatively high value $R_{Fry} = 8.5 \pm 0.4$ (**Fig. 7c**), corresponding to a shear strain $\gamma =$

584 2.6 ± 0.2 . No volume change can be calculated here due to the lack of clear evidence

585 of undeformed (or at least less deformed) protected zones, despite cleavage-foliation

586 being visible with alternating phyllosilicate-rich and albite-quartz-rich tectonic layers

587 indicating significant pressure solution mass transfer (**Fig. 7f**). Nonetheless, the

588 deformation was highly heterogeneous. Using EPMA maps at millimeter size, some

589 strain variations can be measured when using the Fry method in three different zones:

590 the hinge (H) and the limb (L) of an arcuate structure and an intermediate area (I)

591 (**Fig. 7d**) with R_{Fry} ranging from 1.4 ± 0.07 to 4.8 ± 0.2 (**Fig. 7e**). Such strain values,

592 lower at millimeter size than at decimeter scale, implies the existence of some highly

593 localized zones with very high strain values at the sub-grain size. Within these zones,

594 phyllosilicates were oriented sub-parallel to the shear zone with embedded residues of

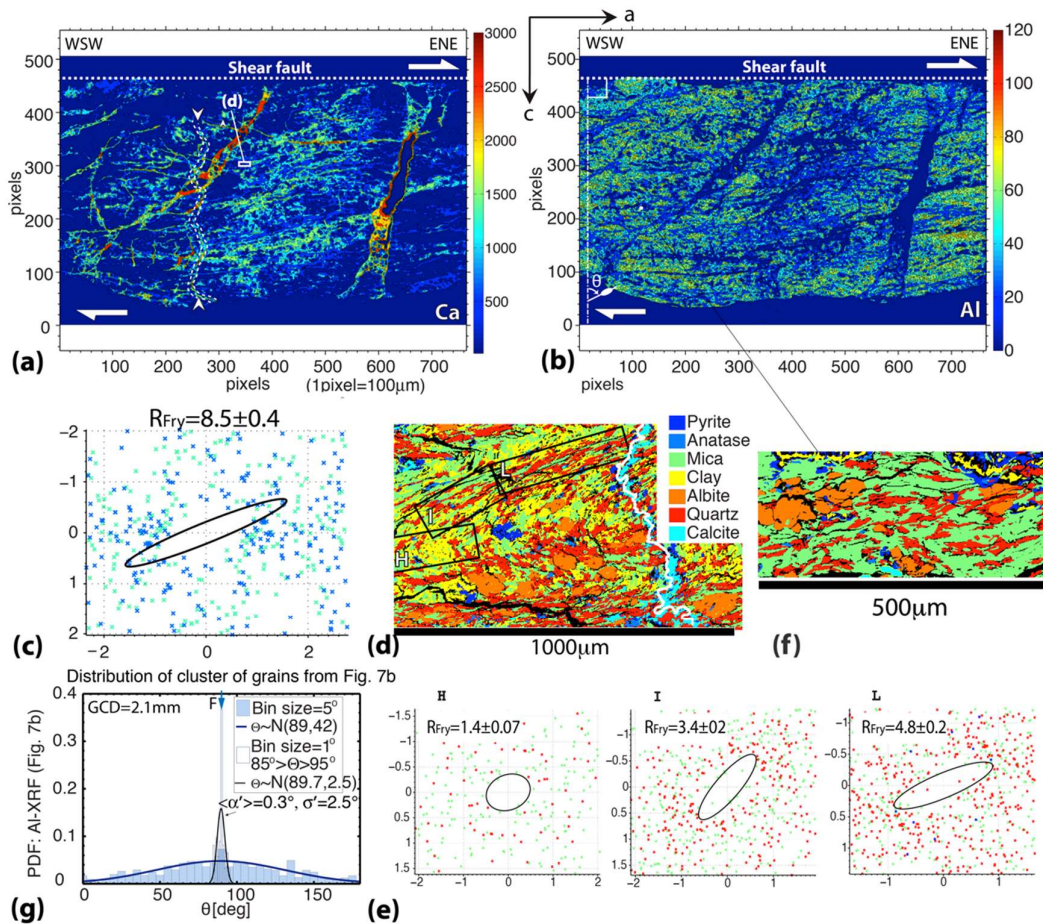
595 feldspar minerals and Fe oxides (**Fig. 7f**) and such fine-grained thin structures were

596 probably not taken into account well enough by the Fry method.

597 This problem was investigated by a statistical analysis of the Al grain clusters. More

598 precisely, relying on the detailed observations of the grain orientation and associated

599 Fry ellipses (**Fig. 7d, f, e**), the distribution of the orientation of elongated grain
600 clusters containing aluminum was fitted with two normal probability density
601 functions. At the decimeter scale (**Fig. 7b**) the cleavage-foliation is overall sub-
602 parallel to the fault at the boundary of the sample. In such ductile context, this fault is
603 presumably parallel to the shear direction and, by measuring the angle θ between the
604 line normal to the fault and the long axis of elongated grain clusters, the amplitude of
605 shear strain is evaluated: the angle between the long axis of elongated grains
606 (assimilated to cleavage plane) and the shear direction (α) is α' when $\alpha = 45^\circ$ (initial
607 cleavage) in **Eq. 2**. This calculation corresponds to the wider normal distribution in
608 **Fig. 7g** that gives an angle $\alpha' = 1^\circ$ associated with γ value of 57. However, at the
609 millimeter scale (**Fig. 7d, f**) we described two types of shear bands: those with grains
610 oblique to the shear direction and those with grains sub-parallel to this shear direction.
611 In order to better constrain the value of the α' angle in the shear bands with grains
612 sub-parallel to the shear direction we choose to concentrate our analysis in these
613 bands (with angles ranging from -5 to $+5^\circ$ with the shear direction). Doing so we
614 avoid the possible perturbing effect of the grains oblique to the shear. This calculation
615 corresponds to the narrower normal distribution in **Fig. 7g** that gives an angle $\alpha' =$
616 0.3° . This value of the α' angle is associated with γ value of 190. However,
617 uncertainty on the angle between the normal distribution of the elongated grains
618 cluster and the shear displacement is probably higher than 1° , leading to γ values
619 ranging from 40 to > 200 . The coexistence of such a wide range of strain values, the
620 uncertainty of their evaluation and their use to calculate the aseismic displacement is
621 discussed in section **5.2**.
622



623

624 **Figure 7.** Mülâyim outcrop. (a) Calcium map in the horizontal displacement plane (ac)
 625 measured by XRF with a folded calcite vein (ptygmatic vein) shown between arrows along
 626 dashed white lines. (b) Aluminum map in the same sample indicates sub-parallel cleavage-
 627 foliation rich in phyllosilicates. In (a) and (b), color bars indicate relative content. (c) Fry
 628 diagram based on XRF-maps (decimeter scale, 7a-b) using the minerals in the legend in (d):
 629 anatase and mica. (d) Mineral maps with fold separated by masks in the hinge, the
 630 intermediate area and the limb with different cleavage angles (locations are given in **Fig. 7a**).
 631 (e) Fry diagram based on the selected zones of the mineral map with increasing values of R_{Fry}
 632 from hinge to limb (millimeter scale) using the minerals in the legend in (d). (f) Mineral map
 633 in a highly localized deformation zone with sub-parallel to the shear displacement micas and
 634 embedded fine-grained quartz. (g) Orientation distribution of clusters of grains containing Al
 635 (**Fig. 7b**) fitted by two normal probability density functions (PDF), with the mean diameter of
 636 the grain clusters containing aluminum (GCD). Parameters of the two normal distributions are
 637 $\langle\alpha'\rangle$ the mean angle with the fault (F) and σ' the standard deviation: $\langle\alpha'\rangle = 1^\circ$ and 0.3° and
 638 $\sigma' = 42^\circ$ and 2.5° for the wide and narrow distributions, respectively.

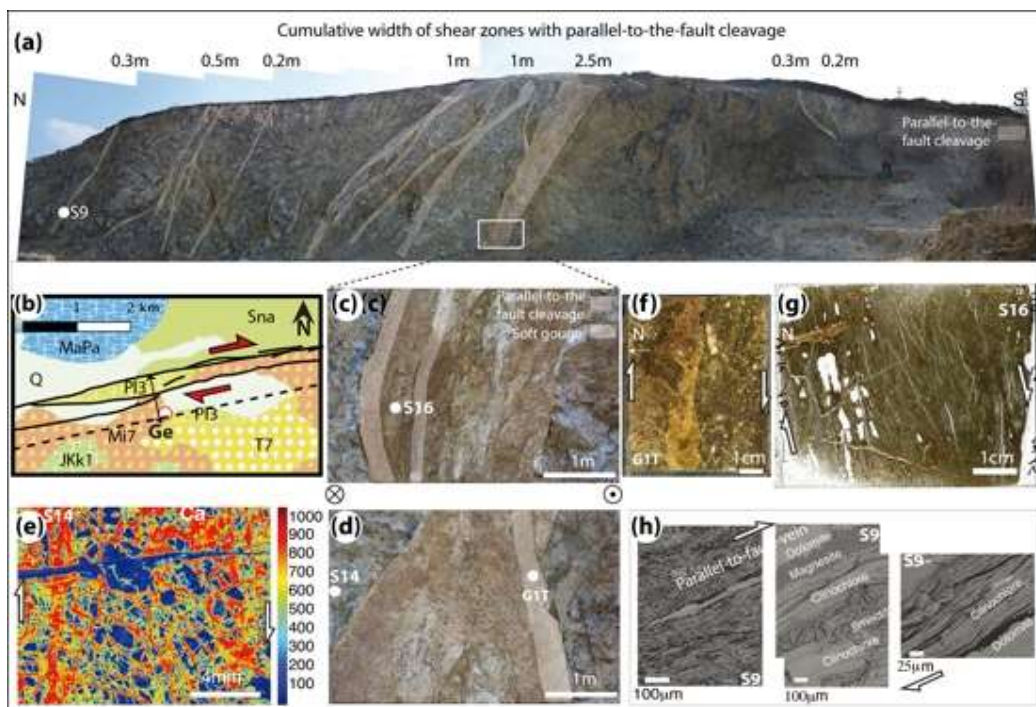
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640 4.6 Gerede outcrop (creeping section)

641 The Gerede outcrop (**Figs. 1** and **8b**) is an abandoned quarry at the western end of the
 642 Ismetpasa creeping segment (Hussain et al., 2016; Cetin et al., 2014). It is located one
 643 kilometer south of the present-day active fault zone with several other major strike-

644 slip faults in between mostly buried under Quaternary sediments (Kaduri et al., 2017).
645 The quarry presents a continuous 150 m-wide outcrop (**Fig. 8a**) that reveals a network
646 of 5 to 30 cm thick clay-rich soft gouges with parallel-to-the-fault cleavage-foliation
647 (**Fig. 8a, 8c, 8d**). These faults cross the volcanic units of the Galatia massif
648 (Adiyaman et al., 2001), which are a mélange of dacite, andesite and trachy-basalts
649 (Wilson et al., 1997). These different units are difficult to distinguish. In the gouges,
650 they have been transformed by deformation coupled to low temperature
651 metamorphism (Kaduri et al., 2017). In the damage zones, they have been highly
652 damaged with fractured blocks and dense networks of vertical veins in all directions
653 (**Fig. 8e**). Such sealed fractures were related to episodic inflow events of carbonate-
654 rich fluids that were associated with the successive earthquakes that affected this
655 section of the NAF (Kaduri et al., 2017). Moreover, the deformation was
656 heterogeneous with more or less rigid fractured blocks surrounded by foliated zones
657 with intense deformation displaying a braided fault pattern at all scales, from thin
658 section (**Fig. 5a**) to regional scale (**Fig. 8a**). The size of such rigid blocks decreased
659 drastically in gouges and in zones with cleavage sub-parallel to the shear fault but the
660 boundaries of such zones were undulated and their widths varied along strike (**Fig.**
661 **8a-c-d**). In thin sections, only some islands of volcanic rocks can be seen between the
662 networks of veins. Consequently, it was not possible to evaluate the strain and shear
663 values using particle distribution as in the other studied outcrops.
664 However, it was possible to evaluate the shear values from the angles between the
665 cleavage and the shear fault planes, similar to what was done at microstructural scale
666 for the Mülayim outcrop (**Fig. 7g**). Such angles can be measured either directly in the
667 field or in thin sections. It was not possible to measure the variations of these angles
668 along a continuous path over the entire 150 m of the outcrop. However, the damage
669 zones appeared to be more layered in areas near the gouges than away from them
670 (**Fig. 8c-d**). Thin sections showed that, in these areas near the gouges, the cleavage
671 orientation was sub-parallel to the shear fault plane (**Fig. 8f-g-h**), whereas away from
672 the gouge the cleavage was oblique to the fault (**Figs. 8e**), as in other outcrops (**Fig.**
673 **5f, 6g**). Such change of cleavage orientation being rather sharp, it was possible to
674 evaluate in the field the width of the shear zones characterized by parallel-to-the-fault
675 cleavage with an uncertainty of about 10%. Such shear zones include all the gouges
676 plus damage zones around them that have the same structural aspect as that seen on
677 thin sections (**Fig. 8f-g**). These shear zones are highlighted in near-transparent white

678 in Fig. 8a-c-d and a cumulative width of 6.0 ± 0.6 m was found along the 150 m
 679 width of the quarry. The ratio between this cumulative width of shear zones with
 680 parallel-to-the-fault cleavage and the total 150 m width of the outcrop may be
 681 evaluated and is equal to 0.04 ± 0.004 . This ratio can be considered as a representative
 682 value at a more regional scale (100-1000 m). The shear strain values associated with
 683 cleavage sub-parallel to the fault were not easy to evaluate and will be discussed in
 684 section 5.2. These strain values and the width of the shear zones with fault parallel
 685 cleavage were used to estimate the total fault displacement accommodated by creep
 686 during geological times in section 5.3.
 687



688 **Figure 8.** Gerede outcrop. (a) Photograph of the 150 m-wide outcrop showing the fault
 689 network and the location and widths of zones (shaded in white) with parallel-to-the-fault
 690 cleavage/foliation. (b) Geological map with the trace of the main fault in the outcrop (dashed
 691 line), Q=Quaternary, P13=Pleistocene conglomerate sandstone, M7=Miocene agglomerate
 692 tuff, T7=Eocene andesite basalt, MaPa=Cretaceous limestone, Sna=Cretaceous limestone,
 693 JKk1=Jurassic limestone. (c) Zone with parallel-to-the-fault cleavage and soft gouge in the
 694 upper part of the quarry: see location in (a). (d) Zone with parallel-to-the-fault cleavage and
 695 soft gouge in the lower part of the quarry (not seen in a). (e) Element map of Ca using XRF
 696 with typical network of calcite - dolomite veins parallel and perpendicular to the cleavage.
 697 The blue perpendicular-to-the-fault vein is a late magnesite vein. (f) Parallel-to-the-fault
 698 cleavage in a thin section of soft gouge. (h) Parallel-to-the-fault cleavage in a thin section
 699 near the gouge (see location in c). (g) Parallel-to-the-fault cleavage in SEM image with
 700 dolomite and magnesite veins parallel to the fault and to the cleavage - foliation (layers of
 701 clinocllore and smectite).
 702
 703

704 **5. Discussion**

705 Several parameters are needed to calculate the displacement associated with creep
706 processes in the NAF shear zone: the quadratic extension and contraction strain
707 values, the volume change, that allow calculating the shear strain when data of all
708 outcrops are combined (**Fig. 9**), and the width of the shear zone that allows
709 calculating the shear displacement (**Eq. 4**). The evaluation of these parameters is
710 discussed below, as well as their uncertainty. In section **5.1** we discuss the volume
711 change in the creeping sections and its effect on shear displacement. In section **5.2** we
712 discuss the strain and width evaluation. We evaluate the total displacement in locked
713 sections. We discuss the effect of heterogeneous deformation in creeping sections and
714 we propose a model of two normal shear strain distributions. Finally, the modeling of
715 the total aseismic displacement along the creeping sections is discussed in section **5.3**.

716

717 **5.1 Volume change evaluation: uncertainty and effect on shear displacement**

718 Here we discuss the relationship between strain and mass-volume change because
719 volume change is an important parameter of the relation between total strain and shear
720 strain (**Eq. 3**). These relations are discussed based on a “strain-ratio/shear-
721 strain/volume-change” diagram (**Fig. 9**). An horizontal plane strain in 2D is assumed,
722 thereby implying homogeneity along the y -axis (Ramsay, 1980; Heilbronner and
723 Barrett, 2014; Fossen and Cavalcante, 2017). For the same measured strain ratio, R ,
724 the higher is the volume change the lower is the associated shear strain (**Fig. 9**). For
725 example, for $R=10$, the shear strain ranges from $\gamma = 2.9$ to $\gamma = 2.0$ for a volume
726 change ranging from $\Delta = 0$ to $\Delta = -0.5$, respectively (**Fig. 9**). For higher strain ratio,
727 such as $R=10000$, when cleavage become sub-parallel to the shear zone at less than
728 0.5° angle, the shear strain ranges from $\gamma = 100$ to $\gamma = 72$ for a volume change
729 ranging from $\Delta = 0$ to $\Delta = -0.5$, respectively (**Fig. 9**). Despite such variations of
730 strain for such large variation from $\Delta = 0$ to $\Delta = -0.5$, it does not change the order of
731 magnitude of the shear strain values. It is worth noting that the calculated mass
732 change is a relative value. It is obtained by comparing a deformed “exposed” area
733 with a “protected” area which is either undeformed (in the best case) or which is just
734 less deformed. This is the reason why we also evaluated the deformation of the
735 protected zone from grain geometry and orientation analysis. Moreover, as the rock
736 density does not significantly change, mass change is equivalent to volume change.

737 As previously discussed in section 2.2, the six samples presented here are
738 representative of all visited sites along the NAF shear zone (**Fig. 1**): Taskesti in
739 limestone, Hamamli in serpentine and the four other sites (Ismetpasa, Yazioreni,
740 Mulayim, Gerede) in various rocks with volcanic origin.

741 All the sites in limestone show several narrow parallel-to-the-fault shear zones of
742 decimeter width with numerous stylolites. The Taskesti sampling is well
743 representative of all these shear zones for which the deformation near the fault
744 measured at decimeter scale shows no significant volume change with $\Delta = 0.08$ (**Fig.**
745 **3**). Observations (**Fig. 3**) always show clear evidence of alternating seismic (fractures)
746 and aseismic (stylolites) processes (Kaduri et al., 2017). However, the aseismic part
747 remains very low, with mean quadratic values of $\sqrt{\lambda_1} = 1.2 \pm 0.1$ and $\sqrt{\lambda_2} = 0.9 \pm$
748 0.02 leading to a low shear strain value $\gamma = 0.2 \pm 0.05$, a value not significantly
749 affected by uncertainties on volume change.

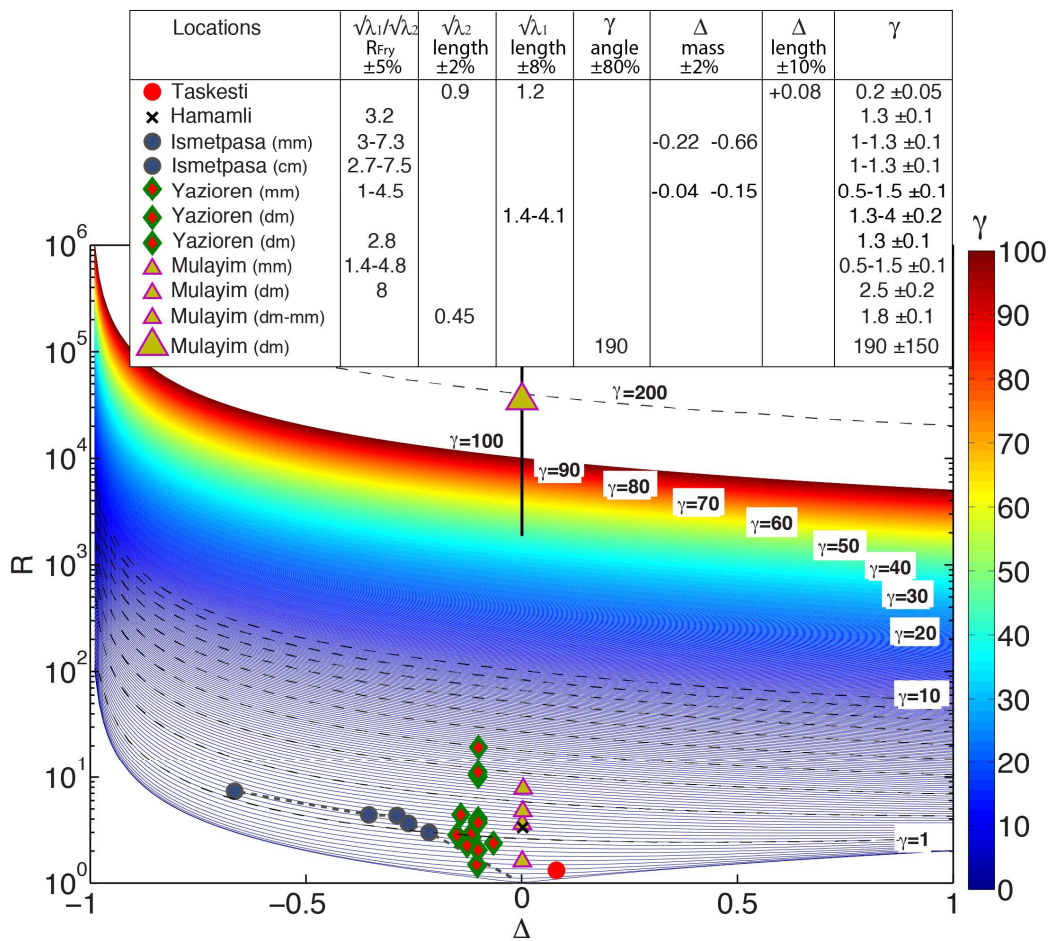
750 At the Hamamli serpentine site, it was not possible to calculate a volume change as no
751 area less deformed than any other was found. This difficulty is a general problem of
752 volume change calculation in serpentine where heterogeneities are rare (Andreani et
753 al., 2005). In Mulayim, the same absence of protected zone prevented us from
754 evaluating the volume change. The corresponding strain values at these sites are then
755 plotted in **Fig. 9** assuming no volume change.

756 At the other sites, volume change was measured by comparative chemical analyses
757 (**Fig. 5 & 6**) at millimeter scale. These calculations assume that in the deformed areas
758 the soluble minerals dissolve and are removed. A minor part of these soluble
759 minerals, as some feldspars, were transformed into newly crystallized phyllosilicate
760 minerals rather than removed, however, this does not significantly alter the
761 calculations (Kaduri et al. 2017). There is a clear correlation between strain and
762 volume change indicating that the same pressure solution mechanism is responsible
763 for the deformation with the development of parallel-to-the-cleavage tectonic
764 layering. When the protected zone is in an undeformed state, as in Ismetpasa (**Fig. 5**),
765 the volume change varies from one layer to another from 0.22 ± 0.005 to more than
766 0.66 ± 0.01 . When the protected zone is slightly deformed, as in Yazioreni (**Fig. 6**),
767 the variations are lower and the uncertainty is much higher. This pattern of successive
768 perpendicular-to-the-compressive-stress layers with various compositions is typical of
769 a pressure solution self-organized process that has been reproduced experimentally

770 (Gratier et al., 2015). This observation was made in the 25 investigated outcrops in
771 volcanic and analogous rocks of the NAF creeping shear zone. It indicates a very slow
772 ductile and aseismic deformation (Gratier et al., 2013). In order to extrapolate this
773 quantitative result to the whole shear zone, some complementary geological
774 observations are needed. No traces of recrystallization of the soluble minerals (quartz
775 and feldspar) are found in the shear zone, so it is likely that the whole shear zone was
776 developed with a decrease of volume at least during the early time of its development.
777 However, later in the NAF development, numerous carbonate veins crosscut some
778 damage zone as seen in Gerede (**Fig. 8e**). As such carbonates need to come from
779 outside since they were not present in the initial rocks, they may have contributed to a
780 local increase of volume. Consequently, in order to evaluate shear strain values
781 representative of the entire NAF shear zone in volcanic and analogous rocks (**Fig. 9**)
782 we used a median regional value of the volume change with a large uncertainty, $\Delta =$
783 -0.20 ± 0.15 .

784 Finally, the possibility of much higher volume changes, that we may have missed,
785 must be discussed because large volume change values below $\Delta = -0.8$ have a large
786 effect on the calculation of shear strain (**Fig. 9**). Common observation is that volume
787 change associated with pressure solution is limited to about $\Delta = -0.7$. It is due to
788 several effects especially the progressive decrease of the soluble mineral content and
789 their progressive isolation in the soft matrix of insoluble species that render their
790 stress-driven dissolution more and more difficult (Gratier, et al., 2013). Moreover,
791 there is another effect in shear zones that limits the volume change, which is linked to
792 the rotation of the cleavage. Pressure solution initiates along solution cleavage planes
793 near rigid objects (**Fig. 5a**) and develops progressively further away at 45° from the
794 shearing direction at the beginning of the shearing process. At this stage, the shearing
795 displacement contributes to the tectonic layering process with volume decrease.
796 However, at a later stage with the rotation of the cleavage that becomes sub-parallel to
797 the shear zone, the displacement is less and less dependent on the volume reduction
798 perpendicular to the cleavage. When the cleavage is sub-parallel to the shear zone
799 (**Fig. 7f**), only the effect of the perpendicular-to-the-shear zone compaction
800 contributes to the dissolution. At this stage, large sliding along parallel-to-the-fault
801 cleavage must switch to grain boundary sliding. This mechanism can always be a
802 pressure solution process at grain size scale in order to accommodate relative grain

803 displacement (Ashby and Verrall, 1973). However, the transfer distance is reduced to
 804 the size of the asperities along the grain boundaries. Consequently, the strain rate of
 805 such viscous behavior increases significantly for a given stress state (Gratier et al.,
 806 2013). This effect contributes to the strain localization parallel to the shear zones. The
 807 system can then evolve to a combined frictional and viscous flow behavior (Bos and
 808 Spiers, 2002) or can depend only on the friction of individual minerals and on the
 809 geometry of their assemblage if mass transfer is no longer possible (Lockner et al.,
 810 2011; Sone et al., 2012; Carpenter et al., 2016). In all cases, this effect increases the
 811 segregation process between shear zones with cleavage oblique to the shear
 812 displacement and shear zones with cleavage sub-parallel to this displacement (**Fig.**
 813 **10a**). The implication for aseismic displacement estimate is discussed in the next
 814 section.
 815



816
 817 **Figure 9.** Plot of finite strain and mass-volume change raw measurements from
 818 various microstructural methods in the displacement plane (*ac*). The principal strain
 819 axis aspect ratio (*R*) is plotted versus the relative mass and volume change (Δ) and the

820 finite shear strain (lines of various colors from $\gamma = 0$ -blue- to $\gamma = 100$ -dark red-).
821 Highly contrasted shear strain values γ are found at the Mülâyim site, depending on
822 the measurement methods, ranging from 0.5 - 2.5 to 40 - >200 using either Fry
823 method (small triangles) or the angle between cleavage and shear zone (large
824 triangle), respectively. Other symbols refer to sites in **Fig. 3** (Taşkesti), **Fig. 4**
825 (Hamamli), **Fig. 5** (Ismetpasa), **Fig. 6** (Yazioren) and **Fig. 7** (Mülâyim). The size of
826 the symbols is larger than the uncertainty except for the high shear strain value
827 deduced from angle measurement where error bars are indicated. Table at top of the
828 figure summarizes strain values at different sites and for different methods: R_{Fry} ,
829 measures of length or angle changes, measures of volume changes from chemical
830 content or length changes with mean percentage uncertainty and calculated shear
831 strain values with specific uncertainty. In all cases only the range of values are given
832 for simplicity.

833

834 **5.2 Strain and width evaluation: uncertainty, heterogeneity and combination** 835 **of two normal shear strain distributions in creeping sections**

836 Uncertainties on strain measurements have been discussed in the methodological
837 section **3** and are given for each measured value in section **4** and **Fig. 9**. However,
838 such uncertainties are valid for homogeneous deformation. The effect of the
839 heterogeneity of the deformation and the validity of extrapolating measurements from
840 thin sections to the entire shear zone are discussed below and integrated in a normal
841 distribution model of the shear strain, which is used to calculate the aseismic
842 displacement in the next section (**5.3**).

843 Several independent strain measurements on several different outcrops show the same
844 trend: data are clustered around the same values in **Fig. 9**. However, observation
845 shows variation in measured strain values: the ratio between the principal strain axes
846 (R) ranges from close to 1 in the massive limestone of Taşkesti (**Fig. 3**), to 8.5
847 (Mülâyim, **Fig. 7**) and 7.5 – 19 (Ismetpasa, **Fig. 4**) in volcanic and analogous rocks.
848 The contrasting behavior of the limestone (locked sections) versus the volcanic rocks
849 (creeping sections) has been discussed in another study (Kaduri et al., 2017), and is
850 only briefly summarized here. In the locked sections, using the low shear strain value
851 of $\gamma = 0.2 \pm 0.05$ for the ductile shear zones and their cumulative width of a few
852 meters leads to calculate a total displacement of few meters, indicating that the creep-
853 related displacement in locked sections is almost negligible. This effect is explained
854 by the contrasting behavior of monomineralic (limestone) rocks and polymineralic
855 (volcanic) rocks. Diffusion at the contact surface between two identical grain minerals
856 as in limestone, whose boundaries are most often healed, is much slower than

857 diffusion at the contact surface between minerals of different nature as in volcanic
858 rocks, where the fluid phase boundary is more continuous (Zubtsov et al., 2004). The
859 difference in behavior is even greater if one of the minerals is a phyllosilicate, which
860 prevents healing (Bos and Spiers, 2000; Zubtsov et al., 2004) or activates dissolution
861 (Renard et al., 2001). This concept that creep deformation is controlled by lithology is
862 supported by a wide range of observations in several major creeping faults (Graymer
863 et al., 2005; Sone et al., 2012; Richard et al., 2014; Thomas et al., 2014; Kaduri et al.,
864 2017).

865 However, in the creeping sections in volcanic or analogous rocks even at the same site
866 the strain values vary widely, for example from $R=1.4$ to 8.5 in Mülayim from the Fry
867 method. This strain ratio is even much greater, up to $R=10000$, when derived from the
868 angles of the cleavage with the shear zone as in Mülayim (**Figs. 7g, 9**). Such
869 heterogeneous pattern is well compatible with the development of the pressure
870 solution layering process discussed in the preceding section (section **5.1**). Not only
871 the volume change but also the strain values vary from one layer to another in a shear
872 zone. This pattern is schematized in **Fig. 10a**. On the left panel, a sketch of the whole
873 kilometer-width NAF shear zone is presented with several faults and parallel-to-the-
874 displacement shear zones (as in **Fig. 8b**). In the middle panel, the hectometer scale is
875 displayed with alternating shear zones with cleavage oblique or sub-parallel to the
876 shear displacement (as in **Fig. 8a**). Finally, on the right panel, we propose a
877 combination of two normal (i.e. Gaussian) shear strain distributions at the scale of the
878 entire NAF shear zone that can be used to evaluate the total aseismic displacement.
879 This model combines (i) a first normal distribution that represents the low strain and
880 wide shear zones with cleavage oblique to the fault zone, with γ_{wsz} and W_{wsz} as
881 maximum shear strain and cumulative width characteristics, respectively, and (ii) a
882 second normal distribution that represents the high strain and narrow shear zones with
883 cleavage sub-parallel to the fault zone, with γ_{nsz} and W_{nsz} as maximum shear strain
884 and cumulative width characteristics. We discuss below how strain measurements on
885 thin sections are extended to regional scale with the associated uncertainties and the
886 evaluation of the cumulative width of these two types of shear zones.

887

888 *Shear zones with cleavage oblique to the shear displacement*

889 The width of the shear zone W_{wsz} is taken as $2\sigma_{wsz}$, where σ_{wsz} is the standard
890 deviation of the normal distribution. There are few possible measurements of the
891 width of this large zone of shearing. Evidence of several parallel faults indicating
892 kilometer width of deformation is seen on all the geological maps (Herece and Akay,
893 2003, Emre et al., 2013, 2016, Akbaş, et al 2016) or in Gerede (**Fig. 8b**), where the
894 presently active fault is located 1 kilometer away from the studied creeping zone. The
895 width W_{wsz} is thus considered to be at least 1 km-wide. Another kilometer-width
896 value can be obtained from the modeling of geodetic data (see section **5.3**).

897 Uncertainty on strain measurements assuming homogeneous deformation is rather
898 low (see table **Fig. 9**). However, heterogeneity is everywhere, amplifying the
899 uncertainty. The normal distribution model implies to choose the highest value of the
900 shear strain γ_{wsz} that is representative of the zones with oblique cleavage (**Fig. 10a**).

901 For a kilometer-width normal distribution, samples for such a γ value representing
902 some meters in size seems realistic. We did not perform strain measurement at this
903 scale but we will consider that measurements on large thin sections at decimeter scale
904 are representative of this meter scale. The higher strain values at thin section scale are
905 from Mulayim and Yaziren sites. The use of the Fry method gives a strain ratio $R =$
906 8.5 ± 0.4 in Mulayim leading to $\gamma = 2.6 \pm 0.2$ (**Fig. 7c**). This value is a minimum due
907 to the limitation of the method to evaluate high strain values, because of the difficulty
908 of measuring a very elongated elliptical void in cloud of points. In Yaziren, it is
909 possible to compare two different methods: boudinage measurements give shear strain
910 values in the range $\gamma = 1.3 \pm 0.1$ to 4.0 ± 0.2 , whereas the Fry method gives $\gamma =$
911 1.3 ± 0.1 (**Fig. 6**). The two methods do not rely on the same markers and the
912 heterogeneity of the deformation with undeformed truncated blocks and highly
913 localized deformation is probably not well accounted for by the Fry method. The
914 boudinage measurement is also limited because with large extension it is difficult to
915 link the truncated blocks. Moreover, the boudinage measurement distributions should
916 also be discussed. As for extension deduced from truncated belemnites (Ramsay,
917 1967), one can either consider that the truncated objects were heterogeneously
918 fragmented in homogeneous deformation and then the maximum extension values are
919 the most representative of the whole deformation, or consider that the deformation
920 was heterogeneous and a mean value is best representative of the whole deformation
921 (Gratier et al., 2013). Here, as we clearly see heterogeneous shear deformation at all

922 scales it seems more realistic to take a mean value of the boudinage measurement of
923 $\sqrt{\lambda_1} = 2.6$, leading to a shear strain of about $\gamma = 2.4$ (with $\Delta = -0.1$), close to the
924 value $\gamma = 2.6 \pm 0.2$ (with $\Delta = 0$) measured in Mulayim. These values must be
925 mitigated by the effect of the volume change ($\Delta = -0.2 \pm 0.15$) as discussed before
926 (section 5.1), leading to a range of shear strain of $\gamma_{wsz} = 1.9 - 2.7$ that will be used
927 in section 5.3.

928

929 *Shear zones with cleavage sub-parallel to the shear displacement*

930 Zones of cleavage sub-parallel to the shear displacement are revealed at thin section
931 scale by two types of observations. The first one results of the statistical analysis of
932 the orientation of elongated cluster of grains, for which the long axis can be
933 assimilated to cleavage planes, that shows normal distribution with very low angles
934 with the shear plane ($\alpha' = 0.3^\circ$), see Fig. 7g. Another key observation is that the
935 shear strain values deduced from particles and length change measurements are
936 sometime higher at decimeter scale than at millimeter scale (Fig. 7). This observation
937 implies the existence of some highly localized zones with very high strain values
938 associated with cleavage sub-parallel to shear displacement (Fig. 7f) that cannot be
939 captured by such measurements. At regional scale, as shown in the Gerede outcrop,
940 there are clear evidences of zones with cleavage sub-parallel to the main shear zone
941 displacement among zones with cleavage oblique to the shear displacement (Fig. 8).
942 These intensively deformed zones include gouges but also some parts of the damage
943 zones that show cleavage planes sub-parallel to the shear zone (Fig. 8f-g). The ratio
944 between the widths of shear zones with parallel-to-the-fault cleavage and the total
945 width has been evaluated in the range 0.04 ± 0.004 . This value can be considered as a
946 representative value at a regional scale (100-1000m). This ratio is used to evaluate the
947 cumulative width of the narrow zones for example $W_{nsz} = 40 \pm 4$ m for a total width of
948 1km.

949 Regarding the reliability of the strain measurements in order to evaluate the maximum
950 value of the normal distribution of the shear strain values, γ_{nsz} , as before, for a
951 kilometer-width normal distribution samples representing some meters in size seem
952 realistic. We used measurements on large thin sections at decimeter scale as
953 representative of meter size. The statistical analysis made in the Mulayim thin section
954 gives a value of $\gamma_{nsz} = 190$ associated with an angle $\alpha' = 0.3^\circ$ with the shear

955 displacement. However, uncertainty on the angle between the normal distribution of
956 the elongated grains cluster and the shear displacement is probably higher than 1° ,
957 leading to γ_{nsz} value ranging from 40 to > 200 (see section 4.5). Using such large
958 uncertainty in the displacement calculation is very difficult, whereas choosing some
959 arbitrary values does not seem realistic either. Moreover we must also integrate the
960 fact that we were not able to evaluate the volume change in this thin section due to the
961 lack of protected zones representative of initial state, despite observations of mass
962 transfer, so we took $\Delta = 0$. But according to other measurements as in **Fig. 5d** it is
963 likely that the volume change in the shear zones with cleavage sub-parallel to the
964 displacement must be very large, up to $\Delta = -0.6$. With such large volume change the
965 shear strain value of 190 decreases down to near 100 (**Fig. 9**). So we finally choose
966 $\gamma_{nsz} = 100$ as a conservative representative order of magnitude. Then, after
967 integrating all the other parameters with lower uncertainty in the modeling to
968 calculate the total displacement, we will be able to evaluate more precisely the
969 possible range of the shear strain values γ_{nsz} in the zones with cleavage sub-parallel
970 to the shear displacement.

971

972 **5.3 Calculation of the aseismic part of the total displacement along creeping** 973 **sections of the North Anatolian Fault**

974 In this section, the relative contribution of aseismic displacement to the total
975 cumulative displacement, along creeping sections of the NAF, is estimated. At the
976 continental scale, rocks deform both by reversible deformation (elastic tectonic
977 loading) and by irreversible deformation (ductile or brittle) that involves plastic
978 dissipative processes and that depends on the system size (Karmakar et al., 2010). We
979 propose to use of a combination of geodetic and geological data in order to evaluate
980 the width of the creeping zone, that we combine with shear strain measurements at the
981 outcrop scale to estimate the aseismic part of the total displacement on the fault
982 (Ramsay, 1980) (**Eq. 4**).

983

984 **5.3.1 Modeling creep behavior at the present time**

985 Geodetic data used in this section are horizontal surface velocity maps across the
986 NAF taken from the InSAR time series analysis of Cetin et al. (2014). These maps are

987 based on Envisat satellite data acquired between 2003 and 2010. Horizontal velocity
988 profiles perpendicular to the fault were extracted from these maps and modeled using:

$$989 \quad V(z) = \frac{S}{\pi} \arctan\left(\frac{z}{d_1}\right) + C \left(\frac{1}{\pi} \arctan\left(\frac{z}{d_2}\right) - H(z)\right), \quad (5)$$

990 where z is the distance perpendicular to the fault. Equation 5 is derived from the
991 classical model of a dislocation into an elastic space of Weertman and Weertman
992 (1964) and Savage and Burford (1973). This equation models surface horizontal
993 velocities V across a strike slip fault locked from the surface to a locking depth d_1 and
994 sliding at the tectonic loading rate S below (first term of the equation). It was
995 modified to take into account creep at velocity C from the surface down to a depth d_2
996 (second term of the equation), as in Hussain et al. (2016). This second term of the
997 equation including the Heaviside step function H thus represents the expected velocity
998 step across fault due to shallow creep.

999 The corresponding shear strain rate $\dot{\epsilon}_{xz}$ along the profile is the spatial derivative of the
1000 velocity profile:

$$1001 \quad \dot{\epsilon}_{xz} = \frac{\partial V(z)}{\partial z} = \frac{S}{\pi} \cdot \frac{d_1}{d_1^2 + z^2} + C \left(\frac{1}{\pi} \cdot \frac{d_2}{d_2^2 + z^2} - \delta(z)\right) \quad (6)$$

1002 where x is the distance along the fault and z along the width of the shear zone (see the
1003 coordinate system in **Fig. 2a**).

1004 **Figure 10 b-c** shows the theoretical surface velocity profiles expected across a fault
1005 that: i) slides at a rate of 20 mm/yr below 15 km (tectonic loading) and is locked
1006 above, ii) creeps between the surface and a depth of 2 km at a rate of 10 mm/yr and is
1007 locked below between 2 km and 15 km, iii) combines both slip processes. These
1008 values correspond to the averaged values of the tectonic loading rate (20 mm/yr),
1009 locking depth (15 km), creep rate (10 mm/yr) and depth of creep (2 km), at the
1010 longitudes of Ismetpasa and Hamamli sampling sites along the creeping section of the
1011 NAF. The first two parameters (tectonic loading rate and locking depth) are based on
1012 GPS measurements (Reilinger et al. 2006). The last two (rate and depth of creep) are
1013 from the creep rate distribution model shown in figure 9 of Cetin et al. (2014) (from
1014 inversion of InSAR data). The corresponding theoretical strain rate profiles are also
1015 derived in **Fig. 10d**.

1016 **Figures 10e-f** show two horizontal velocity profiles derived from Cetin et al. (2014),
1017 perpendicular to the fault (30 km-wide and 240 km-long), and intersecting it at the
1018 Ismetpasa and Hamamli sites. Data are corrected from the tectonic loading at depth,
1019 from (Reilinger et al. 2006), so that the profiles correspond only to the shallow creep

1020 signal. Superimposed on these profiles are models from Cetin et al. (2014), with
 1021 similar parameters as for **Fig. 10b-c** (black lines on **Fig. 10e-f** are computed from the
 1022 right-hand term of **Eq. 5** considering only shallow creep). The creep rate is considered
 1023 to be uniform from the surface to a given depth (10 mm/yr down to 2 km depth).
 1024 Given that the tectonic loading rate is about 25 mm/yr on average for the entire North
 1025 Anatolian Fault (Reilinger et al., 2006; Cetin et al., 2014) and 20 mm/yr near
 1026 Ismetpasa (Cetin et al., 2014), the present-day shallow creep accommodates about 40-
 1027 50% of the tectonic loading.

1028 These models on **Fig. 10e-f** are derived to obtain shear strain rate profiles (**Fig. 10g**)
 1029 at the sampling sites in order to:

- 1030 1) Evaluate the relevance of the proposed combination of two normal (i.e. Gaussian)
 1031 shear strain distributions (**Fig. 10a**).
- 1032 2) Calculate the width of the wide shear zone in the past assuming that it keeps the
 1033 same width (at least the same order of magnitude) throughout the deformation
 1034 process.

1035 We propose to fit the geodetic strain rate model of **Fig. 10g** (green curve) at
 1036 Ismetpasa and Hamamli with a superposition of two normal distributions representing
 1037 the cumulative width and shear strain associated with both types of shear zones (e.g.
 1038 damage and gouge zones in the present time). The shear components can be
 1039 approximated with normal distributions $\gamma \sim N(0, \sigma^2)$ so that:

$$1040 \quad \gamma(z) = \sum_k \gamma_{0,k} \cdot e^{-z^2/2\sigma_k^2} \quad (7)$$

1041 where γ is the shear strain, and index k refers to the type of shear zone. The width of
 1042 the shear zone is considered to be $W_k = 2\sigma_k$. Postulating that the shear strain is the
 1043 only time-dependent variable $\gamma(t)$, the shear strain rate $\dot{\gamma}$ obtained by deriving **Eq. 7**
 1044 over time is $\dot{\gamma}(z) = \sum_k \dot{\gamma}_{0,k} \cdot e^{-z^2/2\sigma_k^2}$.

1045 Considering that strain rate and width values are independent for the damage zone and
 1046 the gouge zone, the strain rate and width of the damage zone are first inverted: $\dot{\gamma}_d =$
 1047 $1.6 \cdot 10^{-6} yr^{-1}$, and $W_d = 4.4$ km, respectively (**Fig. 10g**, orange normal distribution).
 1048 This value of several kilometers for the width of the damage zone is consistent with
 1049 the kilometer scale fault zone width observed in the field (**Fig. 8b**). Owing to the low
 1050 resolution (pixel size 20 m east-west, 4 m north-south) and high noise level (**Fig. 10e-**
 1051 **f**) of InSAR satellite data it is not possible to provide a strain rate model in the fault
 1052 near-field that is sufficiently precise to estimate the width of the gouge zone properly.

1053 This width is therefore taken as $W_g=1$ m, based on field measurements from Kaduri et
 1054 al. (2017). The strain rate of the gouge zone is finally estimated as follows. The
 1055 cumulative displacement rate along a fault, \dot{D} , is given by the integration of the two
 1056 normal distributions over a z-axis perpendicular to the fault plane, in a range
 1057 $(-\infty, +\infty)$ (placing **Eq. 7** in **Eq. 4**):

$$\begin{aligned}
 1058 \quad \dot{D} &= \dot{\gamma}_g \cdot \int_{-\infty}^{\infty} e^{-z^2/2\sigma_g^2} dz + \dot{\gamma}_d \cdot \int_{-\infty}^{\infty} e^{-z^2/2\sigma_d^2} dz \\
 1059 \quad &= \dot{\gamma}_g \cdot \sigma_g \cdot \sqrt{2\pi} + \dot{\gamma}_d \cdot \sigma_d \cdot \sqrt{2\pi} \quad (8)
 \end{aligned}$$

1060 Assuming that \dot{D} is 10 mm/yr based on the InSAR analysis of Cetin et al., (2014),
 1061 **(Fig. 10c-d)**, and given the values of $\dot{\gamma}_d$ and W_d obtained above, it is estimated that
 1062 $\dot{\gamma}_g = 0.89 \cdot 10^{-3} \text{ yr}^{-1}$ from **Eq. 8**, with $W_k = 2\sigma_k$. It is worth noting that this
 1063 partition of the displacement rate between damage zone and gouge is strongly
 1064 dependent on the imposed width of the gouge, and significant detailed improvements
 1065 of this partition would rely on a better spatial definition of present-day geodetic
 1066 measurements. However, this formalism has the merit of representing in a consistent
 1067 framework both structural geologic and geodesic data, which have different spatial
 1068 and temporal scales.

1069

1070 **5.3.2 Modeling the creeping behavior in the past since the beginning of fault** 1071 **sliding**

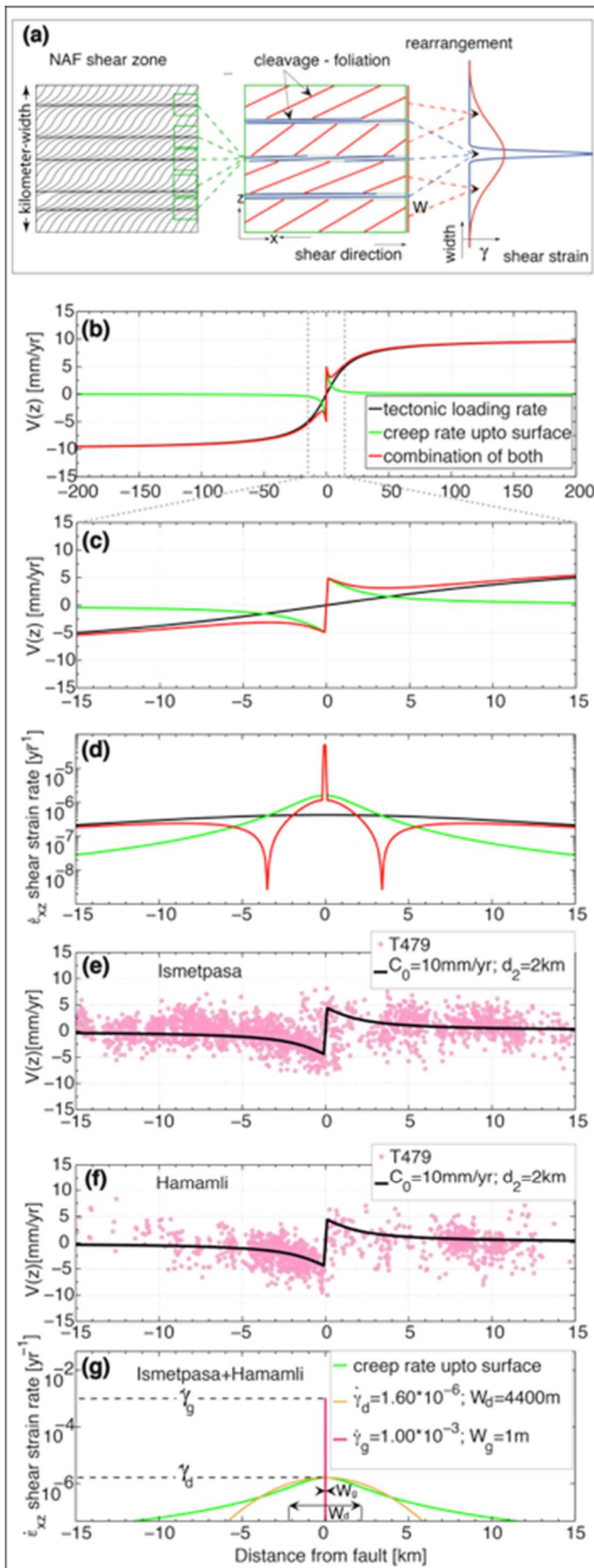
1072 As discussed in section 5.2, two types of strain distributions must be taken into
 1073 account when calculating the displacement associated with creep processes since the
 1074 beginning of fault sliding **(Fig. 10a)**. Firstly, a normal distribution represents the low
 1075 strain and wide shear zones with cleavage oblique to the fault zone, with γ_{wsz} and
 1076 W_{wsz} as maximum shear strain and width characteristics, respectively. Secondly,
 1077 another normal distribution represents the high strain and narrow shear zones with
 1078 cleavage sub-parallel to the fault zone, with γ_{nsz} and W_{nsz} as maximum shear strain
 1079 and width characteristics, respectively. Based on geodetic data the width of the wide
 1080 shear zone W_{wsz} is evaluated to be 4.4 km **(Fig. 10g)**. As discussed in section 5.2,
 1081 geological observations near Gerede lead to a minimum width of the fault zone of 1
 1082 km **(Fig. 8b)**. Strain measurements on field samples give shear strain values ranging
 1083 from $\gamma_{wsz} = 1.9$ to 2.7 when integrating the uncertainty on strain measurements and
 1084 volume change at the scale of the NAF shear zone (see section 5.2). For the narrow

1085 shear zones, as discussed in section 5.2, the width of the narrow zones can vary from
1086 $0.036 \times 1000\text{m} = 36 \text{ m}$ to $0.044 \times 4400\text{m} = 194 \text{ m}$ and a conservative value of $\gamma_{nsz} =$
1087 100 is used for the shear strain (see section 5.2). All these strain and width values can
1088 be used to evaluate the limits on the total displacement associated with aseismic
1089 creep, by integrating the shear strain values of the two normal distributions across the
1090 fault. The maximum and minimum aseismic displacement is calculated using Eq. 8
1091 with γ instead of $\dot{\gamma}$. The maximum displacement results from the combination of the
1092 maximum displacement associated with the low strain and wide shear zones (using
1093 the maximum values $\gamma_{wsz} = 2.7$ and $W_{wsz} = 4.4 \text{ km}$, (corresponding to $\sigma_{wsz} = 2.2 \text{ km}$),
1094 which is 15 km, and of the maximum displacement associated with the high strain and
1095 narrow shear zone (maximum values $\gamma_{nsz} = 100$ and $W_{nsz} = 194 \text{ m}$ (corresponding to
1096 $\sigma_{wsz} = 97\text{m}$), which is 24 km, leading to a total maximum displacement of 39 km.
1097 Similarly, the minimum displacement is estimated using the minimum values of
1098 γ_{wsz} and W_{wsz} for the low strain and wide shear zone ($\gamma_{wsz} = 1.9$; $W_{wsz} = 1 \text{ km}$),
1099 giving 2.4 km of displacement, and the minimum values of γ_{nsz} and W_{nsz} for the high
1100 strain and narrow shear zone ($\gamma_{nsz} = 100$; $W_{nsz} = 36 \text{ m}$), giving 4.5 km of
1101 displacement, leading to a total minimum displacement of 6.9 km. The estimated total
1102 aseismic displacement thus ranges between 9 % and 49 %, of the total shallow
1103 displacement of 80 km during geological times.

1104 The conservative value $\gamma_{nsz} = 100$ can now be discussed in relation with geological
1105 observations. Using shear strain values of $\gamma_{nsz} = 200$ for the zone with cleavage sub-
1106 parallel to the shear displacement would mean that nearly 80% of the total
1107 displacement was accommodated by aseismic creep, which seems a maximum since
1108 we see evidence of earthquakes all along the NAF deformed zone. Using shear strain
1109 values of γ_{nsz} less than 70 would mean that the zones with cleavage sub-parallel to the
1110 shear zone accommodated less displacement than the wide zones, which seems rather
1111 unrealistic. Consequently, these two shear strain values may be considered as upper
1112 and lower bound of the shear strain values in zone with cleavage sub-parallel to the
1113 shear displacement.

1114 From a more general point of view, this type of calculation reveals the difficulty in
1115 quantifying large displacements along regional shear zones. Measurements based on
1116 length changes are limited to shear strain below 2.5 to 4, depending on the methods.
1117 Measurements based on angle changes are much more efficient but they need to be

1118 done with reliable statistical analysis that still need to be developed on the field with
1119 numerous cleavage measurements, which was not possible to do here because of the
1120 quality of the outcrops.



1121

1122 **Figure 10.** Aseismic strain along the North Anatolian Fault. (a) Schematic representation of
1123 the strain distribution in the creeping zone, which is made up of parallel zones of various
1124 strain values representative of **Fig. 8b** and **8a** from left to right, respectively, and that are
1125 represented on the right by two normal (i.e. Gaussian) distributions of the shear strain (b)
1126 Horizontal surface velocity as a function of distance from fault (black profile is for a fault
1127 locked at the surface and creeping at depth, green profile is for a fault creeping at the surface
1128 and locked at depth, red profile is for a combination of both). (c) Zoom on (b). (d) Surface
1129 strain rate as a function of distance from fault, derived from (c). (e) & (f) Geodetic horizontal
1130 creep rate profiles perpendicular to the North Anatolian Fault across Ismetpasa and Hamamli
1131 outcrops, respectively; the data (dots) and model (black lines) are taken from Cetin et al.
1132 (2014). (g) Strain rate profiles derived from creep rate models in (e) & (f). These profiles are
1133 fitted with the superposition of two normal distributions to model the shear strains and the
1134 widths of the gouge and the damage zones for Ismetpasa-Hamamli outcrops.
1135

1136 **6. Conclusion**

1137 The quantitative evaluation of the aseismic part of the total displacement along the
1138 North Anatolian Fault provides methodological and rheological insights on strain
1139 measurements and fault mechanisms. From a methodological point of view, it has
1140 been shown that the measurement of finite strain in shear zones, which can reach very
1141 large strain values, is strongly dependent on the methodology. Methods based on the
1142 analysis of particle distribution, such as the Fry method, or length change, such as
1143 boudinage measurements, do not allow retrieving shear strains higher than 2.5 to 4,
1144 respectively. Conversely, using the variations in cleavage angle with the shear zone
1145 very large strain values up to 100 or more can be estimated, with a large uncertainty
1146 on this value. However, in both cases, evaluation of volume change is important and
1147 must be integrated in the calculation.

1148 The relationship between strain and volume change shows that the penetrative
1149 cleavage observed in the whole NAF shear zone affecting volcanic, or analogous
1150 rocks, is a pressure solution cleavage that evolves from oblique to sub-parallel to the
1151 shear displacement. A combination of two normal (i.e. Gaussian) shear strain
1152 distributions is proposed to represent this pattern of deformation in the kilometer-
1153 width creeping zones. These zones are partitioned into parallel deformation zones
1154 with contrasting shear strain characteristics: (i) wide (10-100 m) shear zones with
1155 cleavage oblique to the shear displacement (ii) narrow (0.2-2 m) shear zones with
1156 cleavage planes sub-parallel to the shear displacement, with sub-grain shear
1157 localization.

1158 Finally, finite shear strain characteristics evaluated from geological analyses were
1159 used in parallel with geodetic data in order to calculate the total cumulative shallow
1160 aseismic displacement that has been accommodated by creep. By modeling the
1161 present-day creep behavior, it is possible to propose a model with the two types of
1162 shear zones and evaluate the maximum width of the creeping zone. By modeling the
1163 creeping behavior since the beginning of fault sliding it is possible to calculate an
1164 aseismic part of the total 80 km displacement in the range 9 to 49% in the creeping
1165 (volcanic and analogous rocks) zone and a negligible displacement in the locked
1166 (limestone) zones, showing the crucial effect of the lithology in the seismic versus
1167 aseismic partition.

1168 The large uncertainty in the evaluation of the NAF aseismic displacement is partly
1169 related to the difficulty of evaluating the large strain values that are associated with
1170 large shear displacement. Based on the methodology developed on thin section
1171 analysis, a promising way to improve such measurements would be to develop
1172 reliable statistical analysis of cleavage orientations in the field.

1173

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