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Anisotropic scaling of tectonic stylolites: a fossilized signature of the stress field?

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21 **Abstract**

22 Vertical stylolites are pressure solution features, which are considered to be caused by
23 horizontal tectonic loading with the largest principal compressive stress being (sub) parallel to
24 the earth surface. In the present study we analyze the roughness of such tectonic stylolites
25 from two different tectonic settings in southern Germany and north-eastern Spain aiming to
26 investigate their scaling properties with respect to the stress during formation. High resolution
27 laser profilometry has been carried out on opened stylolite surfaces of nine samples. These
28 datasets were then analyzed using 1D and 2D Fourier power spectral approaches. We found
29 that tectonic stylolites show two self-affine scaling regimes separated by a distinct crossover-
30 length (L), as known for bedding parallel stylolites. In addition tectonic stylolites exhibit a
31 clear in-plane scaling anisotropy which modifies L . Since the largest and smallest crossover-
32 lengths are oriented with the sample vertical and horizontal directions (i.e. σ_2 and σ_3) and L is
33 a function of the stress field during formation as analytically predicted we conclude that the
34 scaling anisotropy of tectonic stylolites is possibly a function of the stress field. Knowledge of
35 this crossover-length anisotropy would enable the reconstruction of the full 3D stress tensor if
36 independent constraints of the depth of formation can be obtained.

37

38 **1. Introduction**

39 The intriguing variety of pressure solution features and its wide-spread occurrence in
40 monomineralic rock types provoked a continuous interest and attention in various geoscience
41 disciplines over the past decades [*Tada and Siever, 1989*]. One of the most prominent and
42 complex pressure solution features are stylolites, which are rough dissolution interfaces that
43 can be found in a large variety of sedimentary rocks [*Buxton and Sibley, 1981; Dunnington,*
44 *1954; Heald, 1955; Park and Schot, 1968; Railsback, 1993; Rutter, 1983; Stockdale, 1922;*
45 *Tada and Siever, 1989*]. Until recently stylolite morphology has been described qualitatively
46 by the use of a descriptive terminology, which grouped stylolites into generic classes. One

47 classification uses the orientation of the stylolite plane relative to bedding. Bedding-parallel
48 stylolites are supposed to have formed due to the layer-normal overburden pressure, while
49 tectonic stresses cause the formation of stylolites oblique or perpendicular to bedding [*Park*
50 *and Schot*, 1968; *Railsback and Andrews*, 1995]. A second classification is based on the
51 orientation of the stylolite teeth relative to the stylolite plane. Here the term "stylolite" is used
52 for teeth at a high angle to the plane, and 'slickolite' for dissolution surfaces where the teeth
53 are distinctly oblique to the dissolution plane [*Bretz*, 1940; *Gratier et al.*, 2005; *Simon*, 2007].
54 Finally the shape of the characteristic teeth-like asperities and spikes along the interface has
55 been used to characterize stylolites [*Guzzetta*, 1984; *Park and Schot*, 1968].

56 More recently, stylolites have been subjected to more rigorous quantitative analyses to
57 characterise the roughness of the stylolite surface [*Brouste et al.*, 2007; *Drummond and*
58 *Sexton*, 1998; *Ebner et al.*, 2009a; *Ebner et al.*, 2009b; c; *Gratier et al.*, 2005; *Karcz and*
59 *Scholz*, 2003; *Koehn et al.*, 2007; *Renard et al.*, 2004; *Schmittbuhl et al.*, 2004]. It was
60 demonstrated that the 1D stylolite roughness obeys a fractal scaling invariance [*Drummond*
61 *and Sexton*, 1998; *Karcz and Scholz*, 2003]. Investigation of the rough interface of opened
62 stylolite surfaces by means of laser profilometry revealed that the stylolite morphology shows
63 two self-affine scaling regimes with two distinct roughness exponents on their respective
64 scales, which are separated by a characteristic crossover length at the millimeter scale
65 [*Renard et al.*, 2004; *Schmittbuhl et al.*, 2004] for bedding parallel stylolites. Self-affine
66 surfaces define a group of fractals, which remain statistically unchanged by the transform:
67 $\Delta x \rightarrow b \cdot \Delta x$, $\Delta y \rightarrow b \cdot \Delta y$, $\Delta z \rightarrow b^H \cdot \Delta z$, where b is a transformation factor, which can take any real
68 value and H is the Hurst or roughness exponent [*Barabasi and Stanley*, 1995], which is a
69 quantitative measure for the roughness of the signal.

70 Analytical and numerical investigations demonstrated that the growth of the stylolite
71 roughness is induced by heterogeneities in the host rock that pin the interface and is slowed
72 down by two stabilizing forces, the elastic and surface energies. The elastic energy dominates

73 on larger scales and is represented by a small roughness exponent of 0.3 to 0.5 whereas the
74 surface energy is dominant on small scales with a roughness exponent of about 1 [Koehn *et*
75 *al.*, 2007; Renard *et al.*, 2004; Schmittbuhl *et al.*, 2004]. The characteristic crossover length
76 (L) that separates these two scaling regimes is a function of the principal normal stress
77 [Renard *et al.*, 2004; Schmittbuhl *et al.*, 2004] on the interface of a bedding parallel stylolite
78 This analytical predictions were successfully tested by Ebner *et al.* [2009b], who
79 demonstrated on a set of 13 bedding parallel stylolites from varying stratigraphic depth out of
80 a cretaceous succession that this crossover-length decreases with increasing depth (and
81 normal stress) and thus exhibit the analytically predicted behaviour. The 1D scaling of
82 stylolites with two self-affine scaling invariance regimes can be described as the height
83 difference h of points along the surface separated by a distance Δx as [Ebner *et al.*, 2009b]

$$84 \quad h(\Delta x) \approx A \Delta x^{H_s} g(\Delta x / L) \quad \text{with} \quad g(u) = \begin{cases} u^0 & \text{if } u \ll 1 \\ u^{H_L - H_s} & \text{if } u \gg 1 \end{cases} \quad (1)$$

85 where A is a scaling factor, g is a scaling function and u is the ratio $\Delta x/L$ with L being a
86 crossover-length. H_s , H_L correspond to the roughness exponents for small and large scales,
87 respectively. Numerical simulations also demonstrate that the crossover-length is very robust
88 with regard to the kind and amount of quenched noise (heterogeneities initially present) in the
89 rock [Ebner *et al.*, 2009a]. Hence, the use of bedding parallel stylolites as a quantitative stress
90 gauge under the assumption of uniaxial strain (zero horizontal displacement) seems to be
91 verified. Investigations of the surface morphology of bedding parallel stylolites showed that
92 their scaling is isotropic within the plane defined by the stylolite [Renard *et al.*, 2004;
93 Schmittbuhl *et al.*, 2004]. This implies that any arbitrary section through the stylolite interface
94 that contains the principal stress direction (i.e. normal to the plane) fully characterizes the
95 complex self-affine roughness of bedding parallel stylolites. A second mechanism claimed to
96 be responsible for the formation of the characteristic roughness is a stress induced roughening
97 instability along an initially flat solid-solid interfaces [Angheluta *et al.*, 2008] or a solid-fluid-

98 solid interface [Bonnetier *et al.*, 2009]. In both cases the instability is triggered by elastic
99 stresses acting normal on the interface.

100 Up to now no study has quantitatively investigated the 3D topography of tectonic
101 stylolites, which formed due to (sub-)horizontal compression resulting in a vertical stylolite
102 plane. Tectonic stylolites differ in two major characteristics from bedding parallel stylolites.
103 First, the stress field during the formation of tectonic stylolites is non-isotropic i.e. the in-
104 plane normal stresses differ (i.e. $\sigma_{zz} > \sigma_{xx}$) whereas bedding parallel stylolites often have equal
105 in-plane normal stresses $\sigma_{xx} = \sigma_{yy}$ (Figure 1). This would imply that the scaling of tectonic
106 stylolites is not invariant within the plane, since the crossover-length should scale with the
107 (non-isotropic) stress field as was shown analytically [Schmittbuhl *et al.*, 2004]. A second
108 common feature of tectonic stylolites are oblique/tilted teeth with respect to the mean stylolite
109 plane due to overprinting of pre-existing planes of anisotropy such as joints, bedding planes
110 and other interfaces. Tilting of the teeth with respect to the stylolite plane also influences the
111 morphology because it leads to the dominance of long grooves and ridges [Simon, 2007].
112 These features could lead to an anisotropic scaling of the stylolite interface in addition to
113 variations of the in-plane stresses.

114 The present study investigates such tectonic stylolites which formed in a vertical
115 orientation. We mainly concentrate on the influence of (i) the orientation of the dissolution
116 surface with respect to the displacement direction and (ii) the formation stress on the scaling
117 properties of natural stylolites in limestones. To accomplish this task we use laser
118 profilometry data of opened interfaces of tectonic stylolites from flat lying Jurassic limestones
119 of the Swabian Alb in southern Germany and from a Tertiary fold and thrust belt of the
120 Iberian Chain of north-eastern Spain.

121

122 2. Geological setting

123 In the following section we give a brief introduction of the investigated field areas in
124 southern Germany and north-eastern Spain, which both expose upper Jurassic limestones. The
125 Swabian Alb of southern Germany forms a region of flat-lying mainly marine Jurassic
126 deposits [Geyer and Gwinner, 1991]. The studied sections are located 10 km south of the city
127 of Tübingen and comprise upper Jurassic (Oxfordian to Kimmeridgian) limestones. The basal
128 part of the sections (*UTM 32U; E 0515212 m; N 5362240 m*) are made up of well bedded
129 Oxfordian limestones whereas the upper part of the profile contains massive Kimmeridgian
130 limestones representing a riff facies with sponges and algae being the main rock forming
131 species [Etzold et al., 1996; Geyer and Gwinner, 1991]. The bedding is (sub-) horizontal,
132 dipping slightly ($<5^\circ$) to the SE on a regional scale. The principal structural features are ENE-
133 WSW striking graben structures, which exhibit a mixed mode displacement with a major
134 normal and a subordinate dextral component [Etzold et al., 1996; Geyer and Gwinner, 1991]
135 and can be attributed to a later compressional phase (see below). The investigated stylolites
136 (*Samples: Sa6/1a, Sa6/1b, Sa9/2*) form vertical planes which trend WNW-ESE with teeth
137 pointing parallel to the surface normal direction, hence recording a NNE-SSW compression
138 (Figure 2a). Additionally small scale reverse-faults and NNE-SSW trending joints confirm the
139 same kinematic framework. A younger subordinate set of stylolites not investigated in this
140 study form NE-SW trending vertical stylolite planes which can be related to the prominent
141 dextral graben structures found in the area [Geyer and Gwinner, 1991; Kley and Voigt, 2008].
142 Our relative chronological sequence of deformation events is in agreement with data reported
143 by Kley and Voigt [2008], demonstrating a change in the stress field from NNE-SSW directed
144 compression in the late Cretaceous to a NW-SE directed compression in the Neogene. This
145 second compression phase neither altered the shape nor the orientation of the investigated
146 stylolites, since layer parallel shortening did not cause any orientational change and
147 deformation was restricted to stylolite interfaces.

148 The Iberian Chain of north-eastern Spain is located south of the Ebro-basin and trends
149 roughly NW-SE. The succession is composed of up to 6000 m of Mesozoic, mainly Jurassic
150 and Cretaceous sediments [*Capote et al.*, 2002], although the sequence is significantly
151 reduced to only 300-400 m in the investigated area. The investigated area belongs to the
152 Maestrazgo structural domain which forms the transition zone between the NW-SE striking
153 fold and thrust belt of the Aragon Branch and the NE-SW striking Catalanian Coastal Ranges.
154 A regional NNW-SSE compression in the sampling area between the small towns of Molinos
155 and Ejulve is indicated by ENE-WSW striking 100-1000 m scale fold trains with top to the
156 NNW kinematics. The onset of deformation is estimated to be around Early to Middle
157 Eocene, whereas the deformational peak is assumed to be during the Oligocene [*Capote et al.*,
158 2002; *Casas et al.*, 2000; *Liesa and Simón*, 2009]. *Liesa and Simón* [2009] report stylolite
159 data which they argue to be attributed to Betic and Guadarrama compressions both having
160 their deformational peaks during the Oligocene. The investigated section (*UTM 30T; E*
161 *07111963 m; N 4518336 m*) comprises well bedded limestones in an upper Jurassic upright
162 antiform which contains several smaller synforms that plunges 25° to the NW. Stylolites were
163 investigated in a shallow ENE dipping limb (*set A* in Figure 2b) and from an overturned limb
164 which dips steeply to the SE (*set B* in Figure 2b). In the eastwards-dipping limb of the fold the
165 stylolites (*Samples: M4/1, M4/2, M4/3, M4/4*) track the far field shortening direction (SSE-
166 NNW) confirmed from field measurements in other outcrops in the area. In the overturned
167 and steeply dipping fold-limb the stylolites (*Samples: M4c/1, M4c/3*) are rotated around the
168 fold axis into a shallow dipping orientation (i.e. a counter-clockwise or clockwise rotation of
169 65° around the fold axis would transform the stylolite orientation from one limb into the
170 orientation of the stylolites in the other limb of the fold). Hence, the stylolite formation in this
171 outcrop predates the folding event. In addition the angle between the stylolite plane and the
172 bedding (not shown) is consistent in both positions of the fold thus corroborating the evidence
173 that stylolitization predates the folding event. It has to be noted that stylolites in set A and B

174 both form in a vertical orientation. Another important feature to notice is that the stylolite
175 teeth are somewhat oblique ($\sim 10^\circ$) to the mean stylolite plane, which we interpret as a result
176 of pressure-solution overprint of a pre-existing joint-set which strikes NE-SW, sub-parallel to
177 the stylolite planes.

178

179 **3. Methodology**

180 The samples collected in the locations described above were all taken oriented in the
181 outcrop to reconstruct the spatial position of the 3D stylolite morphology after laser
182 profilometry. For analysis only “closed” specimens were considered. Stylolite surfaces that
183 were already open in the outcrop and were subjected to an unknown amount of weathering
184 were ignored. The sampled specimens were opened mechanically along the two opposing
185 interfaces of the stylolite. This method causes some negligible damage to the surface due to
186 the interlocking of asperities. The split surfaces were cleaned from any clay material, i.e. the
187 residuum of the dissolved rock, with a soft brush and distilled water. Areas which did not
188 exhibit visual mechanical damage were chosen for profilometry.

189 Optical profilometry is based on a laser triangulation of the surface similar to previous
190 studies [*Renard et al.*, 2004; *Schmittbuhl et al.*, 2004; *Schmittbuhl et al.*, 2008]. The
191 triangulation technique uses a laser beam that is focused on the surface of the object, which is
192 monitored by a nearby CCD sensor. The distance between the object and the sensor changes
193 as a function of changes of the angle under which the point of consideration is observed. The
194 distance between the object and the laser-head is then calculated from angular relationships
195 [*Schmittbuhl et al.*, 2008]. Before every individual measurement a test run was made to
196 calibrate voltage fluctuations of the laser beam (voltage-height relationship is virtually linear in
197 the chosen range, which gives the estimate of the vertical resolution – small distortions of the
198 profile height, less than 1%, can be expected.). The laser beam is 30 μm wide and horizontal
199 steps between measurement points were $\Delta x = \Delta y = 25\mu\text{m}$ with a horizontal precision of 1 μm .

200 The vertical resolution is $2\mu\text{m}$. Maps were constructed by movement of the laser-head along
201 parallel profiles over the specimen (Figure 3). Eight samples have been measured at high
202 resolution ($\Delta x = \Delta y = 25\mu\text{m}$) with map sizes of 1200×1200 (*Samples: M4/1, M4/4*), 1600×1600
203 (*Samples: Sa6/1a, Sa6/1b, M4/2, M4/3, M4c/1, M4c/3*) & 2000×2000 measurement points
204 (*Sample: Sa9/2*), which corresponds to square maps with physical side lengths of 30, 40 and
205 50 mm. The x- and y-directions are arbitrary choices parallel to the principal axis of the
206 profilometer. The sample is usually oriented in a way to fit the biggest square map on the
207 respective stylolite interface. Care was taken that from the orientation of map x/y direction the
208 sample orientation could be reconstructed.

209 Additionally Sample Sa6/1 was measured twice where the second measurement
210 (Sa6/1b) was rotated 32° clockwise around a vertical axis with respect to the first
211 measurement (Sa6/1a). This was done to test the robustness of the measurements used against
212 possible noise arising from the measurement procedure along discrete profiles. An image
213 registration [Goshtasby, 1986; 1988] of the two measurements in spatial domain revealed the
214 same amount of rotation of 32° with an uncorrelated noise in the height difference between
215 the two images that arises from the discreteness of the two maps (not shown). This height
216 difference is less than 5% (i.e. the ratio of the standard deviation σ of the height difference is
217 0.063 mm to σ of the height of the surface 1.477 mm). Hence, there seems to be no significant
218 error introduced by the measurement procedure.

219

220 **4. Data analysis**

221 Before we analyzed the 2D maps in detail the raw data from the laser profilometry was
222 subjected to a series of pre-treatments (Figure 4). First a mean plane calculated from a least
223 square fit was subtracted from the raw data (Figure 4a), i.e. the x/y-plane is adjusted to a
224 global trend and the vertical (z) axis is set to have zero mean height (Figure 4b). To increase
225 the quality of our Fourier transforms (described below) we used a Hanning window technique

226 [Karcz and Scholz, 2003; Press et al., 2007] to force our data to taper to zero at the
227 boundaries (Figure 4c) in order to reduce spectral leakage (compare Figure 3). This is a
228 standard technique in signal processing, which does not modify the frequency and amplitude
229 of the original signal.

230

231 **4.1. 1D analysis**

232 From the 2D height-field a 1D profile can be extracted either along the x or y-direction
233 or in any arbitrary direction. For an arbitrary 1D profile $f(x)$ the Fourier transform $F(k)$ can be
234 calculated and the power spectrum $P(k) \sim |F(k)|^2$ of the transform can be plotted as a function
235 of the wavenumber $k=2\pi/\lambda$ [m^{-1}], which scales inversely to the wavelength λ [Renard et al.,
236 2004; Schmittbuhl et al., 1995; Schmittbuhl et al., 2004]. In Figure 5 the averaged spectra of
237 Sample M4/3 along the x and y direction of the measured map are shown. The averaged
238 spectra are found from calculating the mean of $P(k)$ for every k -value over all 1D profiles in
239 one direction [Renard et al., 2004; Schmittbuhl et al., 2004]. This averaging procedure
240 reduces the noise attached to an individual 1D profile. A linear slope of the power spectra
241 confirms a self-affine scaling invariance. The power spectrum of a self-affine signal behaves
242 as

$$243 \quad P(k) \sim k^{-D-2H}, \quad (2)$$

244 where D is the topological dimension of the signal ($D=1$ for 1D profiles) and H the Hurst
245 exponent. The Hurst exponent can thus be calculated from the slope of the power spectra.
246 When we study the averaged 1D spectra of a tectonic stylolite along specific directions
247 (Figure 5a) we see that the signal exhibits two slopes, which are separated by a crossover-
248 length (L) in agreement with observations on bedding parallel stylolites [Ebner et al., 2009b;
249 Renard et al., 2004; Schmittbuhl et al., 2004]. The two observed scaling regimes have typical
250 Hurst exponents of $H_S \sim 0.5$ and $H_L \sim 1.1$ for the small and large scale (large and small
251 wavenumber), respectively. These observations indicate that the scaling of bedding parallel

252 stylolites (Eq. 1) can be extended to tectonic stylolites (compare Figure 5a). To enable a more
 253 detailed comparison of the power spectra of our tectonic stylolites from two different
 254 (orthogonal) directions we normalize the power spectra along the x-direction with the power
 255 spectrum of the y direction at $k=1[\text{mm}^{-1}]$ i.e. $P_x(k)/P_y(1[\text{mm}^{-1}])$ as shown in the inset of Figure
 256 5a. This normalization yields a collapse of the large k-values (small scales), but a notable
 257 difference for the small k-values (large scales) of the scaling functions. This is basically the
 258 expression of the shift in cut-off between the two linear sub-branches, which is the crossover-
 259 length L . Figure 5b shows that the calculated cut-off between the scaling regimes and thus
 260 crossover length differs between them. With 1.22 and 0.62 mm for the x and y-directions the
 261 crossover-length changes by 0.6 mm (Figure 5b). The non-linear fitting plotted as a solid line
 262 in both panels of Figure 5b is a linear-by-parts least square fit in logarithmic space with a
 263 weighting function that changes from the small scale to the large scale fraction of the scaling
 264 law [for details compare Ebner et al., 2009b]. This non-linear model uses a minimization
 265 algorithm to find the least square fit for the crossover-length. The differences found between
 266 the two directions also include a discrepancy in the scaling pre-factor, i.e. a vertical shift of
 267 the power spectra, which is clearly higher for all scales in the y-direction.

268 To fully quantify rough surfaces it is necessary to characterise this pre-factor of the
 269 scaling function and thus obtain a full description of the surface morphology. In the following
 270 we use the height-height correlation function, to calculate the scaling prefactor. The height-
 271 height correlation function [Barabasi and Stanley, 1995], which is defined for a function $h(x)$
 272 over the spatial variable x by, $C(\Delta x) = \left[\langle (h(x) - h(x + \Delta x))^2 \rangle \right]^{1/2}$, where $\langle \rangle$ denotes average over
 273 the range of x , which estimates the average height difference between two points of the profile
 274 separated by a distance Δx . For a self-affine profile, the correlation-function follows a power-
 275 law such that $C(\Delta x) \sim t^{1-H} \Delta x^H$, where H is the Hurst exponent and t is the scaling prefactor.
 276 The prefactor can be designed as $C(t)=t$, and thus denotes a length scale, also known as the

277 topothesy [Renard *et al.*, 2006; Schmittbuhl *et al.*, 2008; Simonsen *et al.*, 2000]. The
278 topothesy corresponds physically to the length scale for which the slope of the rough profile is
279 equal to 1. In other words, t is the theoretical length scale over which the rough profile has a
280 mean slope of 45° . The smaller t , the flatter the profile appears on a macroscopic scale.

281 Figure 6a shows a scaling of the correlation function with two linear sub-branches
282 separated by a crossover-length similar to the scaling of the power spectra shown in Figure 5a
283 with only the slopes being different. The correlation function shows, similar to the power
284 spectra, two linear sub-branches separated by a distinct crossover-length. We use the same
285 nonlinear fitting approach as described above (with fixed Hurst exponents of 0.6 and 0.3).
286 The different scaling exponents compared to the power spectral approach is inline with
287 reliability of self-affine measurements performed on synthetic signals [Candela *et al.*, 2009;
288 Schmittbuhl *et al.*, 1995]. These authors have demonstrated that the correlation function
289 underestimates the input Hurst exponents and thus shows lower values than the power spectra.
290 The scaling prefactor and thus the topothesies t_s and t_l for the small and large scale regimes
291 can be found by intersection of the two sub-branches of the scaling function with the 1/1 line
292 (Figure 6a). We estimated the topothesy for all orientations on the surfaces (Figure 6b & c)
293 and found that there is a weak anisotropy in the scaling pre-factor, which shows a correlation
294 with the highest topothesies being parallel to the horizontal direction in the sample orientation
295 (Figure 6b) for most samples but this is only visible in the small scale regime. This
296 observation is similar to what we found from investigation of the power spectra where the
297 small scale regime is shows very consistent results but the large scale regime reveals a higher
298 degree of variability e.g. compare inset in Figure 5a. The small scale topothesy are shown in
299 Figure 6c. The average topothesies range between 0.05-0.15 mm and 0.15-0.3 mm for small
300 and large scales, respectively.

301 Both the power spectra (i.e. the cut-off length between the linear sub-branches) and
302 topothesy of a 1D signal show a considerable degree of anisotropy which is often obscured

303 due to the noise associated with an individual 1D profile. We conclude that to account for this
304 in-plane variation a 1D signal fails to capture all scaling characteristics of tectonic stylolites
305 and the choice of the investigated profile is not arbitrary as for bedding parallel stylolites.
306 Hence, tectonic stylolites have a measurable in-plane anisotropy which we want to
307 characterize in detail with a 2D approach.

308

309 **4.2. 2D analysis**

310 For the 2D analysis we used the processed data as described in section 4 (Figure 4c).
311 First a 2D Fourier transform i.e. a discrete Fourier transform (DFT) was calculated from the
312 data points of the 2D height-field with the Fast Fourier Transform (FFT) algorithm [Cooley
313 and Tukey, 1965] implemented in *Matlab*[®]. Then the DFT is shifted so that the zero-
314 frequency component lies in the centre of the spectra and the 2D power spectrum $P(k_x, k_y)$ is
315 again calculated as the square of the absolute magnitude of the Fourier transform. Figure 7a
316 displays a map of the 2D power spectra $P(k_x, k_y)$ in which the absolute magnitude squared is
317 shown as greyscale values and k_x and k_y range from $-((n/2)*\Delta x)^{-1}$ to $((n/2)*\Delta x)^{-1}$ where n is
318 the number of measurement points in one direction of the map and $\Delta x = \Delta y$ is the step size. To
319 investigate the power law behaviour located in the 1D signals the 2D power spectra had to be
320 transformed to a double logarithmic space originating from the centre of the map i.e. the zero
321 frequency component or the smallest wavenumber. This is accomplished by translating every
322 value pair (k_x, k_y) by $\log\left(\sqrt{k_x^2 + k_y^2}\right)$ along the direction defined by the direction cosine of the
323 position vector (k_x, k_y) with the x-axis of the coordinate system and plotting $\log(P(k_x, k_y))$ on
324 the newly formed logarithmic grid. The central point in this case corresponds to the system
325 size, which imposes the smallest non-zero k . Figure 7b illustrates such a double log-plot of
326 sample M3/4, in which the power spectra are displayed as a 3D surface. Notice that the view
327 direction is along the k_x -axis. The slopes of the surface, which roughly describe an elliptical

328 cone clearly exhibit two linear branches and a distinct crossover region (L) marked by the
329 arrow in Figure 7b. Thus the 3D representation is consistent with the scaling behaviour found
330 from the analysis of the 1D signal.

331 For further analysis of the anisotropy we resample the 3D representation (Figure 7b)
332 with a 2D logarithmic binning (along k_x and k_y direction), to get a constant density of grid
333 points in double-logarithmic representation (Figure 8a). For this reason a fixed grid that
334 covers the 2D power spectra with a constant bin size (bs) in logarithmic space ($\log(bs) = 0.4$)
335 in the x and y direction is used to find all k_x, k_y -value pairs that fall into a certain bin, and the
336 mean of all power spectra that belong to these k_x, k_y -value pairs in this bin is then used to
337 define the binned power spectrum. This procedure allows analyzing the data with an equal
338 importance for the long and small scales, respectively. In addition this method smoothes the
339 data by removing the local fluctuations without an alteration of the overall geometry of the
340 3D representation, that is characterized by the two scaling regimes and the distinct crossover.

341 We use isopach/contour maps of the binned 2D power spectra to quantify the degree
342 of anisotropy. Isotropic signals should reveal concentric circular contour lines, which define
343 the same $\log(P(k_x, k_y))$ value. Concentric circular contour lines would imply that the crossover
344 length, which separates the self-affine scaling regimes for small and large scales are the same
345 in every direction. Figures 8 show that this is clearly not the case for tectonic stylolites (also
346 compare Figure 7a) where the contour lines reveal an elliptical shape (Figure 8a,b). This
347 shape is clearly different from the circular concentric contours found in bedding parallel
348 stylolites (compare e.g. to Figure 4 of Schmittbuhl et al., 2004). We use a least square
349 criterion to estimate the best fit ellipse of the individual contour lines. From the best fitted
350 ellipse, we calculate the aspect ratio of the principal axis (i.e. a/b ; where a and b are the semi-
351 major and semi-minor axis of the best fit ellipse) to get a quantitative measure of the
352 anisotropy of the 2D binned power spectra (Figure 8c). For the direction of the anisotropy we
353 utilize the angle Θ between the long axis (a) of the fitted ellipse and the x-direction of the

354 coordinate system (Figure 8d). For all investigated samples we recognized an increased
355 ellipticity toward the centre of the 2D power spectra but only a moderate or no significant
356 change in orientation of the asymmetry with respect to the position in the power spectra. Note
357 that in this representation (Figure 8a) high contour lines (small wavenumbers) correspond to
358 large physical length-scales whereas low contour lines (large wavenumbers) correspond to
359 small length-scales.

360 The fact that the large wavenumbers display an isotropic power spectrum i.e. aspect
361 ratio close to 1 (Figure 8c), whereas the small ones show an anisotropic one, is very consistent
362 with the result of the 1D data analysis (see previous section). This observation is also in
363 agreement with the physical interpretation of the mechanism of stylolite formation and
364 morphogenesis [Ebner *et al.*, 2009b; Koehn *et al.*, 2007; Renard *et al.*, 2004; Schmittbuhl *et*
365 *al.*, 2004]: At small scales (large wavenumbers), the balance between surface tension and
366 disorder is controlling the shape of stylolites. Both are a priori isotropic along the stylolite. In
367 contrast, the large scale morphologies (small wavenumbers) are normally physically
368 interpreted as resulting from a balance between the elastic field and the material disorder is
369 controlling the shape of the stylolites. The fact that an anisotropy is observed at large scales is
370 thus the signature of an in-plane anisotropy of the stress. Since stylolite teeth are normally
371 parallel to largest stress direction associated with σ_1 , this large scale anisotropy should be
372 associated to a difference between the two principal values of the in-plane stress components,
373 σ_2 and σ_3 .

374 The orientation of the long axis of the fitted ellipse relative to the vertical orientation
375 of the sample is shown in rose diagrams (Figure 9) for all samples. The long axes of the
376 contours of the power spectrum are associated with a shorter crossover-length L (i.e.
377 reciprocal to the wavenumber) between the large k isotropic scaling and the small k
378 anisotropic one (Fig 9j). We will see in the next sections that this can be interpreted as a
379 variation of the difference between the largest principal stress (normal to the stylolite plane)

380 and the two in-plane stress components. The principal stress associated with the direction of
381 the long axis should thus be the smallest one, i.e. σ_3 . Arrows show the orientation of a vertical
382 line projected onto the stylolite plane in its original outcrop orientation. From this
383 representation (Figure 9) it is evident that the vertical direction is roughly normal to the long
384 axis of the anisotropy for all samples except M4c/1 and M4c/3 which formed vertically
385 (compare chapter 2 for details) but were subsequently rotated into a shallow dipping (non
386 vertical) orientation plane due to folding (Figure 9h,i). They thus serve as a cross check to our
387 findings since the vertical direction in these samples does not coincide with the vertical
388 direction during stylolite formation and the anisotropy is therefore not normal to the present
389 vertical direction in these samples as for samples of the upright limb.

390 To estimate the crossover length (L) and thus get quantitative information on the
391 stresses during stylolite formation we again use the elliptical fit as a simplified representation
392 of the 2D Fourier transform of our data. We assume that the crossover is located at the
393 position of the biggest change in the local slope of the 2D Fourier transform (compare Figure
394 7b). We calculate the local slope s as the difference between the long and short axis (a, b) of
395 the best fit ellipse for succeeding $\log(P(k_x, k_y))$ -contours $s = (\Delta a + \Delta b) / 2$. A plot of the
396 $\log(P(k_x, k_y))$ -contours as a function of the local slope s is shown in Figure 10a. The crossover
397 is defined to lie at the minimum local slope in this representation and the crossover is
398 calculated from the principal axis of the best-fit ellipse at this minimum (Figure 10b). It can
399 be noticed that the maximum crossover-length coincides quite well with the vertical direction
400 (indicated by arrow in Figure 10b) this is in agreement with our previous observations that the
401 anisotropy of the power spectra is also oriented (normal) with respect to the sample vertical
402 orientation (compare Figure 9).

403 Before we discuss the orientation of the anisotropy and the determined crossover
404 length-scales in relation to the stress tensor that was present during stylolite growth, we want
405 to investigate the influence of tilted teeth on the scaling results.

406

407 **4.3. Synthetic data analysis**

408 It is important to prove that the large scale anisotropy we found in the investigated
409 samples is really related to the stress field during formation and thus exclude the influence of
410 other factors which might as well cause a scaling anisotropy. The second important
411 characteristic of tectonic stylolites, as stated in the introduction, is the occurrence of inclined
412 teeth i.e. slickolites. It is easy to imagine that the ridge and groove morphology of slickolites
413 with highly inclined teeth can causes a difference in the scaling parallel or transverse to these
414 elongated morphological features and thus an anisotropy. To systematically investigate the
415 influence of a tilt of the asperities or teeth we construct synthetic isotropic self-affine surfaces
416 and tilt the teeth around one arbitrary axis. Tilted or inclined asperities are a common feature
417 of slickolites [*Simon, 2007*] and it is commonly assumed that these structures formed when a
418 stylolite overprinted a pre-existing plane of anisotropy in the host-rock. In this case the
419 principal stresses are oriented oblique to the pressure solution surface, which has recently
420 been proven numerically by Koehn et al. (2007). Synthetic self-affine surfaces can be created
421 following the approaches found in the literature [*Meheust and Schmittbuhl, 2001; Turcotte,*
422 *1997*]. We follow the method described in Meheust and Schmittbuhl (2001) who construct
423 square white noise maps of size $n=512$. The self-affine correlation is then introduced by
424 multiplying the modulus of the 2D Fourier transform of the white noise by the modulus of the
425 wavenumber raised to the power of $-1-H$, where H is the roughness exponent. The self-affine
426 surface is obtained from the inverse Fourier transform. The synthetic surface shown in Figure
427 11a is constructed with a Hurst exponent of $H=0.5$ and its 2D Fourier transform has a true
428 isotropic self-affine behaviour (compare inset in Figure 11a). A pre-defined tilt of the
429 roughness is then attained from adding a linear trend along the x-axis of the map which
430 corresponds to a tilt angle α and a subsequent back-rotation around α i.e. multiplying the data
431 with a 3D rotation matrix of $-\alpha$. Various tilt angles ranging from 1-50° were realised from the

432 map shown in Figure 11a. To analyse single valued functions (with no overhangs) the tilted
433 surfaces are projected on a plane defined by the mean surface. The data were then analyzed as
434 described in the previous chapter (section 4.2). The degree (aspect ratio) and orientation
435 (slope) of the anisotropy is displayed in Figure 11b & c. It is evident that the original data-set
436 is isotropic with aspect-ratios for $\log(P(k_x, k_y))$ contours close to 1. With small tilt angles $\alpha <$
437 10° an anisotropy for the low $\log(P(k_x, k_y))$ contours and thus large wavenumbers and small
438 scales exists, which decreases with increasing α . In addition there is a general increase in the
439 anisotropy in all scales with tilt angles of $\alpha \geq 20^\circ$ (Figure 11b) whereas the orientation is
440 more and more aligned with the rotation/tilt axis (Figure 11c) with increasing tilt angle. The
441 topography of the synthetic surfaces do not exhibit a directional anisotropy but reveal a general
442 decrease of the average topography with increasing tilt angle from a $t \sim 0.22$ for the original
443 data down to $t \sim 0.09$ for a tilt angle of 50° .

444

445 **5. Discussion**

446 We have shown that the tectonic stylolites investigated in this study, i.e. stylolites
447 which form when the principal compressive stress direction is horizontal, differ
448 fundamentally from bedding parallel stylolites since they show anisotropic scaling relations.
449 Two self affine scaling regimes (with Hurst exponents of ~ 0.5 and ~ 1.1 for the small and
450 large scale, respectively), which are separated by a crossover-length at the millimeter scale
451 can be found in bedding parallel and tectonic stylolites. The crossover-length L scales
452 inversely with the formation stress $L \sim \sigma^{-2}$ for bedding parallel stylolites [Ebner *et al.*, 2009b].
453 The analytical solution of Schmittbuhl *et al.* [2004] relates the crossover length (L) to the
454 stress-field during stylolite formation. Their stress term is a product of mean and differential
455 stress and can be used to calculate the stress magnitudes in addition to the determination of
456 principal stress directions. The analytical solution shows that

457
$$L = \frac{\gamma E}{\beta \sigma_m \sigma_d}, \quad (3)$$

458 where E is the Young's Modulus, γ is the solid-fluid interfacial energy,
 459 $\beta = \nu(1 - 2\nu)/\pi$ is a dimensionless constant with ν the Poisson's ratio, σ_m and σ_d , are the mean
 460 and differential stresses respectively. Since for bedding parallel stylolites perfect confinement
 461 can be assumed (that is uniaxial strain or zero horizontal displacement) the stresses and thus
 462 the crossover length L is independent of the orientation within the stylolite surface (Figure
 463 1a). For a tectonic stylolite with a vertical stylolite plane the scenario is different (Figure 1b)
 464 and it can be assumed that the in-plane stresses are dissimilar. One in-plane principal stress
 465 component should be dependent on the amount of overburden and should be oriented
 466 vertically whereas the second stress component should have a horizontal orientation. Since
 467 the crossover-length L scales inversely with the product of mean and differential stress and
 468 the mean stress should be constant, variations of the crossover should reflect variations of the
 469 differential stress $|\sigma_1 - \sigma_{inplane}|$ [compare to Schmittbuhl et al., 2004]. Therefore the crossover-
 470 length has to increase from a minimum in the direction of the least principal stress σ_3 (x-axis
 471 in Figure 1b) and thus the direction of the largest differential stress $|\sigma_1 - \sigma_3|$ to a maximum in an
 472 in-plane orientation normal to this direction, which corresponds to the direction of the largest
 473 inplane stress σ_2 (the vertical direction in Figure 1b), and the smallest differential stress $|\sigma_1 -$
 474 $\sigma_2|$. In conclusion it can be assumed that the orientation of largest and smallest crossover-
 475 length coincide with the vertical and horizontal direction (i.e. $\sigma_{xx} < \sigma_{zz}$) respectively.

476 Indeed we found a scaling anisotropy in our data, which shifts the crossover-length
 477 accordingly (Figure 9). The 1D analysis (Figure 5) and the 2D data analysis (Figure 9 & 10)
 478 reveal that the long axis of the detected anisotropy is normal to the vertical direction with a
 479 crossover-length maximum in this direction implying that σ_2 has a vertical orientation. This
 480 observation holds for both investigated areas although there is a slight deviation of up to $\pm 10^\circ$
 481 for some samples. Only the samples (M4c/1, M4c/3 from the overturned fold limb) which

482 formed vertically but experienced a passive rotation subsequently to stylolite formation due to
483 folding (compare Figure 2b and Figure 9h,i) differ significantly. This can be explained by the
484 fact that the stylolite formation was prior to folding as can be concluded from the structural
485 relationships in the field data (Figure 2). Thus the present orientation of the samples in the
486 overturned fold limb does not coincide with the orientation during formation of the stylolites.

487 We noticed a small difference ($<10^\circ$) between the orientation of the stylolite teeth and
488 the pole of the mean stylolite plane for the samples from north eastern Spain. This is due to
489 the fact that the stylolites overprint a pre-existing joint set that is subnormal to the principal
490 shortening direction, which influenced stylolite formation as stated above. To investigate the
491 effect of the tilt of the stylolite teeth and its contribution to the observed scaling anisotropy we
492 used synthetic self-affine surfaces which were systematically tilted to get slickolite similar
493 structures as explained above (Figure 11). The effect of the tilt of the teeth with respect to the
494 mean plane of the stylolite can be characterized by (i) an anisotropy for the large
495 wavenumbers i.e. on the scale of individual teeth or asperities for small tilt angles ($<10^\circ$) and
496 (ii) a general homogeneous increase of the anisotropy for all scales with an increase of the tilt
497 angle for angles $>10^\circ$. This anisotropy caused by the imposed tilt of the asperities differs
498 significantly from the anisotropy of real stylolites. Therefore we conclude that the 3D
499 formation stress is the dominant force that influences the scaling anisotropy of the
500 investigated samples. However one has to note that tilted teeth imply that the principal stress
501 components are not necessarily oriented within the stylolite plane. Therefore only tectonic
502 stylolites with plane-perpendicular teeth should be used to recalculate principal stress
503 orientations and magnitudes.

504 The analytical solution [Schmittbuhl *et al.*, 2004] is only strictly valid for 2D stress
505 cases where the principal stresses parallel to the stylolite plane are invariant along the third
506 direction, which is truly the case for bedding parallel stylolites as discussed by Ebner *et al.*
507 [2009b]. But since a solution for the 3D case is not available we argue that the above equation

508 (Eq. 3) could serve as an ersatz, of a first approximation to calculate the order of magnitude
509 and the difference between the principal stresses for such tectonic stresses. We assume that
510 the crossover-length in a specific direction is mainly a function of the stresses in the plane
511 normal to the stylolite surface along the direction of investigation and that the out of plane
512 stresses are invariant. This would imply that the differential stresses for the vertical and
513 horizontal directions could be defined as $\sigma_{dv} = \sigma_{yy} - \sigma_{zz}$ and $\sigma_{dh} = \sigma_{yy} - \sigma_{xx}$ and Eq. 3 could
514 be solved if the depth of stylolite formation and the material properties during stylolite
515 formation are known. For the stylolites from the Swabian Alb with a vertical crossover of
516 0.95 mm and a horizontal crossover of 0.7 mm, assuming a Poisson's ratio of 0.25, a surface
517 free energy of calcite of 0.27 J/m^2 , a Young's Modulus of 14 GPa [Ebner et al., 2009b] and a
518 vertical stress component (σ_2) of 6 MPa (assuming a vertical load of 220 m of sediments with
519 a density of 2.7 g/cm^3 in agreement with sedimentological constraints) the tectonic stress
520 component (σ_1) is about 17.7 MPa and the horizontal in-plane stress (σ_3) component is 1.8
521 MPa. See appendix for details of the calculation. The theoretical stresses of stylolite
522 formation calculated here can not serve as realistic values since we unjustifiably borrow from
523 the analytical solution for the isotropic case but should give a first order estimate under the
524 limiting assumptions stated above. Nevertheless we would expect stresses during tectonic
525 stylolite formation to be close to the compressive lithospheric strength, i.e. $\sigma_1 - \sigma_3 \sim 14 \text{ MPa}$
526 [Banda and Cloetingh, 1992] but much smaller than uniaxial compressive strength of
527 laboratory measurements for limestones, which are in the range of $\sim 50\text{-}200 \text{ MPa}$ [Pollard and
528 Fletcher, 2005]. Utilizing the solution given in the appendix the resulting stress magnitudes
529 are surprisingly close to expected values.

530 For our samples in Spain we do not calculate the stresses because the principal stresses
531 are quite likely not aligned with the stylolite plane as discussed above. We argue that even if
532 it would be possible to calculate the stresses for tectonic stylolites in a fold and thrust belt like
533 in northeastern Spain the stresses deduced from stylolites might be completely different form

534 that of the folding event. The main reason is that stylolites probably form rather quick, in the
535 order of hundreds of years [Schmittbuhl *et al.*, 2004]. This would allow several generations of
536 stylolites to form (revealing different finite orientation) during a single folding event the
537 analysis of a single set of stylolites would thus result in a snapshot from the geologic history
538 not necessarily revealing the full picture. Even if the stylolites can be attributed to the same
539 kinematic framework as the folding event both most likely have a rather diverse history in
540 terms of stress.

541

542 **6. Conclusions**

543 Vertical tectonic stylolites investigated in this study show a 1D scaling invariance that
544 resembles those of bedding parallel stylolites investigated in previous studies [Ebner *et al.*,
545 2009b; Renard *et al.*, 2004; Schmittbuhl *et al.*, 2004]. They have a self-affine scaling
546 invariance, which is characterized by a Hurst exponent of 1.1 for long and 0.5 for short scales
547 and a distinct crossover-length at the millimeter scale that separates these two scaling
548 regimes.

549 High resolution laser profilometry of tectonic stylolites provides quantitative 3D
550 information of these pressure solution surfaces that enables a 2D analysis of the surface
551 morphology. We demonstrate that our samples of tectonic stylolites have an anisotropic
552 scaling that is not independent of the orientation of the investigated section within the plane
553 of the stylolite. This anisotropy's main characteristic is a systematic shift of the crossover
554 length that separates the scaling regimes. The presented analysis also confirms that the
555 anisotropy observed in our vertical samples is oriented with respect to the horizontal and
556 vertical direction and thus coincides with the principal stress directions within the stylolite
557 plane for vertical stylolites e.g. σ_2 & σ_3 as depicted in Figure 2b. The long axis of the
558 anisotropy and thus the smallest crossover length consistently coincides with the horizontal
559 direction in the stylolite plane, whereas the largest crossover-length is found in a vertical

560 section. This observation is consistent with the fact that the horizontal in-plane stress is
 561 generally smaller than the vertical in-plane stress, which should be the case for tectonic
 562 stylolites (Figure 1b). They are also both smaller than the normal stress orientated
 563 perpendicular to the stylolite plane, which should be oriented horizontally. Therefore the
 564 crossover-length should be smaller in a horizontal section than in a vertical section (Eq. 3)
 565 using analytical considerations [Schmittbuhl et al., 2004].

566 In addition we studied the influence of inclined teeth and asperities on the scaling
 567 behavior of stylolites. Using synthetic ‘slickolites’ with various tilt angles we found that the
 568 evolving anisotropy is negligible and clearly different from the anisotropy we observed in the
 569 investigated samples. We thus conclude that the scaling anisotropy of the investigated vertical
 570 tectonic stylolites can be related to the 3D formation stress.

571

572 **7. Appendix: Stress Calculation**

573 Part I

574 In this appendix we will show how the tectonic stress (σ_1) and the smaller in-plane stress
 575 component (σ_3) can be calculated if the vertical stress component can be approximated using
 576 vertical loading conditions. According to equation (4) the vertical and horizontal crossovers
 577 (L_v and L_h) can be calculated by [Schmittbuhl et al., 2004]

578

$$579 \quad L_v = \frac{\gamma E}{\beta} \frac{1}{\sigma_m \sigma_{dv}} \quad L_h = \frac{\gamma E}{\beta} \frac{1}{\sigma_m \sigma_{dh}} \quad (\text{A1})$$

580 where E is the Young’s Modulus, γ is the solid-fluid interfacial energy, $\beta = \nu(1-2\nu)/\pi$ is a
 581 dimensionless constant with ν the Poisson’s ratio, σ_m and $\sigma_{dv/h}$, are the mean and differential
 582 stresses respectively. Since the mean stress is the same for both directions we can reformulate
 583 equation A1 to

$$584 \quad \sigma_m = \frac{\gamma E}{\beta} \frac{1}{L_v \sigma_{dv}}, \quad \sigma_m = \frac{\gamma E}{\beta} \frac{1}{L_h \sigma_{dh}} \quad (\text{A2})$$

585 and join both equations so that

$$586 \quad L_v \sigma_{dv} = L_h \sigma_{dh}. \quad (\text{A3})$$

587 If we now define the differential stresses using the main principal stress components with $\sigma_1 =$

588 σ_{yy} ; i.e. acting normal to the stylolite plane; $\sigma_2 = \sigma_{zz}$; i.e. the vertical in plane stress component

589 and $\sigma_3 = \sigma_{xx}$; i.e. the horizontal in plane stress component (compare Figure 1b); as

590 $\sigma_{dv} = \sigma_{yy} - \sigma_{zz}$ and $\sigma_{dh} = \sigma_{yy} - \sigma_{xx}$ equation A3 becomes

$$591 \quad \frac{L_h}{L_v} = \frac{\sigma_{yy} - \sigma_{zz}}{\sigma_{yy} - \sigma_{xx}} \quad (\text{A4})$$

592 and solving for the xx component

$$593 \quad \sigma_{yy} - \sigma_{xx} = \frac{L_v}{L_h} (\sigma_{yy} - \sigma_{zz}),$$

$$594 \quad \sigma_{xx} = \sigma_{yy} - \frac{L_v}{L_h} (\sigma_{yy} - \sigma_{zz}) = \sigma_{yy} - \frac{L_v}{L_h} \sigma_{yy} + \frac{L_v}{L_h} \sigma_{zz} \quad (\text{A5}).$$

595 Part II

596 For simplification we substitute all material parameters of Equation 4 which are assumed to

597 be constant, according to

$$598 \quad a = \frac{\gamma E}{\beta}.$$

599 Then we use equation 4 for the horizontal cross-over

$$600 \quad L_h = a \frac{1}{\sigma_m \sigma_{dh}}$$

601 or

602

$$603 \quad \sigma_m \sigma_{dh} = \frac{\sigma_{xx} + \sigma_{yy} + \sigma_{zz}}{3} (\sigma_{yy} - \sigma_{xx}) = \frac{a}{L_h}$$

604 and

$$605 \quad (\sigma_{xx} + \sigma_{yy} + \sigma_{zz})(\sigma_{yy} - \sigma_{xx}) = 3 \frac{a}{L_h} \quad (\text{A6})$$

606 Now we include equation A5 into equation A6 and solve for σ_{yy}

$$607 \left(2\sigma_{yy} + \sigma_{zz} - \sigma_{yy} \frac{L_v}{L_h} + \sigma_{zz} \frac{L_v}{L_h} \right) \left(\sigma_{yy} \frac{L_v}{L_h} - \sigma_{zz} \frac{L_v}{L_h} \right) = 3 \frac{a}{L_h} \quad (\text{A7})$$

608 and multiplying the components gives

$$609 2\sigma_{yy}^2 \frac{L_v}{L_h} - \sigma_{yy}^2 \left(\frac{L_v}{L_h} \right)^2 + 2\sigma_{yy}\sigma_{zz} \left(\frac{L_v}{L_h} \right)^2 - \sigma_{yy}\sigma_{zz} \frac{L_v}{L_h} - \sigma_{zz}^2 \frac{L_v}{L_h} - \sigma_{zz}^2 \left(\frac{L_v}{L_h} \right)^2 - 3 \frac{a}{L_h} = 0. \quad (\text{A8})$$

610 Rearranging equation A8 in order to solve a binomial formula gives

$$611 \sigma_{yy}^2 + \sigma_{yy} \frac{2\sigma_{zz} \left(\frac{L_v}{L_h} \right)^2 - \sigma_{zz} \frac{L_v}{L_h} - \sigma_{zz}^2 \frac{L_v}{L_h} - \sigma_{zz}^2 \left(\frac{L_v}{L_h} \right)^2 - 3 \frac{a}{L_h}}{2 \frac{L_v}{L_h} - \left(\frac{L_v}{L_h} \right)^2} = 0 \quad (\text{A9})$$

612 and the solution of the binomial formula is then

$$613 \sigma_{yy,1,2} = -0.5 \frac{2\sigma_{zz} \left(\frac{L_v}{L_h} \right)^2 - \sigma_{zz} \frac{L_v}{L_h}}{2 \frac{L_v}{L_h} - \left(\frac{L_v}{L_h} \right)^2} \pm \sqrt{0.25 \left(\frac{2\sigma_{zz} \left(\frac{L_v}{L_h} \right)^2 - \sigma_{zz} \frac{L_v}{L_h}}{2 \frac{L_v}{L_h} - \left(\frac{L_v}{L_h} \right)^2} \right)^2 - \left(\frac{-\sigma_{zz}^2 \frac{L_v}{L_h} - \sigma_{zz}^2 \left(\frac{L_v}{L_h} \right)^2 - 3 \frac{a}{L_h}}{2 \frac{L_v}{L_h} - \left(\frac{L_v}{L_h} \right)^2} \right)}.$$

614 (A10).

615 σ_{xx} can be derived from equation A5.

616

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624

625 **Reference list:**

626

- 627 Angheluta, L., et al. (2008), Stress-Driven Phase Transformation and the Roughening of
628 Solid-Solid Interfaces, *Physical Review Letters*, 100(9), 096105.
- 629 Banda, E., and S. Cloetingh (1992), Europe's lithosphere - Mechanical Structure, in *The*
630 *European Geotraverse*, edited by D. Blundell, et al., Cambridge University Press.
- 631 Barabasi, A. L., and H. E. Stanley (1995), *Fractal concepts in surface growth*, 366 pp.,
632 Cambridge University Press.
- 633 Bonnetier, E., et al. (2009), Does roughening of rock-fluid-rock interfaces emerge from a
634 stress-induced instability?, *The European Physical Journal B - Condensed Matter and*
635 *Complex Systems*, 67(1), 121-131.
- 636 Bretz, J. H. (1940), Solution cavities in the Joliet limestone of northeastern Illinois, *J Geol*,
637 48(4), 337-384.
- 638 Brouste, A., et al. (2007), Variety of stylolites' morphologies and statistical characterization of
639 the amount of heterogeneities in the rock, *Journal of Structural Geology*, 29(3), 422-434.
- 640 Buxton, T. M., and D. F. Sibley (1981), Pressure Solution Features in a Shallow Buried
641 Limestone, *Journal of Sedimentary Petrology*, 51(1), 19-26.
- 642 Candela, T., et al. (2009), Characterization of fault roughness at various scales: Implications
643 of three-dimensional high resolution topography measurements, *Pure and Applied*
644 *Geophysics*.
- 645 Capote, R., et al. (2002), The Alpine system north of the Betic Cordillera, in *The geology of*
646 *Spain*, edited by W. Gibbons and T. Moreno, p. 632, Geological Society of London London.
- 647 Casas, A. M., et al. (2000), Syn-tectonic sedimentation and thrust-and-fold kinematics at the
648 intra-mountain Montalban Basin (northern Iberian Chain, Spain), *Geodinamica Acta*, 13(1),
649 1-17.
- 650 Cooley, J. W., and J. W. Tukey (1965), An Algorithm for Machine Calculation of Complex
651 Fourier Series, *Mathematics of Computation*, 19(90), 297-&.
- 652 Drummond, C. N., and D. N. Sexton (1998), Fractal structure of stylolites, *Journal of*
653 *Sedimentary Research*, 68(1), 8-10.
- 654 Dunnington, H. V. (1954), Stylolite development post-dates rock induration, *Journal of*
655 *sedimentary Petrology*, 24(1), 27-49.
- 656 Ebner, M., et al. (2009a), The influence of rock heterogeneity on the scaling properties of
657 simulated and natural stylolites, *Journal of Structural Geology*, 31(1), 72-82.
- 658 Ebner, M., et al. (2009b), Stress sensitivity of stylolite morphology, *Earth and Planetary*
659 *Science Letters*(277), 394-398.
- 660 Etzold, A., et al. (1996), Schwäbische Alb - Stratigraphie, Tektonik, Vulkanismus,
661 Karsthydrogeologie, *Z. geol. Wiss.*, 24(1/2), 175-215.
- 662 Geyer, O. F., and M. P. Gwinner (1991), *Geologie von Baden-Württemberg*, 4th ed., 482 pp.,
663 Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.
- 664 Goshtasby, A. (1986), Piecewise Linear Mapping Functions for Image Registration, *Pattern*
665 *Recognition*, 19(6), 459-466.
- 666 Goshtasby, A. (1988), Image Registration by Local Approximation Methods, *Image and*
667 *Vision Computing*, 6(4), 255-261.
- 668 Gratier, J. P., et al. (2005), Experimental microstylolites in quartz and modeled application to
669 natural stylolitic structures, *Journal of Structural Geology*, 27(1), 89-100.
- 670 Guzzetta, G. (1984), Kinematics of Stylolite Formation and Physics of the Pressure-Solution
671 Process, *Tectonophysics*, 101(3-4), 383-394.

672 Heald, M. T. (1955), Stylolites in Sandstones, *J Geol*, 63(2), 101-114.
673 Karcz, Z., and C. H. Scholz (2003), The fractal geometry of some stylolites from the Calcare
674 Massiccio Formation, Italy, *Journal of Structural Geology*, 25(8), 1301-1316.
675 Kley, J., and T. Voigt (2008), Late Cretaceous intraplate thrusting in central Europe: Effect of
676 Africa-Iberia-Europe convergence, not Alpine collision, *Geology*, 36(11), 839-842.
677 Koehn, D., et al. (2007), Growth of stylolite teeth patterns depending on normal stress and
678 finite compaction, *Earth and Planetary Science Letters*, 257(3-4), 582-595.
679 Liesa, C. L., and J. L. Simón (2009), Evolution of intraplate stress fields under multiple
680 remote compressions: The case of the Iberian Chain (NE Spain), *Tectonophysics*, 474(1-2),
681 144-159.
682 Meheust, Y., and J. Schmittbuhl (2001), Flow Enhancement of a Rough Fracture, *Geophys.*
683 *Res. Lett.*, 27.
684 Park, W. C., and E. H. Schot (1968), Stylolites: their nature and origin, *Journal of*
685 *sedimentary Petrology*, 38(1), 175-191.
686 Pollard, D. D., and R. C. Fletcher (2005), *Fundamentals of structural geology*, 500 pp.,
687 Cambridge University Press, Cambridge, United Kingdom.
688 Press, W. H., et al. (2007), *Numerical Recipes. The Art of Scientific Computing*, 3 ed., 1256
689 pp., Cambridge University Press.
690 Railsback, B. L., and L. M. Andrews (1995), Tectonic stylolites in the 'undeformed'
691 Cumberland Plateau of southern Tennessee, *Journal of Structural Geology*, 17(6), 911-915.
692 Railsback, L. B. (1993), Lithologic Controls on Morphology of Pressure-Dissolution Surfaces
693 (Stylolites and Dissolution Seams) in Paleozoic Carbonate Rocks from the Mideastern
694 United-States, *Journal of Sedimentary Research*, 63(3), 513-522.
695 Renard, F., et al. (2004), Three-dimensional roughness of stylolites in limestones, *J Geophys*
696 *Res-Sol Ea*, 109(B3), -.
697 Renard, F., et al. (2006), High resolution 3D laser scanner measurements of a strike-slip fault
698 quantify its morphological anisotropy at all scales, *Geophysical Research Letters*, 33(4), -.
699 Rutter, E. H. (1983), Pressure solution in nature, theory and experiment., *Journal of the*
700 *Geological Society of London*, 140(5), 725-740.
701 Schmittbuhl, J., et al. (1995), Reliability of Self-Affine Measurements, *Physical Review E*,
702 51(1), 131-147.
703 Schmittbuhl, J., et al. (2004), Roughness of stylolites: Implications of 3D high resolution
704 topography measurements, *Physical Review Letters*, 93(23), -.
705 Schmittbuhl, J., et al. (2008), Fracture morphology and viscous transport, *International*
706 *Journal of Rock Mechanics and Mining Sciences*, 45(3), 422-430.
707 Simon, J. L. (2007), Analysis of solution lineations in pebbles: Kinematical vs. dynamical
708 approaches, *Tectonophysics*, 445(3-4), 337-352.
709 Simonsen, I., et al. (2000), Wave scattering from self-affine surfaces, *Physical Review E*,
710 61(5), 5914.
711 Stockdale, P. B. (1922), Stylolites: their nature and origin, *Indiana University Studies*, 9, 1-
712 97.
713 Tada, R., and R. Siever (1989), Pressure Solution during Diagenesis, *Annual Review of Earth*
714 *and Planetary Sciences*, 17, 89-118.
715 Turcotte, D., L. (1997), *Fractals and Chaos in Geology and Geophysics*, 2 ed., 398 pp.,
716 Cambridge University Press, Cambridge.

717
718 **Figure captions:**

719
720 **Figure 1:** Schematic drawing of the formation stress state for (a) bedding parallel and (b)
721 tectonic stylolites. The largest compressive stress direction (σ_1) is indicated by a white arrow.

722 Below the sketch map an idealized graph of the in-plane differential stress is plotted as a
723 function of the orientation within the stylolite plane. For bedding parallel stylolites **(a)** the
724 horizontal normal stresses are equal and thus the differential stress is equal in every direction.
725 For tectonic stylolites **(b)** the in-plane normal stresses are dissimilar and σ_{zz} is generally larger
726 than σ_{xx} . Thus the in-plane differential stress scales inversely with the magnitudes of the σ_{xx}
727 and σ_{zz} directions having a maximum along the x-axis.

728

729 **Figure 2:** Lower hemispheric equal area projection (Schmidt's net) of the field data and
730 schematic cross-sections of the investigated outcrops. **(a)** The Swabian Alb of southern
731 Germany (n=22). Right panel shows the flat lying Jurassic strata with vertical stylolites
732 limited to individual beds **(b)** Iberian Chain of north-eastern Spain (n=32). Right panel shows
733 a cross-section of NE plunging fold and the position of *set a* and *set b* within the fold. All
734 samples are taken from well bedded Jurassic strata. In the overlying massif Jurassic
735 limestones (vertical stripes) and conglomerates (circles) no stylolites were found. Notice that
736 in **(a)** only the poles to the stylolite planes are displayed since the shortening direction is
737 normal to that plane. In panel **(b)** two populations are shown which correspond to the two
738 investigated fold limbs. Poles to planes (circles) diverge slightly from the orientation of the
739 long axis of the teeth (triangle); See text for detailed explanation.

740

741 **Figure 3:** Oblique view of the 3D morphology of the surface of an opened stylolite (sample
742 M4/4) reconstructed from optical profilometry. A linear trend is removed from the raw data
743 (compare Figure 4 for details).

744

745 **Figure 4:** Greyscale maps of sample M4/3 where **(a)** shows the raw data from profilometry
746 (notice a general trend from the top left to bottom right); **(b)** detrended data i.e. linear trend is
747 removed and mean height is set to be zero; **(c)** detrended data which is modified with a

748 Hanning window technique where the data is forced to taper off to zero at the boundaries (for
749 explanation see text). Light colours correspond to peaks and ridges and dark colours represent
750 local depressions.

751

752 **Figure 5:** 1D data-analysis of sample M4/3; **(a)** shows the averaged Power spectra $P(k)$ (solid
753 line) and the respective binned spectra (circles) plotted as a function of the wavenumber along
754 the x and the y direction of the measured map. The inset in (a) again shows the power spectra
755 for both directions but the x direction is now normalized with respect to the y direction
756 $P_x(k)/P_y(1\text{mm}^{-1})$. This yields a collapse of the large k-values (small scales), notice that for the
757 small k-values (large scales) the scaling functions deviate considerably **(b)** non-linear fit of
758 the binned spectra for both directions used to estimate the crossover length L (triangle). Along
759 the x-direction the crossover-length is larger ($L=1.22$) than along the y-direction ($L=0.62$).
760 The slope of the branches of the non-linear model corresponds to Hurst exponents of 1.1 and
761 0.5 for small and large scales, respectively.

762

763 **Figure 6:** 1D analysis of the scaling prefactor i.e. the topothesy of tectonic stylolites. **(a)** A
764 loglog plot of the correlation function $C(\Delta x)$ of a 1D slice of sample M4/3 oriented parallel to
765 the x direction of the analyzed surface with the nonlinear fit (compare text for details) and the
766 topotheses t_s and t_l for small and large scale sub-branches. The topothesy is constructed from
767 the intersection of the linear sub-branches with the 1/1 line. **(b)** The topotheses t_s and t_l of
768 sample M4/3 plotted as a function of θ i.e. the counter clockwise angle from the x-direction of
769 the map. Note that the correlation functions are averaged over 5° intervals. Arrow indicates
770 the vertical direction projected onto the stylolite plane. Note that only the t_s shows a clear
771 correlation with the sample orientation. **(c)** The small scale topothesy t_s for all samples plotted
772 as a function of θ .

773

774 **Figure 7:** 2D data-analysis of sample M4/3; **(a)** 2D Fourier transform plotted on a regular
775 grid as a function of k_x and k_y which range from $-((n/2)\Delta x)^{-1}$ to $(n/2)\Delta x)^{-1}$ where n is the
776 number of measurement points in one direction of the map and Δx is the step size. (notice that
777 the zero frequency component lies in the centre of the map). A clear anisotropy of the data
778 can be observed sub-parallel to the k_y -axis (vertical axis). To investigate the power law
779 scaling exhibited by the 1D analysis the 2D Fourier transform is converted to a double log-
780 space where $\log(k_x, k_y)$ is plotted as a function of the logarithm of the power spectra **(b)**; the
781 2D power spectra are plotted as a surface whose height corresponds to $\log(P(k_x, k_y))$. The 3D
782 surface is viewed along the k_x -direction and the arrow indicates the crossover-length L , which
783 separates the two scaling regimes i.e. the two linear subparts of the slope of the cone.

784

785 **Figure 8:** Quantification of the 2D scaling anisotropy of sample M4/3; **(a)** oblique 3D view
786 of the binned 2D power spectra (grey mesh) with an overlay of coloured contour lines of
787 constant $\log(P(k_x, k_y))$ -values. **(b)** Map view of the contours calculated from the conic 2D
788 power spectra. These contours were utilized to calculate best-fitting ellipses using a least
789 squares approach; **(c)** Aspect ratio (a/b) of the fitted ellipse for every $\log(P(k_x, k_y))$ -contour.
790 An increasing aspect ratio towards the centre of the map is characteristic for all samples
791 investigated. **(d)** Slope (i.e. the counter clockwise angle from the x-direction of the measured
792 map) of the long axis of the fitted ellipse plotted for the contour intervals.

793

794 **Figure 9:** Rose diagrams of all samples i.e. a histogram with a constant bin size of 10°
795 plotting the relative orientation of the long axis of the fitted ellipse to the vertical direction of
796 each sample. Arrow in each panel shows the intersection of the vertical direction of the
797 oriented sample with the mean stylolite plane. **(a)** sample Sa6/1a, **(b)** sample Sa6/1b, **(c)**
798 sample Sa9/2, **(d)** sample M4/1, **(e)** sample M4/2, **(f)** sample M4/3, **(g)** sample M4/4, **(h)**
799 sample M4c/1, **(i)** sample M4c/3; Notice that for all samples the long axis and thus the

800 direction with the smallest crossover length is roughly normal to the vertical direction (except
801 for h & i; for explanation see text). This direction corresponds typically to the largest
802 differential stress, which is also the smallest in-plane stress (v and h correspond to the vertical
803 and horizontal directions, respectively). **(j)** Schematic drawing of the relationship between the
804 wavenumber contour [mm^{-1}] (compare Figure 8), the crossover-length L [mm], the principal
805 in-plane stresses and the sample orientation i.e. horizontal and vertical direction. Refer to text
806 for detailed explanation.

807

808 **Figure 10:** Crossover length from the contour data of the maps for sample M4/3 and Sa6/1a.
809 **(a)** Slope of the 2D power spectra calculated as the mean difference between the principal
810 axis of the fitted ellipse (a,b). The biggest change in slope (arrow) is assumed to be the
811 contour at which the crossover is located. **(b)** The crossover-length plotted as a function of the
812 counter clockwise angle from the x-direction of the measured map. The vertical direction in
813 the stylolite plane is indicated for both samples and roughly corresponds to the largest
814 crossover-length i.e. the smallest differential stress as shown in Figure 1.

815

816 **Figure 11:** Greyscale map **(a)** of a synthetic self affine square surface with a side-length of
817 512 and a Hurst exponent of 0.5. Inset displays a 2D Fourier transform of that map, which
818 clearly exhibits isotropy with respect to its centre, similar to bedding parallel stylolites. This
819 dataset is then utilized to construct slickolites i.e. stylolites with oblique teeth and asperities
820 (see text), with various tilt angles (e.g. 10° correspond to oblique asperities that are rotated
821 10° counter clockwise around the x-direction with respect to the mean plane of the synthetic
822 surface). **(b)** Aspect ratio of elliptical fit of synthetic data set. For small tilt angles an
823 anisotropy on small scales (i.e. large wavenumbers and low $\log(P(k_x, k_y))$ -contours) can be
824 observed. For large tilt angles a general increase of the aspect ratio over all scales can be
825 found. **(c)** Orientation of the long axis of the fitted ellipse (compare Figure 8d). Notice an

826 increasing alignment of the long axis of the fitted ellipse towards higher $\log(P(k_x, k_y))$ -
827 contours with increasing tilt angles.

828

829