

# Vertical stress history and paleoburial in foreland basins unravelled by stylolite roughness paleopiezometry: Insights from bedding-parallel stylolites in the Bighorn Basin, Wyoming, USA.

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 unravelled by stylolite roughness paleopiezometry: insights
 from bedding-parallel stylolites in the Bighorn Basin,
 Wyoming, USA.

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- 13

14 ABSTRACT

15 We apply the stylolite roughness inversion technique on sedimentary, bedding-parallel stylolites hosted in the Paleozoic carbonates of the Bighorn and Madison formations cropping 16 out in the Bighorn Basin, Wyoming, USA. The inversion technique applied to bedding-parallel 17 stylolites allows determination of the absolute magnitude of the vertical stress experienced 18 19 at the time dissolution stops along the pressure-solution planes. At the basin scale, reconstructed vertical stress magnitudes range from 19±2 MPa to 35±4 MPa in the Bighorn 20 21 Fm, and from 12±2 MPa to 37±4 MPa in the Madison Fm. Once converted into depth and compared with up-to-date basin models of burial and contractional history, the dataset 22 highlights that bedding-parallel stylolites accommodated compaction from ca. 220 Ma until 23 ca. 90 Ma. We deduce that this correspond to the timing at which the maximum horizontal 24 stress related to the Sevier contraction and related stress build-up became higher than the 25 vertical stress related to burial. This study is key to illustrate how stylolites can be used to 26 consistently access paleoburial and to unravel both stress evolution and timing in foreland 27 settings, and indicates that pressure-solution remains active throughout the carbonate 28 deposition history. 29

# 30 INTRODUCTION

Reconstructing variations in burial depth of strata is a challenging but vital task to 31 constrain depositional, thermal and tectonic histories of sedimentary basins (Beaudoin and 32 Lacombe, 2018; Guidish et al., 1985; Yalcin et al., 1997). A simple but inaccurate approach 33 34 consists in reconstructing the thickness of the overlying sedimentary column by estimating the amount of rock compaction and thickness of eroded strata. Lacombe et al. (2009) used 35 36 paleopiezometric inversion of calcite twins to estimate maximum burial depth under the 37 debatable assumptions that the maximum differential stress related to layer-parallel shortening prevailed at the maximum burial depth and that the upper crust is at frictional 38 39 stress equilibrium (Lacombe, 2007). Beke et al. (2019) recently proposed a relationship between deformation depth and the typology of deformation bands, however authors states 40 that the typology itself depends on the degree of fluid-rock interactions, limiting this 41 42 method's applicability. More common methods that are used to assess paleoburial depth are 43 low-temperature thermochronology and vitrinite reflectance (e.g., Naeser and McCulloh, 2012; Roure et al., 2010; Tissot et al., 1987; Yalcin et al., 1997). Fluid inclusion 44 45 microthermometry on microveins combined with burial models may also help assess depth and timing of burial of strata (e.g., Anders et al., 2014; Becker et al., 2010; English et al., 2003; 46 47 Fall et al., 2012). The three approaches above rely upon assumptions on the past geothermal 48 gradient. In areas possibly affected by uplift and erosion, like fold-and-thrust belts and 49 foreland basins, it is strongly challenging to solve for both temperature and burial for each time interval, so that there is usually a lack of control on paleo-burial estimates. A tool that 50 provides constraints on the paleo-burial of strata without any assumption on the past 51 52 geothermal gradient would thus be a significant step forward (Beaudoin and Lacombe, 2018), 53 allowing for cross-checking between approaches, and hence leading to a gain in accuracy.

54 Stylolites, that are represented by serrated surfaces mainly in carbonate rocks, are common features in sedimentary basins (Andrews and Railsback, 1997; Laronne Ben-Itzhak 55 et al., 2014; Stockdale, 1922; Tavani et al., 2010; Toussaint et al., 2018). Stylolites develop by 56 chemical dissolution under stress (Alvarez et al., 1978; Fletcher and Pollard, 1981; Koehn et 57 al., 2007; Toussaint et al., 2018) and are an early mechanism of accommodation of 58 59 deformation, either related to tectonic shortening or to burial. In the latter case, stylolites are 60 primarily seen as a markers of the amount of compaction (Bathurst, 1987; Koehn et al., 2007; 61 Railsback, 1993). In order to enhance our reservoir prediction capability, for both industrial and societal sake, there is a growing interest to understand how stylolites develop, how they 62 63 impact fluid flow, and how they affect reservoir properties of carbonate rocks (e.g. Baud et al., 2016; Bruna et al., 2019; Heap et al., 2018). Beyond this classical approach, it was 64 proposed recently that stylolite roughness can be reliably used as a paleopiezometer (Ebner 65 66 et al., 2009; Ebner et al., 2010; Rolland et al., 2014; Rolland et al., 2012; Schmittbuhl et al., 67 2004; Toussaint et al., 2018), paving the way to a temperature independent marker of the burial history of carbonate rocks. Indeed, the roughness along a stylolite track, i.e., the 68 69 evolution of the difference in height of nearest neighbors, holds self-affine properties that 70 can be linked to the magnitude of applied stress prevailing in the strata at the time the final roughness of the stylolite fossilized. Various methods were tested to validate the theory of 71 72 the technique (Schmittbuhl et al., 2004; Ebner et al., 2009; Ebner et al., 2010), but cases of 73 application to geological problems remain limited (Beaudoin et al., 2019a; Beaudoin et al., 2016; Bertotti et al., 2017; Ebner et al., 2009; Rolland et al., 2014), especially in basins where 74 contractional tectonics occurred. 75

We take advantage of the well-documented burial history and sedimentological and
 structural framework of the Bighorn Basin (Wyoming, USA) to reconstruct the evolution of

vertical stress magnitudes in sedimentary strata that underwent burial followed by the Sevier-78 Laramide contractional deformation (Amrouch et al., 2010a; Amrouch et al., 2010b; Barbier 79 et al., 2012; Beaudoin et al., 2011; Beaudoin et al., 2012; Bellahsen et al., 2006; Carrapa et al., 80 2019; Craddock and van der Pluijm, 1999; DeCelles et al., 1991; Erslev and Koenig, 2009; 81 Lovely et al., 2010; Neely and Erslev, 2009; Weil and Yonkee, 2012; Yonkee and Weil, 2015). 82 In this context where absolute dating of calcite cements of tectonic veins constrains the 83 84 timing of contractional tectonics in the sedimentary cover (Beaudoin et al., 2018; Beaudoin 85 et al., 2019b), we aim at illustrating the potential of the stylolite roughness inversion by addressing the questions of how deep, and for how long, pressure-solution along a population 86 87 of sedimentary, bedding-parallel stylolites remained an efficient mechanism to accommodate vertical compaction and affected the evolution of the reservoir properties. We also question 88 89 whether the tectonic history impacts such a record, and we reconstruct the absolute 90 magnitude of the maximum vertical stress, hence of the maximum paleodepth of burial within 91 the basin at the onset of layer-parallel shortening related to the far-field Sevier contraction.

92

# 93 GEOLOGICAL SETTING AND SAMPLING STRATEGY

The Bighorn Basin (Figure 1) formed in response to the long-lasting subduction of the 94 Farallon plate, first as part of a broad marine foreland basin related to the thin-skinned Sevier 95 96 orogeny during Cretaceous to early Paleocene times, and subsequently as an endorheic basin 97 within the thick-skinned Laramide province that developed craton-wards by Late Cretaceous to Paleogene times (Yonkee and Weil, 2015). The sedimentary history of the Bighorn Basin 98 (BHB) has been extensively studied in order to (1) reconstruct the geodynamic evolution of 99 100 the North American plate, and (2) understand its georesources potential, e.g. hydrocarbons 101 and geothermal potential. The depocenter of the basin stacked up to 5.5 km of sediments 102 (DeCelles, 2004; Fox and Dolton, 1996; May et al., 2013) since Cambrian times, mainly 103 consisting of shales and sandstones since the Mesozoic (Figure 1a). In this study, we are 104 focusing on the Paleozoic section of the BHB, that crops out in basement-cored folds on the 105 eastern and western edges of the basin (Figure 1b, c).



106

107 Figure 1 : a ) stratigraphic column of the Bighorn Basin after Durdella (2001) and Neely and Erslev (2009), with reported thickness based on a well log located ~25 km west from Cody and located on the 108 109 map b) by a black square. b) Simplified geological map of the Bighorn Basin, where are reported the 110 locations of sampling sites as GPS numbers, see table 1 for correspondence. Location of the area 111 reported in the insert as a black frame labelled B. The red line on the map and the dotted red line in 112 the insert locate the cross-section c). c) NE-SW cross section across the eastern Sevier Belt and Western Laramide belt (modified after Marshak et al. (2000); Lacombe and Bellahsen (2016)). Camb -113 114 Cambrian, Or – Ordovician, Dev – Devonian, Carb – Carboniferous, Perm – Permian, T - Triassic, J – 115 Jurassic, <K – Lower Cretaceous, > K – Upper Cretaceous, RMA – Rattlesnake Mountain Anticline, LSMA 116 – Little Sheep Mountain Anticline, SMA – Sheep Mountain Anticline, BM – Bighorn Mountains, GRB – 117 Green River Basin, WRB – Wind River Basin, BHB – Bighorn Basin.

118

119 The Paleozoic succession of the BHB consists of the Cambrian sandstones, marls and

120 shales alternation of the Flathead, Gros Ventre and Gallatin Formations, the Ordovician

massive dolomites of the Bighorn Formation, the Devonian sandstones of the Three Forks 121 and Jefferson Formations, the Mississippian dolostones and limestones of the Madison 122 123 Formation overlain by the Pennsylvanian sandstones of the Amsden and Tensleep 124 Formations, and by the Permian limestones of the Phosphoria Formation. The competent core of the sedimentary succession lies in the Bighorn- Phosphoria interval, where most of 125 the previously published microstructural work was carried out (Amrouch et al., 2010a; 126 127 Amrouch et al., 2010b; Beaudoin et al., 2012; Bellahsen et al., 2006; Neely and Erslev, 2009). 128 Three systematic, bed-perpendicular joint/vein sets (opening mode I) are encountered at the basin scale (Beaudoin et al., 2014a and references therein). In sequence, the first set 129 130 comprises joints/veins striking N110° (after unfolding), likely formed during layer-parallel shortening related to the Sevier contraction. The two other sets are related to the Laramide 131 132 contraction : the oldest one comprises joints/veins striking N045° (after unfolding) and marks 133 the Laramide layer-parallel shortening; the youngest one comprises joints/veins striking 134 N135° that developed in response to outer-arc extension at the hinge of growing Laramide basement-cored folds (see Tavani et al., 2015 for a complete description of deformation 135 136 patterns in folds and orogenic forelands). Absolute U-Pb ages from the calcite cements of these vein sets (Beaudoin et al., 2018) further suggest that Sevier-related veins developed 137 138 earlier in the west (90 Ma) than in the east (75 Ma) of the basin while Laramide-related veins 139 developed earlier in the east (72 Ma and 45 Ma) than in the west (60 Ma and 28 Ma) of the 140 basin. Tectonic, bedding-perpendicular stylolites have also been described in the basin (Amrouch et al., 2010; Beaudoin et al., 2012; Beaudoin et al., 2020); their peaks are mostly 141 oriented ~N110° and N045°, thus reflecting Sevier and Laramide layer-parallel shortening, 142 143 respectively. Finally, mesoscale reverse and strike-slip faults have been extensively documented at the basin scale (Neely and Erslev, 2009; Amrouch et al., 2010; Beaudoin et al.,

145 2012) and have also been interpreted as related to either Sevier or Laramide shortening.

146 Bedding-parallel stylolites were collected within 4 different basement-cored folds 147 located in the BHB along an E-W transect (Figure 1b, c): the Rattlesnake Mountain Anticline in the west, the Sheep Mountain and the Little Sheep Mountain Anticlines in the east, and the 148 western part of the Bighorn Mountain arch that bounds the basin to the east. Because 149 150 stylolite occurrence depends on lithology (e.g., Marshak and Engelder, 1985), sampling was 151 limited to the dolomitic parts of the Bighorn and Madison Formations. We collected bedding-152 parallel, pre-folding sedimentary stylolites, the development of which is related to vertical 153 compaction related to the burial of the strata (Figure 2). Sedimentary facies were checked following the description established by Barbier et al. (2012) to ensure bedding-parallel 154 155 stylolites were hosted by clay-poor dolomite of which dolomitization corresponds to the early 156 deposition surface chemistry (eogenesis), i.e., predating burial and related stylolitisation 157 (mesogenesis). That is important in order to limit the variability of mechanical and chemical parameters used for the stylolite roughness inversion. Hence samples were collected from 158 159 the top of the Madison Formation, i.e. the Little Tongue and Bull Ridge members which are massive dolomite of which dolomitization occurred before stylolite development (Figure 3c 160 161 d). The Bighorn Formation represents pure dolomite that completely dolomitized before 162 burial (Figure 3e, Blackwelder, 1913). We collected samples in the top part of the Bighorn 163 Formation at the two locations it crops out, the Rattlesnake Mountain Anticline and the Bighorn Mountains. We focused as much as possible on bedding-parallel stylolites showing a 164 morphology of suture and sharp peak type (Figure 3a b, Koehn et al., 2016), that have been 165 shown to reflect the maximum burial depth experienced by the strata whilst the maximum 166 167 principal stress ( $\sigma_1$ ) was vertical (Beaudoin et al., 2019a).

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Figure 2: Field photographs of bedding-parallel stylolites observed in the different structures and formations. Bedding is reported as a dotted black line, bedding-parallel stylolites are pointed out by white arrows, and the top of the pictures is up. a) Bighorn Mountains, Madison Formation. b) Bighorn Mountains, Bighorn Formation. c) Rattlesnake Mountain Anticline, Bighorn Formation. d) Rattlesnake Mountain Anticline, Madison Formation. e) Rattlesnake Mountain Anticline, Madison Formation, the insert shows the sequence between bedding-parallel stylolite and Sevier related tectonic vein. f) Sheep Mountain Anticline, Madison Formation.



Figure 3 : a-b) High-resolution scans of polished slabs showing bedding-parallel stylolites, top of the
strata is up. a) Scan of the sample RM-S26B, in the Madison Formation, that evidences that
dissolution along the bedding-parallel stylolites predates development of tectonic vein. b) Scan of the
sample LSM-S8, in the Madison Formation. c-e) microphotographs showing the texture of the host
rock in the Madison Formation. (c, d at Sheep Mountain Anticline) and in the Bighorn Formation (e, in
the Bighorn Mountains).

# 183 STYLOLITE ROUGHNESS INVERSION FOR STRESS

The growth and the morphology of a stylolite are rate-dependent and are governed 184 by the kinetics of dissolution (Stockdale, 1922) and by the distribution of heterogeneities, as 185 well as by the amount of clay (Renard et al., 2001). Once dissolution starts, there is a 186 187 thermodynamic competition between a destabilizing (roughening) force due to pinning particles on the stylolite surface that resist dissolution, and two stabilizing (smoothening) 188 forces, long-range elastic forces and local surface tension, that tend to flatten the stylolite 189 190 surface by preferentially dissolving areas of local roughness (Schmittbuhl et al., 2004). The inversion of the stylolite roughness exploits the self-affine properties of the stylolite 191 192 roughness by treating it as a signal that is governed by two different processes according to 193 the scale of observation (Schmittbuhl et al., 2004), i.e., the surface energy at small-scale (typically < 1 mm) and the elastic energy at larger scales. For all the stylolites that follow the 194 195 growth model the method is valid for (Koehn et al., 2007), the spectral analysis of the 196 roughness returns two different roughness exponents (Hurst exponents) (Schmittbuhl et al., 197 2004; Toussaint et al., 2018). The scale at which there is a transition from one Hurst coefficient to another one is referred to as the cross-over length (Lc), and is analytically linked to the 198 applied mean stress magnitude ( $\sigma_m = \frac{\sigma_1 + \sigma_2 + \sigma_3}{3}$ , in Pa) and differential stress magnitude ( $\sigma_d$ 199 =  $\sigma_1$ - $\sigma_3$ , in Pa), to the elastic parameters of the rock (Young modulus E (in Pa) and Poisson 200 ratio) and to the solid-fluid interfacial energy  $\gamma$  (in J.m<sup>-2</sup>) (Ebner et al., 2009; Schmittbuhl et 201 al., 2004): 202

203

$$Lc = \frac{\gamma E}{\beta \sigma_m \sigma_d} \tag{1}$$

where  $\beta = v(1 - 2v)/\pi$ , a dimensionless constant with v being the Poisson's ratio.

While the shape of a stylolite is affected by the strain rate and the stress orientation 205 during its growth (Ebner et al., 2009; Koehn et al., 2012), the final roughness of a stylolite is a 206 saturation state that is reached over a short period of time at the end of dissolution (Ebner et 207 al., 2009; Rolland et al., 2012). Hence the final Hurst exponent that characterises the 208 209 roughness of a stylolite is different from the width and the amplitude parameters of a stylolite that are related to the growth rate (Koehn et al., 2012) and to the amount of chemical 210 211 compaction (e.g., Angheluta et al., 2012). The final stylolite roughness can be then treated as 212 a snapshot of the prevailing ambient stress at the time it stopped being active (Schmittbuhl et al., 2004; Ebner et al., 2009; Rolland et al., 2012). Finally, as the dissolution occurs along a 213 214 fluidic film, stylolite roughness inversion is not sensitive to the local fluid pressure, allowing to calculate the depth at which bedding-parallel stylolites stopped growing under a vertical 215 216 maximum principal stress and using the average dry density of the overlying sandstones and 217 carbonate rocks in the area (2.4 g.cm<sup>-3</sup>, Manger, 1963).

218 A number of studies explored in depth which signal analysis tool is the best for roughness inversion applied on stylolites. Three main methods can be used, the Fourier Power 219 220 Spectrum, the Height Correlation Function, and the Average Wavelet Coefficient (Ebner et al., 2009; Ebner et al., 2010; Renard, 2004; Rolland et al., 2014; Schmittbuhl et al., 2004; Toussaint 221 222 et al., 2018). For our study, we have chosen to conduct the signal analysis with the method 223 that has proven to be the less impacted by sample number and quality: the Average Wavelet 224 Coefficient (AWC) method with Daubechies D4 wavelets (Ebner et al., 2009; Simonsen et al., 1998). The AWC analysis reconstructs the signal as a sum of different wavelets, starting with 225 a mother function (Simonsen et al., 1998), the scale a (mm) and the averaged wavelet 226 coefficient W(a) being related as W(a)= $a^{(H+0.5)}$ , where H is the roughness exponent, or Hurst 227

exponent. AWC must return a Hurst exponent of 0.5 at the large scale and of 1.1 at the small-228 scale (Ebner et al., 2009; Rolland et al., 2014; Schmittbuhl et al., 2004; Toussaint et al., 2018). 229 230 In order to apply the stylolite roughness inversion for stress, samples were cut perpendicular to the stylolite planes. Stylolites with peaks that are oriented perpendicular to 231 the solution planes were selected, then slabs were manually polished using abrasive mats 232 from coarse (250  $\mu$ m) to extra fine (2.5  $\mu$ m) in order to avoid alteration to the stylolite track. 233 234 The stylolites were then scanned at a resolution of 12800 dpi using a commercial 2-D scanner. 235 Stylolite tracks were hand drawn as 8 bits, 5 pt- thick pixelated lines using the drawing software GIMP (Figure 4). The inversion process is using Matlab scripts that have been made 236 237 available by Ebner et al. (2009) for AWC. The 1-D signal is then analyzed, and the consistency 238 to the theory of the two governing processes is tested by fitting a non linear least square regression through the signal (Ebner et al., 2009). We fix the slopes of the non-linear 239 240 regression to the two theoretical Hurst exponents. When the stylolite roughness is consistent 241 with the model, the script will return two slopes with a corresponding cross-over length Lc (Figure 4a-b). Otherwise, the script will return a single slope with an extreme cross-over 242 243 length value (Figure 4c), that cannot be used for further stress inversion, the stylolite is then discarded. By applying this method on numerical signals, of which Hurst exponent and cross-244 245 over length were set beforehand, the error on determination of the cross over length due to 246 the non-linear regression has been estimated around 23% (Rolland et al., 2014).

In the case of bedding-parallel-stylolites we assume a zero horizontal displacement in the stylolite plane, corresponding to a perfect isotropy of the horizontal principal stresses, such that  $\sigma_v > \sigma_H \approx \sigma_h$  where  $\sigma_v$ ,  $\sigma_H$  and  $\sigma_h$  are the absolute magnitude (i.e. without considering any effect related to fluid pressure) of the vertical principal stress, of the maximum horizontal principal stress and of the minimum horizontal principal stress, respectively. This leads to the
 simplification of equation 1 as follows

$$\sigma_{\nu}^2 = \frac{\gamma E}{\alpha L c} \tag{2}$$

with  $\alpha = \frac{(1-2\nu)*(1+\nu)^2}{30\pi(1-\nu)^2}$  (Ebner et al., 2009). In order to obtain the vertical stress, we need to 254 consider the appropriate values for the solid-fluid interfacial energy  $\gamma$  (J.m<sup>-2</sup>), the Young 255 modulus E and the Poisson ratio v at the time the dissolution ended. For y we used the known 256 value for dolomite, γ=0.24 J.m<sup>-2</sup> (Wright et al., 2001). For the mechanical parameters, we used 257 258 the results of mechanical tests conducted on the most homogeneous, fracture-free samples 259 of the Madison and Bighorn Formations that it was possible to collect. The tests on the Madison Formation were published elsewhere (Amrouch, 2010), and we produce new results 260 261 for the Bighorn Formation (Supplementary Material). In nature and in the tests, the variability of the Poisson ratio is negligible for a given material, so we use the average value from the 262 mechanical tests, i.e. 0.2 for the Madison Formation and 0.26 for the Bighorn Formation. We 263 264 also use the average values returned by the tests for the Young modulus, i.e. 29 GPa for the 265 Madison Formation and 43 GPa for the Bighorn Formation. It is noteworthy that the validity of the estimates of the Young modulus is more problematic as (1) it is highly variable in nature, 266 even for a given material; (2) it shows some variability in the mechanical testing; and (3) it is 267 expected to vary during the burial history. When considering all parameters the overall 268 uncertainty on the stress value was estimated previously to ca. 12% (Rolland et al., 2014). 269 270 Finally, the depth h is obtained using  $\sigma_v = \rho g h$ , with  $\rho$  the dry density (2.4 g.cm<sup>-3</sup>), and g the gravitational field acceleration (9.81 m.s<sup>-2</sup>). Here we consider the uncertainty on the depth to 271 272 reflect the uncertainty on the stress estimate, i.e. 12%. As bedding-parallel stylolites can form 273 at nearly any depth as long as the maximum principal stress is vertical, we consider a

- 274 population of stylolites to be representative of the range of depths reached by the strata of
- 275 interest.



Figure 4: Three examples of application of the stylolite roughness inversion technique. For each example given here the polished slab is scanned in 2D (top), then hand drawn with a 5 pixel-thick line (middle), then analyzed using an average wavelet spectrum method (bottom). See text for more details. Example c) illustrates a failure in the inversion process, since a single slope is derived.

276

278 **RESULTS** 

At the basin scale, inversion of stylolite roughness was successfully applied on 51 279 bedding-parallel stylolites (Rattlesnake Mountain, n=12; Little Sheep Mountain-Sheep 280 281 Mountain, n=22; Bighorn Mountain, n=17) out of the 58 stylolites tested (Table 1), that were 282 hosted in the Madison Formation (n= 38) and in the Bighorn Formation (n=13). All results are reported in table 1; and distribution of vertical stress magnitudes at the basin-scale (Figure 5) 283 284 ranges (1) in the Bighorn Formation from 19 ±2 MPa to 35 ±4 MPa, corresponding to depths 285 ranging from 800 ±90 m to 1500 ±170 m (median = 1104 m) and (2) in the Madison Formation from 12 ±1.5 MPa to 37 ±4 MPa, corresponding to depths ranging from 510 ±60 m to 1570 286 287 ±190m (median = 850 m). At Rattlesnake Mountain Anticline, the vertical stress ranges from 15  $\pm$ 2 MPa to 27  $\pm$ 3 MPa, corresponding to depths ranging from 930  $\pm$ 110 m to 1150  $\pm$ 140 m 288 289 for the Bighorn Formation (median = 1060 m), and from 640 ±80 m to 1150 ±140 m for the 290 Madison Formation (median = 720 m). At Little Sheep Mountain – Sheep Mountain Anticlines, 291 the vertical stress ranges from 14 ±1.5 MPa to 37 ±4 MPa, corresponding to depths ranging 292 from 640 ±80m to 1570 ±190m for the Madison Formation (median = 990 m). Finally, in the 293 western flank of the Bighorn Mountains, the vertical stress ranges from 12 ±1.5 MPa to 35 ±4 MPa, corresponding to depths ranging from 800 ±90 m to 1500 ±170 m for the Bighorn 294 Formation (median = 1147 m), and from 510 ±60 m to 980 ±120 m for the Madison Formation 295 296 (median = 765 m) (Figure 6). The success of the inversion process (88%) was tested regarding 297 the morphological type of the stylolites, the percentage of failure being much higher for the rectangular layer and seismogram pinning types (40%, for both types) than for the suture and 298 299 sharp peak type (6.5%). However, as the population is very small for both rectangular layer 300 and *seismogram pinning* types (7 and 5, respectively), no definitive behaviour can be deduced 301 from this study only.

Sample name and location (WGS84, decimal											
		degrees	)		(mm)*	Vertical Stress	Depth (m)**				
Name	GPS	Longitude	Latitude	Formation	(1111)	(MPa)**					
Rattlesnake Mountain Anticline											
RM-S1	27	-109,148	44,511	Bighorn	1,466	22	934				
RM-S3	27	-109,148	44,511	Bighorn	0,921	27	1147				
RM-S4	27	-109,148	44,511	Bighorn	0,998	26	1104				
RM-S5	27	-109,148	44,511	Bighorn	1,15	24	1019				
RM-S9	28	-109,197	44,501	Madison	1,658	17	722				
RM-S9-2	28	-109,197	44,501	Madison	0,865	23	977				
RM-S18	31	-109,196	44,501	Madison	1,885	16	680				
RM-S31	39	-109,136	48,858	Madison	0,636	27	1147				
RM-S26T	33	-109,191	44,501	Madison	1,579	17	722				
RM-S26B	33	-109,191	44,501	Madison	2,103	15	637				
RM S27	39	-109,191	44,501	Madison	1,579	17	722				
RM-S29	39	-109,136	48,858	Madison	0,973	22	934				
RM-S35	39	-109,136	48,858	Madison	N.A.	N.A.	N.A.				
Little Shee	ρ Μοι	untain Anticl	ine								
LSM-S1	38	-108,19	44,513	Madison	1,36	19	807				
LSM-S2	38	-108,19	44,513	Madison	0,572	29	1232				
LSM-S3	38	-108,19	44,513	Madison	0,759	25	1062				
LSM-S4	38	-108,19	44,513	Madison	0,972	22	934				
LSM-S5	38	-108,19	44,513	Madison	0,445	32	1359				
LSM-S6	38	-108,19	44,513	Madison	0,482	31	1317				
LSM-S8-1	38	-108,19	44,513	Madison	2,283	14	595				
LSM-S8-2	38	-108,19	44,513	Madison	0,838	24	1019				
LSM-S14	38	-108,19	44,513	Madison	1,673	17	722				
LSM-S15	38	-108,19	44,513	Madison	0,43	33	1402				
LSM-S16	38	-108,19	44,513	Madison	1,443	18	765				
Sheep Mou	untain	Anticline									
SM-S1	1	-108,139	44,606	Madison	0,34	37	1572				
SM-S5	4	-108,138	44,61	Madison	0,639	27	1147				
SM-S15	9	-108,135	44,613	Madison	1,491	18	765				
SM-S17	10	-108,134	44,613	Madison	0,69	26	1104				
SM-S18-1	11	-108,134	44,613	Madison	0,88	23	977				
SM-S18-2	11	-108,134	44,613	Madison	0,912	23	977				
SM-S18-3	11	-108,134	44,613	Madison	1,748	16	680				
SM-S6-1	17	-108,14	44,611	Madison	0,336	37	1572				
SM-S7	17	-108,14	44,611	Madison	0,53	30	1274				
SM-S11	17	-108,14	44,611	Madison	1,043	21	892				
SM-S26	34	-108,143	44,625	Madison	1,504	18	765				
Western Part of the Bighorn											
Mountain											
BM-S1	18	-107,709	44,569	Madison	3,408	12	510				
BM-S4-1	18	-107,709	44,569	Madison	1,346	19	807				

Table 1. Sample name and location, and results from stylolite roughness inversion by average wavelet coefficient

Sample	name	and location	n (WGS84,	Lc (mm)*	Vertical Stress	Depth (m)**				
		degrees	)							
Name	GPS	Longitude	Latitude	Formation	(11111)	(MPa)**				
BM-S4-2	18	-107,709	44,569	Madison	1,679	17	722			
BM-S5-1	18	-107,709	44,569	Madison	1,294	19	807			
BM-S5-2	18	-107,709	44,569	Madison	2,626	13	552			
BM-S8	18	-107,709	44,569	Madison	0,884	23	977			
BM-S19	19	-107,7	44,575	Madison	2,866	13	552			
BM-S20	19	-107,7	44,575	Madison	1,359	19	807			
BM-S22	20	-107,691	44,579	Bighorn	0,567	35	1487			
BM-S25-1	21	-107,689	44,579	Bighorn	1,118	25	1062			
BM-S25-3	21	-107,689	44,579	Bighorn	0,911	27	1147			
BM-S25-2	21	-107,689	44,579	Bighorn	0,584	34	1444			
BM-S26-2	21	-107,689	44,579	Bighorn	0,698	31	1317			
BM-S27-1	21	-107,689	44,579	Bighorn	1,144	24	1019			
BM-S27-2	21	-107,689	44,579	Bighorn	1,694	20	849			
BM-S30	22	-107,968	44,795	Bighorn	1,956	19	807			
BM-S32	24	-107,967	44,795	Bighorn	0,9	27	1147			
BM-S21	20	-107,691	44,579	Bighorn	N.A.	N.A.	N.A.			
BM-S26-1	21	-107,689	44,579	Bighorn	N.A.	N.A.	N.A.			
BM-S28	21	-107,689	44,579	Bighorn	N.A.	N.A.	N.A.			
BM-S29	21	-107,689	44,579	Bighorn	N.A.	N.A.	N.A.			
BM-S35	24	-107,967	44,795	Bighorn	N.A.	N.A.	N.A.			
BM-S39	26	-107,97	44,792	Madison	N.A.	N.A.	N.A.			

Table 1. Suite

\*: Crossover length Lc is given within 23%, N.A. represents samples for which the method failed (12% of the population)

\*\*: Vertical stress and depth values are given with 12% relative error

302



303

304 Figure 5: Bar plot showing the distribution of the depth values obtained from stylolite roughness

inversion of the bedding-parallel stylolites population at the basin-scale, irrespective of the fold

306 structure where samples were collected. Depths related to bedding-parallel stylolites sampled in the

307 Madison Formation are reported in grey while the ones coming from the Bighorn Formation are

308 reported in red. – SRIT – Stylolite roughness inversion technique

#### 309 DISCUSSION

Stylolite roughness inversion was used to consistently estimate the maximum vertical 310 311 stress experienced by dolomitic strata for as long as the vertical stress magnitude stayed 312 higher than the horizontal stress magnitude, i.e. since early burial to Sevier - Laramide shortening. Our results show that the range of vertical stress values and so the burial depths 313 are similar for both considered formations (Figure 5), and that the maximum depth at which 314 315 bedding-parallel stylolites stopped being active is lower than the maximum depth predicted 316 by the cumulated thickness of the overlying formations (Fox and Dolton, 1996; Neely and Erslev, 2009)(Figure 6). Considering the median depth of the populations (Figure 5), the 317 318 stylolites from the Madison Formation recorded shallower depths (median = 850m) than the ones from the Bighorn Formation (median =1104 m), and the difference of 250 m. Considering 319 the median depth by structure (Table 1, Figure 6), the difference is of 300 m in the Rattlesnake 320 321 Mountain Anticline and of 300 m in the western flank of the Bighorn Mountains. These 322 differences in depth are well in line with the thickness between the top of each formation estimated from the sedimentary column and well logs (~ 300 m, Figure 1, Fox and Dolton, 323 324 1996; Durdella, 2001). Results of this study add up to the growing number of evidence that stylolite roughness inversion is a powerful tool to access paleoburial without any assumption 325 on the past thermal gradient or fluid pressure (Ebner et al., 2009; Rolland et al., 2012; 326 327 Beaudoin et al., 2016; Bertotti et al., 2017; Beaudoin and Lacombe, 2018; Beaudoin et al., 328 2019a). Each of the depth values we obtained represents the stage at which some part of the stylolite population stopped being active (Toussaint et al., 2018). Consequently, the minimum 329 depths recorded (from 500 to 800 m) correspond to the minimum depths at which the 330 331 development of some bedding-parallel stylolites halted; in other words pressure solution 332 likely started at burial depth shallower than 500m, even though the minimum depth required for stylolites formation still remains unknown. This illustrates that chemical compaction is a
mechanism that can be active at very shallow depth (Ebner et al., 2009; Rolland et al., 2014),
discarding theories that predict a minimum depth of 800 m to form bedding-parallel stylolites
in the absence of clay-enhanced reactions (Finkel and Wilkinson, 1990 and others; Railsback,
1993).

In order to discuss the timing and the duration of bedding-parallel stylolite development, we projected the range of calculated depths onto the burial models for the eastern and the western parts of the Bighorn basins proposed in Beaudoin et al. (2014b) after May et al. (2013), based on well data and thermochronology (Figure 6). The validity of this burial-exhumation model is supported by the recent burial models based on organic matter in the basin (Ellis et al., 2017; Gottardi et al., 2019).



Figure 6: Burial curves (modified after May et al. (2013), Beaudoin et al. (2014b)) valid for the western 345 margin of the basin (left-hand side) and for the eastern margin of the basin (right-hand side). The 346 347 absolute age of the development of the tectonic veins are reported after Beaudoin et al. (2018) as blue highlights (Sevier) and green highlights (Laramide). In each case, the range of depth obtained from 348 applying the stylolite roughness inversion technique on the local population of bedding-parallel 349 350 stylolites is reported on the y-axis and on the graph in the considered formations. Corresponding timing of activity for chemical compaction along bedding-parallel stylolites is projected on the x-axis. 351 352 Uncertainties of ca. 12% are reported along as lighter grey area. Bar plots of the distribution of the depths vs the number of stylolite analyzed are reported for each side of the basin, with data from the 353 354 Madison Formation in grey and data from the Bighorn Formation in red.

The bedding-parallel stylolites we studied happen to show some clear field evidence that they developed prior to the formation of the tectonic, bed-perpendicular veins (Figures 2, 3), the orientations of which that indicate that they are related to either Sevier contraction (~E-W) or Laramide contraction (NE-SW) (Amrouch et al., 2010a; Beaudoin et al., 2012; Bellahsen et al., 2006; Craddock and van der Pluijm, 1999; Varga, 1993). Therefore, one can safely consider that the bedding-parallel stylolites have developed prior to the Sevier and Laramide shortening, i.e., during burial and not during subsequent Paleogene exhumation.

363 The timing of opening of the Sevier and Laramide related veins in the studied structures has been set by means of U-Pb absolute dating (Beaudoin et al., 2018; 2019b) and 364 is reported on the burial curves (Figure 6). With respect to the fold structure and the 365 formation considered, it appears that the maximum depths recorded by the bedding-parallel 366 367 stylolites correspond to an age which is always older than, or equal to, the one of the developments of tectonic veins related to Sevier layer-parallel shortening. This supports the 368 369 results of the inversion as one can expect bedding-parallel stylolites have stopped being active 370 when the maximum principal stress  $\sigma_1$  switched from vertical to horizontal, as peaks of 371 stylolites are parallel to the  $\sigma$ 1 orientation (e.g. Koehn et al., 2007). In the Bighorn Basin, the stress switch happened when the magnitude of the maximum horizontal stress transmitted 372 373 forelandward from the Sevier front overcame the magnitude of the vertical stress related to burial. We can estimate that this switch from vertical to horizontal maximum principal stress 374 occurred for an absolute value of at least 35 ±4 MPa. 375

Paleopiezometric studies of calcite twinning in the Madison Formation at the Sheep Mountain Anticline (Amrouch et al., 2010) and at the Rattlesnake Mountain Anticline (Beaudoin et al., 2012) documented differential stress ( $\sigma_1 - \sigma_3$ ) magnitudes, values of the

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stress ratio ( $\varphi = \frac{(\sigma_2 - \sigma_3)}{(\sigma_1 - \sigma_3)}$ ), and stress regimes prevailing in the rocks during the 379 Sevier layer-parallel shortening. From these data, it is possible to derive the absolute 380 magnitude of  $\sigma_1$  if we set the depth of deformation. Considering the range of burial depths 381 valid at Sheep Mountain Anticline during the Sevier layer-parallel shortening, as given by the 382 383 absolute age of the tectonics veins projected on the burial model (Figure 6), we can estimate 384 that the absolute magnitude of  $\sigma_1$  ranges from 55±10 MPa at the minimum depth (1500 m) 385 to 79 ± 10 MPa at the maximum depth (2500 m). At the Rattlesnake Mountain Anticline, the same approach yields absolute magnitude of  $\sigma_1$  ranging from 65 ± 7 MPa at minimum depth 386 (1300 m) to 100 ± 7 MPa at maximum depth (2800 m). A more recent paleopiezometric study 387 based on tectonic stylolite roughness inversion in the Madison Formation at the Sheep-Little 388 Sheep Mountain Anticlines (Beaudoin et al., 2020) further provides absolute magnitudes of 389 390 the maximum horizontal principal stress  $\sigma_1$  related to the Sevier layer-parallel shortening 391 ranging from 50±2 MPa at minimum depth (1500 m) to 67 ± 5 MPa at maximum depth (2500 m). All these independent estimates indicate that the absolute magnitudes of the maximum 392 horizontal principal stress  $\sigma_1$  associated with calcite twinning and tectonic-related pressure 393 394 solution related to the Sevier contraction were consistently higher than the one (35 ±4 MPa) 395 recorded at the time  $\sigma_1$  switched from vertical to horizontal. The present study therefore 396 reveals that the Sevier-related maximum horizontal principal stress increased faster than the 397 vertical stress related to burial during orogenic stress loading. This is consistent with a regional stress build-up model, with the maximum horizontal stress expectedly overcoming 398 the vertical stress by a margin before being able to trigger anisotropic deformation such as 399 400 joints/veins and stylolites. Furthermore, our study supports that the orogenic stress build-up

401 was rather fast since there is little time (~5 Ma, Figure 6) between the activity of the last
402 bedding-parallel stylolite and the opening of the first Sevier-related tectonic vein.

403 Finally, our results indicate that the investigated bedding-parallel stylolite population was active at least from, and likely before, ca. 240 Ma ago until 85 Ma ago in the western part 404 405 of the basin, and at least from, and likely before, 230 Ma ago until 80 Ma ago in the eastern 406 part of the basin (Figure 5). By comparing the paleodepth to the burial curve, we can provide for the first time a time bracket for the expected long-lasting development of a population of 407 408 bedding-parallel stylolites. Such information is of importance when it comes to reservoir 409 property evolution during burial, as stylolites have a strong influence on the porosity, permeability and mechanical properties of carbonates (Aharonov and Karcz, 2019; Bruna et 410 al., 2019; Martín-Martín et al., 2018; Toussaint et al., 2018). 411

412

# 413 **CONCLUSIONS**

This study uses stylolite roughness paleopiezometry to constrain the magnitude of the vertical 414 stress and the burial depth of selected strata during the foreland evolution of the Bighorn 415 416 Basin, Wyoming, USA. The results show that the paleopiezometric analysis of the roughness 417 of a relatively small population (n=51) of bedding-parallel stylolites can reliably return the 418 pre-shortening burial evolution over a long period of time (~150 Ma). Bedding-parallel 419 stylolites also yield the depth and timing at which the maximum principal stress switched 420 orientation from burial-related vertical to orogenic contraction-related horizontal, both being supported by the available absolute ages of the kinematically compatible tectonic veins. 421 422 Beyond regional implications, this study illustrates the potential of the inversion of the 423 bedding-parallel stylolite roughness, conducted with a wavelet analyses, as a reliable and

- 424 powerful paleopiezometric tool for basin and structural analyses, allowing for paleoburial
- 425 estimates independently of past geothermal gradients, and adding important information
- 426 about the timing of orogenic stress build-up in orogenic forelands.
- 427

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