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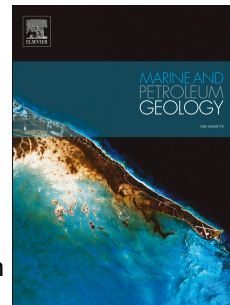
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## **CREdIT Author Statement**

**Marta Gasparrini** : Conceptualization, Writing - Original Draft, Supervision, Investigation, Visualization

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**Tristan Euzen**: Writing - Review & Editing, Resources

Journal Pre-proof

1 **Natural mineralized fractures from the Montney-Doig**  
2 **unconventional reservoirs (Western Canada Sedimentary Basin):**  
3 **timing and controlling factors**

4  
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13

14 **ABSTRACT**

15 Characterizing the origin of natural fractures in organic-rich fine-grained deposits is key to  
16 constraining permeability evolution in these potential source rocks and tight reservoirs, as well as to  
17 assess the hydraulic connectivity of the fluid systems in which they develop.

18 Differently oriented calcite-filled fractures (i.e. veins), hosted by organic-rich mudrocks of the  
19 Lower-Middle Triassic Montney-Doig unconventional resource play (Western Canada Sedimentary  
20 Basin), were sampled in sub-surface well cores from British Columbia. A multidisciplinary approach  
21 (including sedimentology, Rock-Eval pyrolysis, petrography, O-C-Sr isotope geochemistry and fluid  
22 inclusion microthermometry) was applied to host-rocks and fracture filling calcites.

23           The results demonstrate the relevance and usefulness of such multidisciplinary studies to  
24 gather insights on: 1) the lithology-related factors controlling fracture occurrence; 2) the timing of  
25 fracture opening and the origin of the circulating paleo-fluids; 3) the openness of the fluid system  
26 through time.

27           More specifically, host-rock facies (particularly grain size) and vertical facies changes appear  
28 to be the leading factors controlling fracture occurrence. A less relevant role was played by the  
29 occurrence of diagenetic carbonates, while TOC possibly did not control fracture occurrence.

30           Three generations of calcite cemented fractures were identified. Vertical fractures (first  
31 generation) post-dated the onset of oil generation (Late Cretaceous). Horizontal, bedding-parallel  
32 fractures (second generation) post-dated the onset of gas generation and possibly opened close to  
33 maximum burial, corresponding to peak hydrocarbon (CH<sub>4</sub>) generation (Late Cretaceous - Early  
34 Paleogene). Vertical fractures (third generation) post-dated the horizontal ones and opened during  
35 basin uplift (Middle to Late Paleogene).

36           The consistent petrographic and geochemical features of all the calcite cements point to  
37 parent fluids in equilibrium with the host-rock lithologies, that possibly behaved as a closed hydraulic  
38 system during Late Cretaceous to Paleogene time; this would support the hypothesis that, at least in  
39 the portion of the basin investigated, the Montney Fm also acted as source rock of the  
40 unconventional system.

41           Keywords: Montney-Doig, fracture diagenesis, paleo-fluids, source rocks, tight reservoirs

42

## 43           **1. INTRODUCTION**

44           Fine-grained siliciclastic deposits like shales and siltstones (i.e. mudrocks, *sensu* Lazar et al.,  
45 2015) largely contribute to the filling of sedimentary basins and most commonly constitute low-

46 permeability rock units, unless they are affected by fracturing processes. Natural fractures commonly  
47 occur in mudrocks (e.g. Gale et al., 2014; Ukar et al., 2017; Hooker et al., 2019), though their  
48 occurrence and the extent to which they enhance the overall rock permeability still remain poorly  
49 constrained (Dewhurst et al. 1999).

50         The development and sealing of natural fractures within organic-rich mudrock successions  
51 have received great attention since the interest of exploration in unconventional gas and oil plays  
52 rose, particularly in North America (Curtis, 2002; Gale et al., 2007; Engelder et al., 2009). These plays  
53 encompass a variety of low-permeability reservoirs that require hydraulic fracturing to be  
54 commercially productive and are generally found within or in close association with source rock  
55 intervals. The rising interest for unconventional oil and gas plays revealed that source rocks in these  
56 settings may also represent the reservoirs of HC accumulations which may be generated *in-situ* or  
57 may have migrated from external sources. Discriminating among the possible behaviors (open *versus*  
58 closed system) of organic-rich mudrock successions during burial and thermal maturation in such  
59 plays is relevant to, among others, better assess the amounts of ultimate HC in place (e.g. Jarvie et  
60 al., 2007).

61         On the other hand, characterizing the processes that cause natural fractures in fine-grained  
62 organic-rich deposits is crucial to better understand permeability evolution within these potential  
63 tight reservoirs (Gale and Holder, 2010; Cobbold et al., 2013; Gasparrini et al., 2014). Several  
64 mechanisms may act independently or in combination to cause fracture growth such as the  
65 conversion of organic matter in HC (Spencer, 1987; Ozkaya, 1988; Vernik, 1994; Jochum et al., 1995;  
66 Marquez and Montjoy, 1996; Zanella et al., 2014; Meng et al., 2017), the build-up of tectonic stress  
67 at different moments of the basin history (Ukar et al., 2017; Hooker et al., 2019), as well as the  
68 occurrence of specific diagenetic processes (e.g. Van de Kamp, 2008; Gasparrini et al., 2014; Hooker  
69 et al., 2017a). Regardless of the mechanisms, mineral cementation may accompany fracturing to  
70 form mineralized fracture (i.e. veins), which investigation may help discriminating fracture growth  
71 timing and/or genetic mechanism. This may be relevant for predicting evolution of hydraulic

72 connectivity through time and joint relationships with organic matter maturation that can provide  
73 constraints to calibrate numerical models for the exploration and production of unconventional  
74 resources (e.g. Sassi et al., 2012; Romero-Sarmiento et al., 2013).

75         Additionally, the localization of the fractured intervals in the organic-rich mudrocks may be  
76 controlled by different factors capable of affecting the mechanical properties of the rocks (e.g.  
77 Young's modulus, Poisson's ratios). Primary sedimentary features of these deposits (e.g. grain size,  
78 total organic carbon (TOC), mineralogy), their diagenetic modifications during burial (e.g.  
79 cementations, replacements) and the vertical heterogeneities of the succession (e.g. bed thickness,  
80 facies stacking pattern, fissility), may control mechanical properties and therefore the localization of  
81 fractures and consequently, the rock response to hydraulic fracturing (e.g. Vishkai et al., 2017).  
82 However, consensus is not reached in literature on the role of these different factors (Engelder and  
83 Peacock, 2001; Rijken and Cooke, 2001; Peacock and Mann, 2005; Wang and Gale, 2009; Gale et al.  
84 2014; Wang et al., 2016; Ilgen et al., 2017; Hooker et al., 2020; Peng et al., 2020) and this is especially  
85 true when referring to the coarser mudrock end-member, i.e. the siltstones (e.g. Vaisblat et al.,  
86 2017a, 2019; Chatellier et al., 2018).

87         For the different reasons mentioned above, characterizing the factors controlling the  
88 occurrence of natural fractures in organic-rich fine-grained deposits, as well as framing the timing  
89 and mechanism of fracturing within the local geological evolution, is key to evaluate permeability and  
90 fluid pathway potential through time of these rocks, and hence their present reservoir properties and  
91 sealing capacity.

92         Our study focuses on the mudrocks from the Lower-Middle Triassic Montney-Doig Fms in the  
93 Western Canadian Sedimentary Basin (WCSB). More specifically, the Montney Fm is one of the  
94 largest economically feasible resource plays and is classified as an unconventional tight oil and gas  
95 siltstone reservoir (Davies et al., 1997; Moslow, 2000; Euzen et al., 2018). Since the failure mechanics  
96 of such rocks is poorly understood and seldom incorporated in hydraulic fracturing or geomechanical  
97 models, this formation has received much attention in terms of characterization of physical and

98 mechanical properties from laboratory tests (Ghanizadeh et al., 2015a,b; Vishkai et al., 2017; Vaisblat  
99 et al., 2017a, 2019; Riazi et al., 2017; McKean and Priest, 2019). Geometric features and structural  
100 fabrics of fractures in the Montney Fm have been investigated from well cores by few authors  
101 (Davies et al., 2014; Gillen et al., 2019). Surprisingly however, poor attention has been paid to the  
102 characterization of the distribution (in space and time) of natural mineralized fractures.

103 In an attempt at shedding new light onto these last issues, we investigated mineralized  
104 fractures occurring in the Montney-Doig lithologies from two well cores located in British Columbia  
105 by applying a multidisciplinary approach (including sedimentology, Rock-Eval pyrolysis, petrography,  
106 O-C-Sr isotope geochemistry and fluid inclusion microthermometry) to host-rocks and fracture filling  
107 calcites. The main purposes of this survey were: 1) to constrain the factors controlling fracture  
108 occurrence in relation with the host-rock properties, 2) to define the relative timing of fracture  
109 opening in the framework of the burial and geodynamic history of the WCSB, and 3) to assess the  
110 origin of the circulating paleo-waters precipitating the fracture-sealing calcites, in order to gain  
111 insight on the fluid system openness through time.

112

## 113 2. GEOLOGICAL SETTING

114 The WCSB is a complex, polyphase basin system the evolution of which includes a succession  
115 of rift and passive-margin basins, intracratonic basins, and foreland basin (Mossop and Shetsen,  
116 1994). A first collisional pro-foreland developed during the Permian, that was followed by a second  
117 stage of foreland (retro-foreland) as early as the the Triassic (Ferri and Zonneveld, 2008; Golding et  
118 al., 2015; Rohais et al., 2018). Then, eastward subduction of the Farallon plate and subsequent  
119 collision of terranes along the western margin of North America during the Middle-Late Jurassic gave  
120 birth to the North American Cordillera and its associated retro-foreland basin (Price, 1994; Gillespie  
121 and Heller, 1995; DeCelles, 2004; Miall et al., 2008; Fuentes et al., 2011). Sediments were deposited  
122 in this retro-foreland basin until the Eocene (Dawson et al., 1994), forming an eastward- thinning



123 wedge that is ~5–6 km thick in front of the present-day Canadian Rocky Mountains fold-and-thrust  
124 belt, and pinches out onto the Canadian Shield (Price, 1994; Wright et al., 1994; Tufano and Pietras,  
125 2017).

126 The history of the WCSB has a close link to the tectonic evolution of the nearby Rocky  
127 Mountains. On the basis of radiometric age dating of gouges from major thrusts located in the Rocky  
128 Mountains and Foothills of Alberta, Pană and van der Pluijm (2015) showed that the eastward  
129 propagation of the southern Canadian Rocky Mountains fold-and-thrust belt occurred in sequence  
130 through four orogenic pulses (separated by periods of tectonic quiescence) that correspond to  
131 depositional patterns in the adjacent foreland basin. The development of the first clastic wedge in  
132 the basin was partly coeval with the initiation of the westernmost thrusts that moved during a first  
133 pulse (Late Jurassic, 163-145 Ma) in relation to the onset of thin-skinned deformation from Jurassic  
134 terrane accretion. Development of Cenomanian deltaic deposits in the basin was contemporaneous  
135 with a second thrusting pulse (mid-Cretaceous, 103-99 Ma), and the last major transgression in the  
136 southern Alberta foreland was related to a third pulse of tectonic loading (Late Cretaceous, 74-75  
137 Ma). The outermost thrusts to the east (e.g. the McConnell thrust) moved during the last pulse (early  
138 Eocene, 52-54 Ma) and witnessed the last phase of regional contraction which triggered the  
139 accumulation of large volumes of entirely continental deposits in the foreland basin. These thrusting  
140 events presumably occurred under a NE-SW to ENE-WSW direction of the maximum principal stress  
141  $\sigma_1$  associated to a compressional or transpressional stress regime, as revealed by paleostress  
142 analyses in the Rocky Mountains fold-and-thrust belt at the latitude of Calgary (Vandeginste et al.  
143 2012). The stress regime later changed to extensional/transensional in type during the Eocene.

144 Within the WCSB, the Lower–Middle Triassic deposits of interest belong to the Montney-Doig  
145 Fms (Fig. 1) and consist primarily of fine-grained sandstones, siltstones and shales with locally  
146 occurring sandy, bioclastic packstone and grainstone beds that were accumulated in a large variety of  
147 marine (deep turbidite) to marginal-marine environments (Armitage 1962; Davies 1997; Davies et al.  
148 1997; Zonneveld et al. 1997, 2001; Dixon 2000; Orchard and Zonneveld 2009; Crombez et al., 2016;

149 Moslow et al., 2016, 2018; Furlong et al. 2018a; Davies et al., 2018; Euzen et al., 2018 ; Zonneveld  
150 and Moslow, 2018). This sedimentary succession was initially sub-divided into three formations, from  
151 older to younger: the Montney, Doig and Halfway, with a recent update between the Montney and  
152 Doig Fms corresponding to the localized Sunset Prairie Fm (Furlong et al. 2018b). The Lower-Middle  
153 Triassic deposits are preserved in between the Permian Belloy Fm and the Late Triassic Charlie Lake  
154 Fm or the regional pre-Cretaceous erosional unconformity. The Lower–Middle Triassic deposits were  
155 affected by reactivation of inherited structures from both the Paleozoic Peace River Arch (PRA)  
156 collapse, and the older Hay River Shear Zone (HRSZ) (Fig.1; Davies, 1997; Peirce et al., 2001).

157 The synthetic stratigraphic architecture of the Montney-Doig interval (Crombez et al., 2019)  
158 is presented in figure 2. Four sequence boundaries are presented, delimiting four stratigraphic  
159 sequences. The first three (sequences 1–3) belong the Montney Fm, and the last (sequence 4)  
160 encompasses the Doig and Halfway Fms. This study is mainly focused on the three third order  
161 sequences of the Montney Fm (sequences 1-3), and on the basal part of the sequence 4 including the  
162 Doig Phosphatic Member (Doig Fm).

163 Triassic source rocks only reach maturity in the western part of the basin and become  
164 overmature close to the Cordillera deformation front. The peak HC generation ranges from  
165 approximately 100 Ma (mid-Albian) to 60 Ma (Paleocene) for the Doig Fm based on 1D basin  
166 modeling (Ness, 2001). According to Ness (2001), the timing of HC generation for the Montney Fm is  
167 more complex with possible early generation between 120 and 130 Ma, and additional later oil and  
168 gas generation during burial at approximately 100 to 110 Ma. At basin scale, secondary gas  
169 generation probably started at 90-100 Ma and at about 70 Ma in the studied sector (Ness, 2001).  
170 More recently, Ducros et al. (2017), based on 2D petroleum basin modeling, suggested that the  
171 Montney Fm started to generate hydrocarbon in this western part of the basin during the Late  
172 Cretaceous (88,5 Ma). These authors also suggested that the Montney Fm reached the maximum

173 transformation rate just before reaching the maximum burial, which was followed by a major  
174 exhumation and erosional event starting in the Early Paleogene.

175

### 176 3. MATERIAL AND METHODS

177 Two non-oriented well cores from British Columbia (ca. 70 Km North-West of Fort St. John;  
178 Fig. 1A) intercepting the Montney and Doig Fms were investigated. They correspond to 00/16-17-  
179 083-25W6/0 and to 00/12-36-083-25W6/0, vertical wells (BC Oil & Gas Commission), hereafter  
180 referred as 16-17 and 12-36, respectively. The wells are located ca. 70 km North-West of Fort St.  
181 John and between the HRSZ and the inherited structure from the Paleozoic PRA collapse (Fig. 1A).  
182 They are located 7.2 km from each other and based on available basin models (see later in this  
183 section) have experienced a similar burial and thermal history. In particular, the 16-17 continuous  
184 well core includes the entire section of the Montney Fm (sequences 1, 2 and 3; Crombez et al., 2016)  
185 and the basal part of the Doig Fm, whereas the 12-36 well core is composed of three discontinuous  
186 cores of the Montney Fm only for which mineralogical data for two of them (core 1 and core 3) are  
187 available from the British Columbia Oil & Gas Commission public database.

188 Sedimentological logs for the two well cores were described at the 1:50 scale by some of the  
189 co-authors in the frame of a PhD thesis (Crombez, 2016). Wire-log data for the two well cores were  
190 available from the British Columbia Oil & Gas Commission. Investigated well core depth goes from  
191 2531 to 2233 m for well 16-17 and from 2369 to 2088 m for well 12-36. Lithology, sedimentary  
192 structures, textures and paleontological contents allowed characterizing different macro-facies.  
193 Occurrence and orientation of fractures was also reported on the logs (Fig. 3). The majority of the  
194 observed fractures are cemented (i.e. veins), though a few non-mineralized fractures also occur.

195 Descriptive properties derived from macroscopic observations (i.e. facies, depositional  
196 environment, occurrence of differently oriented fractures) and analyses (i.e. TOC) of the two studied

197 well cores were classed and converted into numerical codes (Table 1). A continuous sampling along  
198 the two logs was done and Excel spreadsheets were completed with the evolution of the different  
199 properties with depth (Table S3a and Table S3b of the Supplementary Material). A total of 232 and  
200 576 points were sampled on the logs for well cores 12-36 and 16-17, respectively. These  
201 spreadsheets served as input in the EasyTrace™ software in order to convert the original descriptive  
202 information into a quantitative dataset to achieve frequency distribution analysis of the different  
203 properties (see details on this approach in Gasparrini et al., 2017). Using EasyTrace™ multivariate  
204 statistical analysis with the compiled spreadsheets enabled us to highlight the link between the type  
205 of depositional facies (with their TOC) and the occurrence and orientation of the fractures, as well as  
206 the fracture distribution within the different stratigraphic sequences of the studied formations.

207 Fifty samples, mostly containing mineralized fractures were collected from the two studied  
208 wells (23 from 16-17 and 27 from 12-36). The samples were used to make thin polished and thick  
209 double-polished sections for petrographic, geochemical and fluid inclusion studies. The vein host-  
210 rock from these samples was milled and analyzed with Rock-Eval 6 pyrolysis. Here follow the details  
211 on the different methodologies and apparatus used.

212 Thirty-seven polished thin sections (30-35  $\mu\text{m}$  thick) were prepared for conventional and  
213 cathodoluminescence (CL) microscopy, as well as for staining. Thin section petrography performed  
214 on a Nikon ECLIPSE LV100 POL equipped with a mercury vapor lamp (100 W) allowing UV-light  
215 observations. Calcite crystal habitus and crystal spatial arrangement (texture) in the fractures were  
216 described using the terminology suggested by Bons et al. (2012) and Woodcock et al. (2007). For  
217 samples affected by dolomitization the dolomite texture classification from Sibley and Gregg (1987)  
218 was used. The cold-cathode CL device used is a 8200 Mk5-2 (CITL) which was employed under  
219 vacuum ( $<0.1$  mBar) and with 250  $\mu\text{A}$  and 10 kV operating conditions. A portion of all thin sections  
220 was stained with a solution composed of diluted HCl (10%) with Alizarin red –S and potassium

221 ferricyanide to distinguish calcite from dolomite and qualitatively estimate their Fe content (Dickson,  
222 1966).

223         Seven double-polished thick sections (100-120  $\mu\text{m}$  thick) were prepared for Fluid Inclusion  
224 (FI) petrography and microthermometry measurements. Different FI assemblages were distinguished  
225 based on their location within the crystals, e.g. crystal cores, growth zones, along trails, isolated. For  
226 bi-phase FIs, the volumetric proportion of the liquid phase relative to the total volume of the FI,  
227 referred as degree of fill (F), was calculated at room temperature from 2D screen images by  
228 measuring areas. Microthermometry was carried out with a Linkam MDS 600 stage mounted on a  
229 Nikon LV100 Eclipse, with a 100 W Mercury vapor lamp which allowed UV-light observations. The  
230 stage was calibrated with synthetic FIs in the temperature range  $-56.6$  to  $+135^\circ\text{C}$ . The Linksys 32  
231 software enabled all the operations for FI microthermometry. The measurement accuracy was of 1  
232 and  $0.2^\circ\text{C}$  for heating and cooling runs, respectively. For aqueous fluids, heating runs were  
233 accomplished before cooling runs. Salinities were calculated from final melting of ice ( $T_{\text{mi}}$ ) in the  
234 binary  $\text{H}_2\text{O}$ - $\text{NaCl}$  system (Bodnar, 1993). Oil density was qualitatively estimated from fluorescence  
235 color of hydrocarbon FIs based on the chart from McLimans (1987).

236         Carbonate (calcite) powders from the targeted veins were extracted from polished rock slabs  
237 by means of a dental drill to be analyzed for oxygen (O), carbon (C) and strontium (Sr) isotopes.

238         Forty-one calcite powders were analyzed for O and C isotopes. The samples were reacted  
239 with 100% phosphoric acid at  $70^\circ\text{C}$  up to 7 hours. A Gasbench II connected to a Thermo Finnigan Five  
240 Plus mass spectrometer was used. All values for carbonate phases are reported in per mil relative to  
241 V-PDB by assigning a  $\delta^{18}\text{O}$  value of  $-2.20$  ‰ and a  $\delta^{13}\text{C}$  value of  $+1.95$  ‰ to the NBS19 standard.  
242 Reproducibility was checked by replicate analysis and is better than  $\pm 0.07$  ‰ for  $\delta^{18}\text{O}$  and  $\pm 0.06$  ‰  
243 for  $\delta^{13}\text{C}$  (1 std. dev).

244         Eleven calcite powders were also analyzed for Sr isotopes. Carbonate samples were leached  
245 in ammonium acetate to remove groundwater salts and displace contaminant Sr on exchangeable

246 sites (e.g. Bailey et al., 2000). The remaining material was rinsed twice in deionized water, and then  
247 dissolved in dilute HCl. After ion exchange chemistry, samples were loaded onto purified Re filaments  
248 in a Ta emitter solution (Birck, 1986). Isotopic analyses were made on a VG-Sector-54 thermal  
249 ionization mass spectrometer using a three cycle dynamic multicollector routine and an exponential  
250 mass fractionation correction relative to  $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$  (e.g. Hans et al., 2013). Repeated  
251 measurements of reference material NBS987 at similar run conditions during the period over which  
252 the analyses were made yielded a value indicating that the measurement repeatability is  
253 commensurate with the within-run uncertainty.

254 The Rock-Eval 6 pyrolysis apparatus was used to calculate, among other parameters, the  
255 TOC, the Hydrogen Index (HI) and the Mineral Carbon (MinC) of 46 fracture host-rock samples. The  
256 Shale Play Method<sup>®</sup> developed at IFP Energies nouvelles (Pillot et al., 2014; Romero-Sarmiento et al.  
257 2016a, 2016b) was used on bulk-rock samples, allowing for a better estimate of the free or absorbed  
258 hydrocarbons in the rocks. 50-70 mg of powdered sample is heated in an open pyrolysis system  
259 under non-isothermal condition (from 300 to 650°C). During this pyrolysis, the amount of  
260 hydrocarbons released is measured by a flame ionization detector (FID) and CO and CO<sub>2</sub> release are  
261 monitored with an infrared (IR) detector. The residual sample is then put in an oxidation oven where  
262 it is heated (from 300 to 800°C) under artificial air (N<sub>2</sub>/O<sub>2</sub>: 80/20). During combustion, the amount of  
263 CO and CO<sub>2</sub> released are monitored with an IR detector.

264 In addition, among the 46 bulk-rock samples analyzed, 10 samples were selected for organic  
265 solvent extraction. The powdered samples were placed in a solution of dichloromethane and  
266 methanol (1/1) in an ultrasonic bath at 40°C for 30 minutes. The samples were then filtered and  
267 placed in a drying oven for one day. Finally, the samples of extracted organic matter were analyzed  
268 again by using the Shale Play Method<sup>®</sup> of the Rock-Eval 6 apparatus described above.

269 A 3D basin model under TemisFlow<sup>™</sup> software available in-house was used to estimate the  
270 thermo-barometric evolution of the studied stratigraphic interval with time. In this basin model a

271 constant heat flow of 47 mW/m<sup>2</sup>, and then 57 mW/m<sup>2</sup> was imposed at the base of the sedimentary  
272 pile, based on a lithospheric model, and thermal calibration was achieved by making use of vitrinite  
273 reflectance data to test the two heating scenarios (Ness 2001; Ducros et al., 2017). This model takes  
274 into account the amplitude of the Tertiary uplift, with an estimated thickness of eroded sediments  
275 that ranges from 4000 m close to the Rocky Mountains (Roure et al., 2010) to less than 400 m near  
276 the Precambrian shield in Saskatchewan (Ness, 2001; Higley et al., 2005; Roure et al., 2010).

277

## 278 4. RESULTS

279 The main results from well core logging, facies analysis and stratigraphy are reported in  
280 figure 3 together with the location of the studied samples and their TOC and MinC from Rock-Eval  
281 pyrolysis. Table S1a and Table S1b (see Supplementary Material) summarize details on the samples  
282 collected from wells 16-17 and 12-36, respectively, together with the sample depth, sequence of  
283 provenance, fracture orientation and kinematic aperture, petrography of the calcite sealing cements  
284 and their O-C isotope composition. Table S2a and Table S2b (see Supplementary Material) report the  
285 results of Rock-Eval pyrolysis for the rock samples hosting the fractures (on bulk-rock and extracted  
286 organic matter, respectively) for the two well cores investigated. Finally, Table S3a and Table S3b  
287 (see Supplementary Material) report the data-sheets with the evolution of the different properties  
288 with depth, used for frequency distribution analysis with Easytrace™ software.

289

### 290 4.1 Sedimentology

291 In this contribution, we primarily refer to detailed descriptions of the sedimentary facies  
292 previously published (Crombez et al., 2016; Playter et al., 2018). The two studied wells are located in  
293 the distal area of the WCSB where a narrow range of grain size and a dolomitic-feldspathic  
294 composition is expected to dominate the Montney lithologies, as due to arid climate conditions

295 during deposition (e.g. Euzen et al., in press; Zonneveld and Moslow, 2018). They contain relatively  
296 hard grains (e.g. quartz, feldspars, carbonates), moderate amounts of organic matter and, locally, a  
297 relatively high content of clay minerals (e.g. Euzen et al., 2018; Vaisblat, 2020).

298 Petrographic observations on thin sections indicate that the investigated samples have grain  
299 size from very fine sandstone to fine siltstone, locally with clays. In particular, the Montney samples  
300 are often characterized by consistent grain size with silt-sized particles of 10-25  $\mu\text{m}$  being dominant,  
301 locally with very fine sands (<80  $\mu\text{m}$ ), whereas the more clay-rich intervals could not be further  
302 characterized due to the limited resolution of the conventional optical microscopy used. On the  
303 other side, QEMSCAN images available for part of well 12-36 (British Columbia Oil and Gas  
304 Commission public database) suggest that the grain size mostly ranges between 15 and 30  $\mu\text{m}$ . Both  
305 these datasets allow to consider the studied lithologies as mudrocks, according to the definition of  
306 Lazar et al. (2015) which includes silt-sized particles (fine mud<8 $\mu\text{m}$ , medium mud=8-32 $\mu\text{m}$  and  
307 "coarse mud"=32-64 $\mu\text{m}$ ). In this contribution, we primarily refer to detailed descriptions of the  
308 sedimentary facies previously published (Crombez et al., 2016; Playter et al., 2018). Ten facies  
309 associations were identified (Fig. 3 and 4) by conducting a higher resolution lithology and facies  
310 analysis combined with the microfacies analysis from thin section petrography : 1. Lag; 2. Massive  
311 fine-grained sandstones; 3. Massive very fine-grained sandstones to siltstones; 4. Sand-dominated  
312 alternation of very fine-grained sandstones to siltstones; 5. Silt-dominated alternation of very fine-  
313 grained sandstones to siltstones; 6. Massive siltstones; 7. Organic-rich massive siltstones; 8. Bioclastic  
314 siltstones; 9. Laminated siltstones with mm-thick sand laminae (low density turbidite); 10. Laminated  
315 siltstones.

316 The ten facies associations identified were then grouped into four main depositional  
317 environments: (1) lag, (2) shoreface, (3) offshore transition, (4) offshore. Given the distal position of  
318 the wells in the basin, the very proximal depositional environments (e.g. foreshore or tidal) were not



319 encountered. Lag deposits do not strictly represent a depositional environment since they are rather  
320 related to erosional and reworking processes influenced either by wave-, or storm-action.

321

#### 322 **4.2 Rock-Eval pyrolysis**

323 The Montney and Doig Fms contain organic matter of type II and III (Crombez et al., 2016;  
324 Riediger, 1997; Romero-Sarmiento et al., 2016b). The mean TOC is in the range 0.5% - 4wt% and up  
325 to 8.2wt% TOC in the Montney Fm, and may reach 11 % in the Doig Fm within the Doig Phosphatic  
326 Member at the base (Riediger et al., 1990a; Ibrahimbas and Riediger, 2004; Crombez et al., 2016;  
327 Romero-Sarmiento et al., 2016b).

328 Table 2 illustrates the mean TOC and HI values for the 46 fracture host-rock samples analyzed  
329 in this study by stratigraphic sequence and by well . The complete dataset acquired in the present  
330 study by Rock-Eval analysis is reported in Table S2a and Table S2b (see Supplementary Material).

331 The studied 46 samples have TOC between 0.56 and 6.65 % (mean is 1.74%) with only 13% of  
332 the samples having TOC above 3.00%. In particular, a mean TOC of 1.68% is recorded from the 26  
333 samples of the 12-36 well, whereas a mean of 1.91% is recorded from the 20 samples of the 16-17  
334 well; the highest value of 6.65% is recorded in the 16-17 well at the TST of sequence 4, corresponding  
335 to the Doig Phosphatic Member.

336 All samples analyzed here display HI below 50 mgHC/gTOC, pointing towards a low  
337 hydrocarbon potential, possibly due to the fact that maturation in the studied intervals has already  
338 occurred.

339 The pyrolysis results obtained from the 10 samples analyzed twice (i.e. on bulk-rock and on  
340 extracted organic matter) were compared. The procedure of organic matter extraction eliminated all  
341 the light hydrocarbons (free and adsorbed) and part of the heavy and very heavy hydrocarbons. The

342 TOC values obtained on bulk-rock and extracted organic matter are similar (see Table S2a and Table  
343 S2b in Supplementary Material), suggesting that only the carbon from organic matter contributes to  
344 the TOC computation. This ensures the reliability of the TOC values from the remaining 36 samples  
345 which were analyzed on the bulk rock only.

346 Crombez et al. (2016) previously published TOC and MinC values from continuously sampled  
347 cores and cuttings (155 samples) of the same wells. Measurements were done by means of the same  
348 Rock-Eval 6 apparatus, though by using the classic method (Lafargue et al., 1998; Pillot et al., 2014).  
349 TOC values measured by these previous authors were compared to those measured on close samples  
350 during this study for a quality check; the values from the two independently obtained datasets  
351 mostly overlap, the TOC difference being within the uncertainty range. This also confirms that the  
352 classic and the Shale Play Method® of the Rock-Eval 6 apparatus lead to equivalent TOC computation  
353 (as already underlined by Romero-Sarmiento et al., 2016a, 2016b).

354 Figure 5 reports the frequency distribution of the available TOC data (in wt. %) by comparing the  
355 values measured in this study (on fracture host-rock samples only) with the full dataset by also  
356 including the values measured by Crombez et al. (2016). The figure illustrates that the two TOC  
357 datasets have an overlapping distribution, the only minor difference being that in our dataset  
358 samples with TOC of 1-2 wt. % are more represented. This suggests that the TOC data at fractures  
359 here produced is representative of the whole TOC dataset available for the studied well cores.

### 360 **4.3 Host-rock diagenesis**

361 Thin section optical petrography combined with CL observations of the fracture host-rocks  
362 revealed that several diagenetic events affected the studied formations after deposition. The study  
363 was focused on the carbonate diagenetic phases occurring as cements or replacements, therefore  
364 potentially affecting the rock mechanical properties. Here follow the details on the main diagenetic  
365 phases observed.

366 A first calcite (C1) may pervasively cement the most proximal (bioclastic) samples by filling  
367 the intergranular pores between bioclasts. It is composed of non-ferroan calcite crystals (10-60  $\mu\text{m}$ )  
368 displaying a uniform and bright orange response under CL. It is rarely observed and is considered of  
369 minor extent in the studied wells.

370 A first dolomite (D1) may totally or partially replace the detrital carbonate grains. It consists  
371 of non-ferroan euhedral to sub-hedral crystals (10-30  $\mu\text{m}$ ), building up a planar-E texture and  
372 displaying a uniform to zoned dull orange to dull red CL (Fig. 6A).

373 These first two phases (C1 and D1) are interpreted to have precipitated during eogenesis (i.e.  
374 early diagenesis) because of their non-ferroan nature. Furthermore, samples affected by C1  
375 cementation lack major mechanical compaction features, suggesting that C1 precipitation occurred  
376 before significant burial.

377 Mechanical compaction is witnessed in many samples by micro-fracturing of bioclasts, as well  
378 as by the presence of grain to grain (locally interpenetrated) contacts .

379 A second dolomite (D2) mainly occurs as a cement since it forms overgrowths (5-10  $\mu\text{m}$  thick)  
380 around detrital carbonate grains commonly replaced by D1 (Fig. 6B), although locally it also replaces  
381 the carbonate grains.. It is composed of ferroan dolomite crystals showing a uniform to zoned bright  
382 red CL (Fig. 6B).

383 A third dolomite (D3) may totally replace specific bioclastslike the calcispheres (Fig. 6C). It  
384 consists of ferroan anhedral crystals (10-50  $\mu\text{m}$ ), which make up a non-planar texture, do not  
385 luminesce under CL (Fig. 6D) and acquire a deep blue color after staining.

386 The ferroan nature of these latter two dolomite phases (D2 and D3) points towards a  
387 precipitation during mesogenesis, in the burial environment, where reduced  $\text{Fe}^{2+}$  may commonly be  
388 available for uptake into the carbonate lattice upon crystallization.

389 A second calcite (C2) phase mainly occurs as a cement within intergranular pores. It consists  
390 of non-ferroan granular to blocky crystals (10-50  $\mu\text{m}$ ) with a uniform bright orange luminescence. It  
391 possibly precipitated after major mechanical compaction as suggested by the presence of  
392 compaction features (e.g. grain to grain interpenetrated contacts) in samples pervasively affected by  
393 C2 cementation. C2 may also occur as a replacement of previously precipitated carbonate phases  
394 (Fig. 6E and F).

395 Pyrite occurs in all samples both as small framboids (<5-10  $\mu\text{m}$ ) or in large euhedral crystals  
396 (10-50  $\mu\text{m}$ ). It replaces mainly the bioclasts and the early diagenetic calcites but also the third  
397 (ferroan) dolomite occurring within calcispheres.

398 Some of the above diagenetic phases may preferentially occur in facies from one or more  
399 depositional environments. For instance, the first dolomite (D1; eogenetic replacement) is observed  
400 only in samples from the 16-17 well, mainly from the shoreface environment and secondarily from  
401 the offshore transition environment. Furthermore, the third dolomite (D3; burial replacement)  
402 always occurs in facies containing calcispheres which are dominantly found in samples from the  
403 offshore transition and the offshore environments. The other diagenetic phases described above do  
404 not seem to be facies specific and are observed in similar proportions in all of the identified  
405 depositional environments.

406 PDiagenesis studies on the Montney Fm were conducted by previous authors (Vaisblat et  
407 al.,2017b, Vaisblat 2020). Vaisblat et al. (2017b) investigated 83 samples coming from two wells  
408 located in British Columbia for petrography (SEM), mineralogy (XRD) and geochemistry (QEMSCAN).  
409 In particular, 68 samples out of the 83 come from well 16-17, also investigated in the present study.  
410 Our observations (though based on optical petrography and CL only) were therefore compared with  
411 those of Vaisblat et al. (2017b); the main differences are illustrated in figure 7. This comparison  
412 reveals that in the present study authigenic clay minerals, quartz cement and feldspar  
413 dissolution/precipitation were not observed, due to the limited resolution of the conventional optical

414 microscopy. On the other hand, some carbonate phases were here documented for the first time  
415 (Fig. 6); these are the non-ferroan first calcite (C1) and first dolomite (D1) phases, precipitated during  
416 eogenesis and the ferroan third dolomite (D3), precipitated during burial within calcispheres.

417 Interestingly, no diagenetic sulfates were detected in the Montney-Doig lithologies of the  
418 studied well cores, in contrast with the observations made in well cores located further to the south  
419 (Liseroudi et al., 2020), i.e. within the inherited structure from the Paleozoic PRAcollapse (Fig. 1A).

420

#### 421 **4.4 Fracture orientation and abundance**

422 All the fractures (both mineralized and non-mineralized) observed along the 16-17 and 12-36  
423 vertical well cores were reported on the sedimentological logs. Macroscopically, the fractures were  
424 classified according to their orientation (Fig. 8): horizontal (bedding-parallel) or vertical (vertical s.s.  
425 or at high angle to bedding). The persistence of the vertical to high angle fractures mostly ranges  
426 from 3 to 10 cm and only rarely exceeds 20 cm; the persistence of the horizontal fractures could not  
427 be estimated since it exceeded the core width.

428 For the properties recorded on the sedimentological logs (i.e. facies associations,  
429 depositional environments, stratigraphic sequences, occurrence and orientation of the fractures) and  
430 for those measured in the laboratory (i.e. TOC) different classes were established and expressed by  
431 means of numerical codes (Table 1).

432 The complete spreadsheets used for statistical treatment with EasyTrace™ software are  
433 reported in Table S3a and Table S3b (see Supplementary Material) for wells 16-17 and 12-36,  
434 respectively. The treatment of these data highlighted different possible correlations between the  
435 abundance and orientation of the fractures and the host-rock features as well as their distribution  
436 within the four stratigraphic sequences (Fig. 9).

437 The 10 facies associations (property here referred to as FACIES) previously identified (Fig. 3  
438 and 4) were grouped into four broad depositional environments (property here referred as  
439 ENVIRONMENT) as follows: the lag (code 1), the shoreface (code 2) which groups the facies  
440 associations 2 and 3, the offshore transition (code 3) which groups the facies associations 4 and 5  
441 and the offshore (code 4) which groups the facies associations 6, 7, 8, 9 and 10 (see Table 1).

442 The four stratigraphic sequences (property here referred to as SEQUENCE) of the studied  
443 formations were classed from 1 to 4 (Table 1) as follows: sequence 1 (code 1), sequence 2 (code 2),  
444 sequence 3 (code 3), sequence 4 (code 4).

445 The TOC contents (property here referred as TOC) from fracture host-rock samples analyzed  
446 in this study (Table 2 and Fig. 5) were classed from 1 to 6 (Table 1) as follows: <1% (code 1), 1-2%  
447 (code 2), 2-3% (code 3), 3-4% (code 4), 4-5% (code 5), 5-11% (code 6).

448 Finally, the presence/absence of fractures and their orientation along the logs (property here  
449 referred as FRACTURE) were classed by means of 3 codes (Table 1) as follows: no fracture occurrence  
450 (code 1), occurrence of horizontal fractures (code 2), occurrence of vertical to high angle fractures  
451 (code 3).

452 Figure 9 illustrates the results of the statistical treatment with EasyTrace™ software for the  
453 two wells. The red lines represent the distribution of the whole sampled points where fractures  
454 occur (i.e. only the points with codes 2 and 3 of the property FRACTURE). Figure 9A and 9B show  
455 respectively the distribution of sampled points characterized by the presence of horizontal and  
456 vertical to high angle fractures; the black bars indicate in which SEQUENCE, ENVIRONMENT and TOC  
457 classes fall the sampled points corresponding to fracture occurrence. The values reported as “777”  
458 correspond to undefined, though not null, values that were indeed attributed to cells of the log  
459 spreadsheets when information on a given property was not available. This is because at the same  
460 well depth information may not be available for all properties (see Table S3a and Table S3b of the  
461 Supplementary Material).

462 Figure 9A indicates that the horizontal fractures (code 2 of the property FRACTURE)are  
463 similarly abundant in 3 of the 4 stratigraphic sequences (property SEQUENCE) and are rare only in  
464 sequence 3. They occur in nearly all facies associations (property FACIES). In particular, the massive  
465 siltstones (code 6) host 28% of the fractures, whereas the bioclastic siltstones (code 8) and the silt-  
466 dominated alternation (code 5) host respectively 7 and 5% of the fractures. Globally, the depositional  
467 environment (property ENVIRONMENT) named offshore (code 4) hosts the majority of horizontal  
468 fractures. No relationship between the abundance of the horizontal fractures and the host-rock TOC  
469 could be highlighted.

470 Figure 9B indicates that the vertical to high angle fractures (code 3 of the property  
471 FRACTURE)are similarly abundant in 3 of the 4 stratigraphic sequences (property SEQUENCE) and are  
472 fairly less represented in sequence 3. They occur in almost all facies associations (property FACIES). In  
473 particular, the sand-dominated alternation (code 4) hosts 29% of the fractures, whereas the massive  
474 fine-grained sandstones (code 2) and the bioclastic siltstones (code 8) host respectively 4 and 3% of  
475 the fractures. Globally, the depositional environment (property ENVIRONMENT) named offshore  
476 transition (code 3), characterized by sand-rich facies, hosts the majority of the vertical fractures.  
477 These fractures seem to occur mainly in rocks with low TOC, i.e. between 1 and 2% (code 2 of the  
478 property TOC).

479

#### 480 **4.5 Vein petrography**

481 All the investigated mineralized fractures (i.e. horizontal and vertical to high angle veins) are  
482 fully sealed by calcite cements, leaving no fracture porosity left. Despite the monotonous mineralogy,  
483 the calcite cements display various crystalline habitus (e.g. fibrous, blocky and elongated blocky) and  
484 spatial arrangement (i.e. texture) within the veins. Details on the petrographic features of all the  
485 studied samples (including the fractures and the sealing calcite cements) can be found in Table S1a

486 and Table S1b (see Supplementary Material). The nomenclature used is after Woodcock et al. (2007)  
487 and Bons et al. (2012).

488 The horizontal, bedding-parallel veins are sealed by fibrous, blocky or elongated blocky  
489 crystals (Fig. 10). Commonly, different crystal habitus may occur in the same vein, where for instance  
490 the crystals evolve from the vein wall to the vein center from fibrous to elongated blocky or from  
491 elongated blocky to blocky. In all these cases, the crystals have syntaxial (both symmetrical or  
492 asymmetrical) or antitaxial arrangements suggesting that calcite precipitation occurred during  
493 fracture opening (syn-kinematic calcite). Features suggesting a crack-seal mechanism for vein  
494 opening are common; we could observe several bands of calcite oriented parallel to the vein walls  
495 which represent repeated opening events progressively sealed by calcite (Ramsay, 1980; Laubach,  
496 2003). The calcite cement may therefore be considered as syn-kinematic, since crack-seal textures  
497 form as cement precipitates during progressive vein widening. Our observations also suggest that in  
498 most cases, the horizontal veins opened in mode I, even though we could observe in very few cases  
499 evidence for mixed mode I-mode II opening (Fig. 10C and D).

500 All the vertical to high angle fractures are sealed by blocky to elongated blocky calcite crystals  
501 (Fig. 10). In the first case a drusy texture may be observed with the crystal size increasing towards the  
502 vein center, whereas in the second case the crystals have a syntaxial arrangement (both symmetrical  
503 and asymmetrical). Features corresponding to crack-seal mechanism of vein opening have been only  
504 locally observed. More commonly the vein terminations display tortuous forms, with axial planes  
505 lying broadly in the plane of bedding. In most cases also the vertical veins opened in mode I (Fig. 10A  
506 and B).

507 All the veins show twinned calcite crystals (Fig. 10). Most twins are thin (thickness < 5  $\mu\text{m}$ )  
508 and rectilinear, suggesting low strain (a few percent) and deformation at temperature below 170-  
509 200°C (Ferrill et al., 2004; Lacombe, 2010).



510 Interestingly, all calcite cements sealing the fractures, irrespective of the well core of  
511 belonging, fracture orientation and crystal habitus/texture, display, when observed under UV-light,  
512 the same response given by uniform and very dull green fluorescence. Similarly, the calcite cements  
513 cannot be distinguished when observed under CL as they all show a relatively uniform dull to very  
514 dull orange color.

515 Cross-cutting and abutting relationships between horizontal and vertical to high angle veins  
516 were only locally observed, therefore the vein relative chronology is not easily assessed from  
517 petrography only. However, where cross-cutting relationships are observed, they indicate that some  
518 of the vertical veins post-date the horizontal ones since the calcite crystals of the vertical vein replace  
519 the calcite crystals of the horizontal one at the vein intersection (Fig. 10G and H).

520

#### 521 **4.6 O-C-Sr isotope geochemistry**

522 The results of the O-C stable isotope analyses performed on the vein calcite cements are  
523 reported in figure 11. The  $\delta^{18}\text{O}$  ranges between -8.27 to -11.18‰ (mean -10.45‰), whereas  $\delta^{13}\text{C}$   
524 ranges from 1.34 to -4.38‰ (mean -1.09‰). When separating the calcite samples per well of  
525 belonging it comes that calcite from well 16-17 has  $\delta^{18}\text{O}$  ranging from -8.27 to -10.7‰ (mean -  
526 9.69‰) and  $\delta^{13}\text{C}$  ranging from 1.04 to -4.3‰ (mean -1.86‰), whereas calcite from well 12-36 has  
527  $\delta^{18}\text{O}$  varying from -10.72 to -11.18‰ (mean -10.72‰) and  $\delta^{13}\text{C}$  varying from 1.34 to -3.43 (mean -  
528 0.82‰).

529 Analyzed calcites display a very negative  $\delta^{18}\text{O}$  compared to calcite precipitated in equilibrium  
530 with Early Triassic seawater (Fig. 11; Veizer et al., 1999). Based on the well-established dependency  
531 of O isotope fractionation on temperature (O'Neil, 1969), such negative  $\delta^{18}\text{O}$  values suggest calcite  
532 precipitation at burial temperatures.

533 The slight  $\delta^{13}\text{C}$  variability of the calcites from the two wells encompasses a comparable range  
534 (~4‰) and may be due to the contribution of different amounts of light  $^{12}\text{C}$  from the host-rocks to  
535 the calcite parent fluids. Indeed, in well 12-36, the most negative  $\delta^{13}\text{C}$  values are recorded by calcites  
536 hosted in sequence 3 (Fig. 11A), which is characterized by the highest mean TOC among the three  
537 sequences recorded in this well (Table 2). Similarly, in well 16-17, the most negative  $\delta^{13}\text{C}$  values are  
538 recorded in samples from sequence 4 (Fig. 11B), which records the highest TOC values of the whole  
539 dataset analyzed (Table 2).

540 Calcites from different well cores and differently oriented veins have overlapping Sr isotope  
541 composition (Table S1a and Table S1b of Supplementary Material). Indeed, 10 of the 11 calcites  
542 investigated display  $^{87}\text{Sr}/^{86}\text{Sr}$  falling in the narrow range comprised between 0.7112 and 0.7124, with  
543 only one calcite having lower values (0.7091). Compared with calcites precipitated in equilibrium  
544 with Early to Middle Triassic seawater, which have  $^{87}\text{Sr}/^{86}\text{Sr}$  in the range 0.706940 to 0.708219  
545 (Veizer et al., 1999), all of the fracture calcites investigated result to be enriched in the radiogenic  
546  $^{87}\text{Sr}$  isotope.

547 Overall, these results indicate that the investigated calcites show little variations in isotope  
548 compositions irrespective from crystal habitus and texture, vein orientation and well of belonging.

549

#### 550 **4.7 Fluid inclusion (FI) study**

551 Of the 7 samples prepared for Fluid Inclusion (FI) study only 4 could be investigated for  
552 microthermometry. This is because the three other samples consist of fractures sealed by fibrous  
553 calcite crystals for which microthermometry measurements were unfeasible due to the small size of  
554 the FIs. Four different types of FIs (Type I, Type II, Type III, Type IV) were distinguished  
555 petrographically and may show shapes like negative crystals or controlled by crystallographic planes  
556 (Fig. 12 and Table 3). The main results of FIs microthermometry are reported in Table 3 and Fig. 13.

557 Type I consists of aqueous, bi-phase, liquid-rich FIs with degree of fill (F) between 0.89 and  
558 0.95; they occur in crystal cores, along trails or isolated and do not show fluorescence under UV-light  
559 (Fig. 12A and B). Homogenization occurs in the liquid state at temperatures ( $T_h$ ) between 78 and  
560 118°C with mode value at 100°C. During cooling after homogenization most FIs develop a gas bubble  
561 at temperatures between 88 and 62°C. During further cooling they freeze at temperatures between -  
562 67 and -70°C. During reheating after freezing, on some FIs (n=15) the occurrence of a first liquid (i.e.  
563 the apparent eutectic, which overestimates the real eutectic) was observed optically at temperatures  
564 between -21.9 and -18.4°C, suggesting that the aqueous fluid is dominated by NaCl. Ice is possibly  
565 the last phase to melt (as deduced from the roundish shape and the whitish color of the crystals)  
566 with a stable behavior (gradual melting allowing cycling procedure). Final melting of ice ( $T_{mi}$ ) occurs  
567 between -19.3 and -16.0°C. Fluid salinity, calculated on 15 FIs, varies between 19.4 and 21.9 eq. NaCl  
568 wt% (mode is 20). The salinity of Type I FIs is about six times higher than normal seawater, suggesting  
569 that the calcites precipitated from saline basinal brines. The  $T_h$  measured represent FI minimum  
570 trapping temperatures. To estimate the real trapping temperatures an isochore was constructed  
571 using the mode values of  $T_h$  (100°C) and salinity (20 eq. NaCl wt%) and combined with a thermo-  
572 barometric gradient of 30°C/km. This gradient is consistent with the basin evolution during Mesozoic  
573 time, as deduced from the available TemisFlow<sup>TM</sup> model. A pressure correction of 10°C was evaluated  
574 for Type I FIs pointing to a possible trapping temperature of 110°C.

575 Type II is made of oil, bi-phase, liquid-rich FIs with F between 0.8 and 0.92; they occur in  
576 crystal cores, along trails or isolated and show a green fluorescence under UV-light (Fig. 12C and D).  
577 Homogenization in the liquid state occurs at temperatures ( $T_h$ ) falling within the 24 to 58°C range  
578 with mode value of 31°C. During cooling after homogenization all FIs develop a gas bubble between  
579 48 and 26°C. During cooling runs no solid phases are observed to form.

580 Type III consists of oil, mono-phase FIs with F=1; they occur in crystal cores, isolated, along  
581 growth zones, along trails and show a green fluorescence under UV-light. No phase changes are

582 observed during heating and cooling runs and the FIs remain monophasic. No microthermometry  
583 measurements could be performed.

584 Type IV comprises mono-phase, liquid FIs with  $F=1$ ; they occur in crystal cores, along growth  
585 zones, isolated or along trails. They consistently show shapes from negative crystal to  
586 crystallographic controlled. Some of them display weak yellow to green fluorescence whereas some  
587 others do not show any fluorescence under UV-light (Fig. 12E and F). During cooling from room  
588 temperature, the mono-phase liquid FIs develop a gas bubble at temperatures between  $-60$  and  $-$   
589  $74^{\circ}\text{C}$ . The FIs persist in the bi-phase state down to  $-150^{\circ}\text{C}$  with no observable solid phase being  
590 formed. During reheating the bi-phase FIs homogenize in the liquid state. Homogenization  
591 temperatures ( $T_h$ ) are comprised between  $-76$  and  $-62^{\circ}\text{C}$  with mode value at  $-70^{\circ}\text{C}$ . The behavior of  
592 Type IV FIs suggests they may be carbonic FIs, belonging to the  $\text{CH}_4\text{-CO}_2\text{-N}_2$  fluid system. The behavior  
593 of these fluids depends on their density and on the relative proportions of the different components  
594 (Van den Kerkhof and Thiery, 2001; Conliffe et al., 2017). In particular, for a pure  $\text{CH}_4$  system,  $T_h$   
595 values of  $-80^{\circ}\text{C}$  are expected with  $T_h$  becoming less negative with increasing proportions of  $\text{CO}_2$ .  $T_h$   
596 values measured for Type IV FIs (Table 3) are in line with a system dominated by  $\text{CH}_4$  with some  
597 quantities of  $\text{CO}_2$ , although the presence of other hydrocarbons, such as  $\text{C}_2\text{H}_6$  or  $\text{C}_3\text{H}_8$ , cannot be  
598 excluded. Preliminary characterization of Type IV FIs with Raman spectroscopy also revealed the  
599 presence of  $\text{CH}_4$ , though the high fluorescence of the host calcite mineral did not allow a more  
600 quantitative evaluation.

601

## 602 5. DISCUSSION

603

### 604 5.1 Host-rock controls on fracture occurrence

605 A number of papers have addressed the multiple factors controlling the occurrence of  
606 fractures in siliciclastic tight reservoir rocks (Engelder and Peacock, 2001; Rijken and Cooke, 2001;  
607 Peacock and Mann, 2005; Wang and Gale, 2009; Gale et al. 2014; Wang et al., 2016; Ilgen et al.,  
608 2017; Hooker et al., 2020; Peng et al., 2020). Many factors, such as tectonic stress, lithology,  
609 reservoir thickness, abundance of organic matter, mineralogy and cementation have been recognized  
610 as potentially exerting a control on the brittleness of the rock (i.e. the easiness of a rock to be  
611 fractured and its ability to maintain a fracture).

612 Rock brittleness is a key criterion to identify the most suitable stratigraphic intervals for  
613 hydraulic fracturing and hence, for successful development of unconventional tight oil and gas plays  
614 hosted in fine siliciclastic rocks such as the shales and the siltstones. Present day brittleness can be  
615 evaluated based on either lithology or elastic properties. In general, the higher the Young's modulus  
616 and the lower the Poisson's ratio, the more brittle the rock. For instance, rock mechanical tests  
617 indicate that under the same stress conditions, a shale will have higher brittleness and smaller tensile  
618 strength if it has lower Poisson's ratio and higher Young's modulus (Ding et al., 2012).

619 The composition of the shale exerts a first-order control on brittleness, and therefore on  
620 fracture system development, both natural and stimulated. The content in relatively hard minerals  
621 such as quartz, feldspar and carbonates has been found to be positively correlated with the number  
622 of fractures and/or brittleness of the shale (Perez and Mafurt, 2013; Zeng et al., 2013; Hu et al. 2015;  
623 Labani and Rezaee 2015; Rybacki et al., 2015; Wang et al., 2017). Conversely, a high clay content  
624 makes the shale more ductile (Aoudia et al., 2010, Perez and Mafurt, 2013; Dong et al., 2017, 2018).  
625 The brittleness of a shale can therefore be predicted from the volumetric fraction of clay, quartz,  
626 feldspar and carbonate contents by treating the rock as a composite material and by calculating its  
627 elastic properties by the rule of mixture (Pei et al., 2016), though this approach does not take into  
628 account the effect due to the form of grains (Peng et al., 2020). The influence of TOC and organic  
629 maturity on brittleness of tight siliciclastic rocks is still debated. In the Nunitang Shale the TOC shows  
630 a positive correlation with the number of fractures (Zeng et al., 2013). In the Barnett Shale, Perez and

631 Mafurt (2013) show that contrary to the commonly held understanding, the increased TOC does not  
632 make the rock more ductile, whereas Yasin et al. (2017) report that the TOC content has a positive  
633 correlation with the amount of quartz, and that organic maturity has a positive correlation with  
634 brittleness. Dong et al. (2018) also conclude that in the Duvernay shale, increased organic maturity  
635 results in greater hardness for rocks of similar geochemical compositions. Conversely, other studies  
636 suggest that the effect of organic matter content and maturity on the mechanical properties of a  
637 shale is subordinate to the rock mineral composition (Aoudia et al., 2010; Labani and Rezaee, 2015).

638 In contrast with the well documented literature on shales, the relationship between  
639 composition and mechanical properties in siltstones is not well established. Brittleness of the  
640 Montney siltstones has been calculated from well-log averages (sonic, density, XRF, XRD) calibrated  
641 to static and measured data on cores (Vaisblat et al. 2017a, 2019; Chatellier et al., 2018).

642 The results of the present study derived from the analysis of cored intervals of wells 12-36  
643 and 16-17 which dominantly corresponds to the Montney Fm and to a minor extent to the Doig Fm  
644 (Fig. 3), allow considering the succession investigated as being mostly composed by siltstones.  
645 According to the frequency distribution analysis accomplished (Fig. 9), based on a rather continuous  
646 sampling of the different properties (i.e. SEQUENCE, FACIES, ENVIRONMENT, FRACTURE, TOC) along  
647 the logs of the two well cores it could be concluded that:

648 - The depositional environment named offshore transition (and including facies associations  
649 such as sand-dominated alternation and silt-dominated alternation), chiefly composed by coarse silt  
650 and fine sand, hosts the majority of the vertical to high angle fractures. These rocks display low to  
651 moderate TOC (i.e. between 1 and 2%).

652 - The depositional environment named offshore (and including facies associations such as  
653 massive siltstones, organic-rich siltstones, bioclastic siltstones, laminated siltstones), chiefly  
654 composed by fine silt (locally with clay), hosts the majority of horizontal fractures. No relationship  
655 between the abundance of the horizontal fractures and the host-rock TOC has been identified.

656 On the other hand, the petrographic analysis accomplished under CL, based on local  
657 observations (i.e. where thin sections were made; see list in Table S1a and Table S1B of the  
658 Supplementary Material) from the two investigated wells, suggests that the abundance of diagenetic  
659 carbonates (early or late diagenetic; Fig. 6 and 7) does not seem to be a main factor controlling the  
660 fracture occurrence, since most of the fracture host-rocks observed in thin section do not contain  
661 abundant diagenetic carbonates. An exceptions is represented by the offshore samples which seem  
662 to develop vertical fractures where abundant burial dolomite (D3) occurs.

663 A further aspect that can be deduced from the results of the present study on wells 16-17  
664 and 12-36 (Fig. 3) is that the vertical facies changes of the studied lithologies (stacking pattern) seem  
665 to also influence the horizontal fracture occurrence: the higher the vertical facies heterogeneity, the  
666 more numerous the horizontal fractures.

667 According to these general results issued from the two studied wells it could thus be  
668 concluded that fracture occurrence in the Montney siltstones seems to be chiefly controlled by host-  
669 rock facies and vertical facies changes (stacking pattern) and to a minor extent by the abundance of  
670 diagenetic dolomite. Conversely, host-rock TOC seems not to be correlated with the fracture  
671 abundance.

672 In order to further characterize the factors controlling fracture occurrence in the studied  
673 lithologies, these general results issued from the study of the two wells were integrated with  
674 available data from well 12-36 on bulk brittleness (computed from XRD and XRF continuous logs and  
675 calibrated with core measurements; Chatellier et al., 2018) and mineralogy (British Columbia Oil and  
676 Gas Commission public database). Mineralogy is however only available for two of the three  
677 investigated cores (i.e. core 1 and core 3; Fig. 14). In spite of the unfortunate lack of mineralogy  
678 analysis for core 2 that shows the highest bulk brittleness, the comparison between mineralogy and  
679 bulk brittleness of cores 1 and 3 (Fig. 14) suggests that clays/mica content is negatively correlated  
680 with brittleness and stiffness. This comparison further shows (Fig. 14) that the calcite/dolomite and

681 K-feldspar/plagioclase contents of core 1 is higher than in core 3, suggesting that these minerals may  
682 also control the bulk brittleness. Following Pei et al. (2014), it could be concluded that similarly to  
683 shales, the content in relatively hard minerals like feldspars and carbonates may enhance the  
684 brittleness of siltstones.

685 By combining the data on bulk brittleness for well 12-36 (Chatellier et al., 2018) with the  
686 occurrence/orientation of fractures (this study) it appears that the relationship between bulk  
687 brittleness and fracture occurrence is not straightforward (Fig. 14), so that fracture occurrence  
688 cannot be simply predicted from bulk brittleness derived from compositional logs only. Indeed,  
689 bedding-parallel horizontal fractures are observed in all of the three cores of well 12-36 despite the  
690 strong difference in bulk brittleness. This rather argues in favor of a control being also exerted by  
691 mechanical layering. Indeed, in contrast to core 1 which displays more homogeneous sedimentary  
692 facies and bulk brittleness, cores 2 and 3 include a number of depositional environments and facies  
693 and exhibit high frequency vertical changes in brittleness (Fig. 14). This results in horizontal fractures  
694 being less abundant in core 1 than in cores 2 and 3. and conclusively suggests that the occurrence of  
695 horizontal fractures is mainly related to 1) the occurrence of internal facies laminations and 2) the  
696 number of interfaces between layers of different brittleness (i.e. sedimentary facies transitions).  
697 Therefore, the Montney siltstones like shales seem to exhibit a rather high strength anisotropy  
698 (weakness parallel to bedding), though they are stiffer and more brittle than common shales.

699

700 Concerning the vertical fractures of well 12-36 (Fig. 14), they mostly occur in core 3 which is  
701 composed by facies with the lowest bulk brittleness (likely due to the presence of high clays/mica  
702 contents; Fig. 14). This is again hard to explain from the bulk brittleness log only. The results of the  
703 present study suggested however that the facies coarseness and/or the presence of D3 ferroan  
704 dolomite had a potential role in concentrating these fractures.



705 Summarizing, our results from wells 16-17 and 12-36 together with the literature data from  
706 well 12-36 only (Chatellier et al., 2018, and British Columbia Oil and Gas Commission public database)  
707 show that the Montney Fm siltstones have a peculiar mechanical behavior. Indeed, the fracture  
708 occurrence seems not to be controlled by the host-rock TOC (Fig. 5 and Fig. 9). Furthermore, the lack  
709 of direct correlation between the occurrence of fractures and bulk brittleness (calculated according  
710 to Chatellier et al., 2018; Fig. 14) suggests that bulk brittleness cannot be used alone as a reliable  
711 proxy for predicting the occurrence of natural fractures in such lithologies. Here other factors may  
712 play a role, such as the vertical facies heterogeneity and/or the facies coarseness, locally associated  
713 with the presence of specific carbonate diagenetic phases.

714

## 715 ***5.2 Fracture timing and foreland evolution***

716 Timing of fracture opening is difficult to constrain, and without this information, ascribing  
717 them to a given specific mechanism, is problematic. In the absence of absolute radiometric dating of  
718 calcite vein cements (Beaudoin et al, 2018; Hansman et al., 2018; Mangenot et al., 2018; Parrish et  
719 al., 2018), relative constraints on fracture timing may come from field/core evidence of  
720 abutting/crosscutting relationships and petrographic evidence, combined with isotope geochemistry  
721 and thermometry evidence (Gabellone et al., 2013; Quesnel et al., 2016; Hooker et al., 2017b, 2019).  
722 In particular, the chemical and thermal signatures of the syn-kinematic vein cements reflect fluid  
723 conditions during fracturing, and hence may help constraining the timing of fracture opening if  
724 combined with thermal basin modeling accounting for the regional tectonic evolution (e.g. Roure et  
725 al., 2010; Gasparrini et al., 2014).

726 In this study, both vertical and horizontal mineralized fractures were investigated from wells  
727 16-17 and 12-36 of the Montney-Doig Fms (Fig. 8 and 10). The strike for the vertical fractures is  
728 unknown, which hampers the definition of fracture sets based on their orientation. Therefore, the  
729 term “generation” is here used to distinguish calcite veins formed at different times, and possibly by

730 different mechanisms, chiefly based on petrography and FI evidence. Having said this, three  
731 generations of veins were identified in the studied wells cores: the first and third oriented vertical to  
732 high angle and the second oriented horizontal. Syn-kinematic features like fibrous to elongated  
733 blocky calcite crystal habitus arranged in syntaxial to antitaxial textures and/or crack-seal mechanism  
734 evidences are common in the three vein generations (see figure 10 and Table S1a and Table S1b in  
735 the Supplementary Material), allowing to discuss the timing of vein opening from assessing calcite  
736 precipitation time.

737         The three generations of veins are hereafter integrated to the thermal/burial history of the  
738 Montney-Doig Fms and interpreted within the geodynamic evolution of the WCSB. (Fig. 15). With  
739 such aims, the temperature/depth evolution with time of the Montney-Doig Fms, modeled by the  
740 TemisFlow<sup>TM</sup> software, was used. A 1D thermal curve was extracted for the studied location from the  
741 available 3D model, by assuming a thickness of 3300m of eroded sediments during the Tertiary uplift,  
742 according to previous structural restorations (Faure et al., 2004; Hardebol et al., 2009; Roure et al.,  
743 2010). The curve highlights four main stages in the geological evolution of the WCSB since the  
744 Triassic (Fig. 15): 1) a first stage governed by a retro-foreland regime in the Triassic; 2) a transition  
745 toward the Jurassic collisional foreland regime; 3) two phases of fast subsidence in the Middle  
746 Cretaceous and Latest Cretaceous – Early Paleocene, respectively, separated by a subsidence plateau  
747 in the Late Cretaceous; 4) a phase of exhumation and associated sediment erosion or bypass starting  
748 from the Early Paleogene, in association with the Laramide Orogeny. The two Cretaceous phases of  
749 subsidence and the Early Paleogene exhumation are well time-correlated with the tectonic pulses  
750 identified by Paná and van der Pluijm (2015). Notably, the temperatures modeled for the maximum  
751 rock burial do not exceed 150°C (given uncertainties), in agreement with the observation that the  
752 calcite from all veins displays thin and rectilinear twins (Fig. 10), suggesting that twinning occurred at  
753 temperatures below 170-200°C (Ferrill et al., 2004; Lacombe, 2010).

754         The first generation of vertical veins is cemented by calcite precipitated at about 110°C from  
755 basinal brines (20 eq. NaCl wt%) and which also carries oil inclusions (Table 3 and Fig. 13), recording

756 that the oil generation had already started. Triassic source rocks from the WCSB started to generate  
757 hydrocarbons during the Late Cretaceous (88,5 Ma) based on Ducros et al. (2017), and even earlier  
758 based on Ness (2001). Accordingly, the age of the first vein generation is bracketed between 100 Ma  
759 and 70 Ma, i.e. atimespan post-dating the beginning of oil generation and consistent with a 110°C  
760 burial temperature for the studied succession (Fig. 15). These vertical veins therefore possibly  
761 opened during the Late Cretaceous when vertical movements of the foreland were still limited and  
762 fast sedimentation of the Colorado Group had occurred (Pană and Van der Pluijm, 2015; Ducros et  
763 al., 2017, Rohais et al., 2018). Since the strike of these vertical veins is unknown, it is not possible to  
764 determine under which stress regime they developed. They could have formed either under a  
765 burial/flexure related vertical maximum principal stress  $\sigma_1$  or under a horizontal maximum principal  
766 stress  $\sigma_1$  oriented NE-SW that was likely prevailing during the Cretaceous in the Rocky Mountains  
767 (Vandeginste et al., 2012) if tectonic stress magnitude had already overcome the burial-related stress  
768 magnitude in the basin at that time (e.g. layer-parallel shortening, Tavani et al., 2015).

769 The horizontal, layer-parallel, veins (second generation) are cemented by calcite which hosts  
770 mono-phase liquid  $\text{CH}_4 \pm \text{CO}_2$  inclusions, together with oil inclusions, indicating that they formed after  
771 the onset of gas generation. Though the trapping conditions for these hydrocarbon inclusions could  
772 not be estimated, the presence of  $\text{CH}_4$  strongly suggests that the calcite cements precipitated at  
773 temperatures higher than 110°C (i.e. higher than the temperature inferred for the vertical veins of  
774 the first generation). Triassic source rocks from WCSB reached the maximum transformation rate just  
775 before maximum burial (~58-57 Ma) which also corresponds to a maximum overpressure in these  
776 rocks, due to peak hydrocarbon generation and secondary cracking of oil into gas (Ducros et al.,  
777 2017). Therefore, the horizontal veins are interpreted to have formed during Late Cretaceous to Early  
778 Paleogene, just before the host lithologies reached the thermal/burial maximum (Fig. 15).

779 Despite the multiple sources of uncertainties, the age for thermal/burial maximum (~58-57  
780 Ma) is rather close to the age of the last tectonic pulse (~54-52 Ma) identified by Pană and van der  
781 Pluijm (2015). The formation of such horizontal, layer-parallel veins at nearly the maximum burial

782 depth was likely associated with pore-fluid pressures exceeding the vertical overburden load (e.g.  
783 Price, 1975; Henderson et al., 1990). The mechanisms for opening horizontal fractures has been  
784 recently summarized by Hooker et al. (2019). In case of prevailing strike-slip faulting stress regime,  
785 the horizontal fracture opening would have required a poroelastic effect in response to fluid pressure  
786 increase in order to counteract the overburden and the lateral stress in proportion (Engelder and  
787 Fischer, 1994), flipping the vertical stress to be the least compressive stress by the time the tensile  
788 failure criterion is reached. However, the most likely explanation is that the bedding-parallel veins  
789 were related to catagenesis and formed from fluid overpressure within a thrust-faulting regime (in  
790 which the overburden is the least compressive stress), in line with the Eocene pulse of regional  
791 contraction (Pană and van der Pluijm, 2015) and the compressional or transpressional stress regime  
792 reconstructed by Vandeginste et al. (2012) in the Rocky Mountains to the South East. The  
793 overpressure increase, here likely induced by CH<sub>4</sub> generation, is in agreement with earlier works  
794 emphasizing the role of hydrocarbon generation as potential drivers of fracture growth in organic-  
795 rich mudrocks (Spencer, 1987; Ozkaya, 1988; Vernik, 1994; Jochum et al., 1995; Marquez and  
796 Montjoy, 1996; Zanella et al., 2014; Meng et al., 2017). Although the horizontal fractures are  
797 interpreted to be chiefly governed by gas generation, they are not more abundant in high TOC levels  
798 (Fig. 9). This apparent paradox may be explained by the entire succession being overpressured  
799 throughout together with the high frequency vertical facies changes of the studied rocks (already  
800 underlined in section 5.1). Indeed, these rocks consist of dm to m scale alternations of OM-rich and  
801 more arkosic/carbonatic siltstones; it seems likely that peak gas generation provided distributed fluid  
802 overpressures that concentrated fracturing in the more brittle beds located above or below the OM-  
803 rich ones.

804 Finally, the third generation of vertical veins shows petrographic evidence for post-dating the  
805 bedding-parallel ones of second generation (Fig. 10G and H) and also contains monophasic liquid  
806 CH<sub>4</sub>±CO<sub>2</sub> inclusions. We propose that these vertical veins opened during basin uplift (Middle-Late  
807 Paleogene) that occurred after peak burial. Vein opening is here interpreted as resulting from the

808 progressive decrease of the far-field horizontal stress magnitude and the transition from Laramide  
809 compression to regional transtensional faulting regime (Middle Paleogene). This new tectonic stress  
810 likely combined to residual fluid overpressure to drive the opening of the vertical fractures at depth.  
811 Indeed, the proposed fracture timing is in line with the study of Chatellier et al. (2018) who identified  
812 overpressure associated with tectonic uplift by comparing the reservoir pressures against normal  
813 hydrostatic pressure trends and also with the model of Ducros et al. (2017) indicating that a fluid  
814 overpressure regime persisted after thermal/burial maximum in areas where HC occurred.

815 In conclusion, the opening of the horizontal veins related to gas generation certainly required  
816 fluid overpressure to overcome the weight of the overburden. The occurrence of relatively abundant  
817 vertical veins at depth may indicate that their formation was also probably assisted by elevated fluid  
818 pressures. In this study however, the quantification of fluid pressure remained out of reach. Only the  
819 fine investigation of immiscible co-genetic fluid inclusions (e.g. Roure et al., 2010) or the joint  
820 application of thermometric techniques (e.g. Mangelot et al., 2017; Honlet et al., 2018) may  
821 potentially give access to fluid pressure conditions during calcite crystal growth and confirm the  
822 proposed hypotheses.

823

### 824 **5.3 Calcite parent fluids and openness of the system**

825 Paleo-water (or paleo-fluid) circulation in sedimentary basins plays a key role in governing  
826 burial diagenetic processes (Bjørlykke, 1994; Kyser et al., 2005; Roure et al., 2005). Previous studies  
827 have stressed the importance of establishing the origin of burial diagenetic fluids in siliciclastic  
828 successions (Morad et al., 2002; Bjørlykke and Jahren 2012; Deschamps et al., 2012). Indeed, aiming  
829 to predict the spatial and temporal distribution of diagenetic alterations at basin scale and the  
830 present day heterogeneities in the reservoirs, it is of prime importance to define whether reactions  
831 were accomplished by fluid advection (active flow) or diffusion (no flow), or in other words whether  
832 the fluids were locally sourced (closed system) or derived from external sources (open system). In

833 this respect, the openness of fluid systems driving diagenetic reactions (including carbonate  
834 cementations) may be assessed by investigating the occurrence of mineral assemblages and their  
835 isotope and elemental geochemistry (Day-Stirrat et al., 2010; Fantle et al., 2010; Khalifa and  
836 Gasparri, 2014; Swart, 2015). Organic-rich mudrocks represent a peculiar case of siliciclastic rocks,  
837 since together with the paleo-waters, they may host locally sourced hydrocarbon liquids. Here, the  
838 investigation of both diagenetic minerals and organic components may help to characterize the  
839 openness of the paleo-water and petroleum systems.

840         The Montney-Doig Fms have been long investigated at the scale of the WCSB to determine  
841 the origin of their hydrocarbons to finally discriminate if they mostly consist in tight reservoirs  
842 (where hydrocarbons migrated from an external source) or if they locally also behaved as source-  
843 rocks (where hydrocarbons are internally sourced). It is commonly accepted that the Doig Fm  
844 contains a prolific source rock interval (the Doig Phosphate Zone) and produced liquid hydrocarbons  
845 (Creaney and Allan, 1990; Riediger et al., 1990; Allan and Creaney, 1991; Ness, 2001, Ejezie, 2007;  
846 Ducros et al., 2017). Conversely, despite the Montney Fm contains huge unconventional oil and gas  
847 resources (Neb, 2013), there is less consensus on the potential contribution of the Montney  
848 lithologies to the overall produced hydrocarbons. Indeed, the studies on organic components  
849 conducted so far do not find a common agreement and both hypotheses (tight reservoirs *versus*  
850 source rocks) are presently supported.

851         On one side, it is believed that the organic matter from the Montney rocks is composed  
852 almost entirely of solid bitumen or pyrobitumen (Chalmers and Bustin, 2012; Sanei et al., 2015;  
853 Wood et al., 2015, 2018, 2020; Ardakani et al., 2020). In particular, a modified Rock-Eval procedure  
854 and organic petrography were applied by Sanei et al. (2015) on Montney Fm samples from one well  
855 core located ca. 50 Km South of Fort St. John, within the inherited structure from the PRA collapse.  
856 Based on the absence of primary deposited kerogen, these authors concluded that the majority of  
857 the TOC in the studied samples consists of solid bitumen (or pyrobitumen). The latter resulted from

858 the secondary cracking of a former liquid oil phase which migrated into the larger paleo-intergranular  
859 pore spaces of the organically lean rocks of the Montney Fm. Sanei et al. (2015) also believe in  
860 previous oil-source rock correlation studies (Riediger et al., 1990a, 1990b) based on biomarker  
861 signatures and suggesting that all oils produced from Triassic rocks in the WCSB have been sourced  
862 either from the Doig Fm or the Nordegg Member of the Fernie Fm.

863 On the other hand, a study conducted by Romero-Sarmiento et al. (2016b) on core and  
864 cutting samples (partly falling close to wells 16-17 and 12-36) led to the conclusion that the Montney  
865 Fm may contain potential source rock beds. Based on artificial thermal maturation, organic  
866 petrography and gas chromatography, these authors indicate that the samples represent a series  
867 from immature and early mature source rocks with preserved liptinite, to post-peak-mature and  
868 over-mature source and tight reservoir rocks containing abundant solid bitumen.

869 Montney lithologies from cores and cuttings distributed in a large area and encompassing  
870 well 16-17, were analyzed by Crombez et al. (2016) for organic content and for major and trace  
871 element concentrations. Their results illustrate that sequence 3 presents propitious redox conditions  
872 and primary productivity for source rocks development and emphasize the occurrence of primary  
873 deposited kerogen in this stratigraphic interval.

874 The presence of source-rock intervals in the Montney Fm is also supported by Feng et al.  
875 (2016, 2017), who identified three different oil families with distinct geochemical signatures within  
876 hydrocarbon accumulations of the Montney Fm. One of the oil family appears to be originated from  
877 the Montney Fm since it bears striking resemblance to the Montney rock extracts.

878 Finally, very recent studies also highlighted the presence of primary organic facies, coexisting  
879 with disseminated and pore-filling solid bitumen in the Montney Fm from Alberta (Ardakani et al.,  
880 2020; Becerra et al., 2020). These observations further call into question the possible local source the  
881 hydrocarbons within this formation.

882 The present survey, conducted on mineralized fractures from two wells located ca. 70 km  
883 North-West of Fort St. John (i.e. in between the HRSZ and the inherited structure from the PRA  
884 collapse; Fig. 1A), may bring some light from independent observations to this controversy. In the  
885 samples here investigated, three different generations of veins were identified (Fig. 10), carrying  
886 different types of HC fluids (Table 3) and having different timing within the WCSB geological and  
887 geodynamic evolution (Fig. 15). Many of the studied veins are filled by calcite crystals with fibrous to  
888 blocky fibrous habitus in syntaxial to antitaxial arrangements and may display features of crack-seal  
889 mechanism (Fig. 10 and Tables S1a and S1b from Supplementary Material). This suggests that the  
890 mineral precipitation commonly occurred during vein opening (syn-kinematic calcite) so that the  
891 fractures were never particularly porous throughout growth. The UV-light and CL response (chiefly  
892 governed by Fe and Mn contents), together with the isotope composition, are very similar for all the  
893 studied calcite cements, irrespective of the well core of belonging, the vein generation and the  
894 crystal habitus/textures (Fig. 10 and 11). In conclusion, despite the different timing of the vein  
895 calcites (Late Cretaceous to Paleogene), precipitating at different maturity stages of the HC system,  
896 the calcite cements display similar petrographic and geochemical features (Fig. 10 and 11). This  
897 conclusively points at fluids in continuous equilibrium with the host-rocks (Hooker et al., 2017b,  
898 2019). According to this scenario, the calcite veins identified probably formed as hydraulic fractures  
899 from a closed fluid system which would have persisted during the burial history of the studied  
900 lithologies.

901 Another argument in favor of this closed-system hypothesis is the lack of diagenetic sulfate  
902 minerals within the studied samples (Fig. 7). This contrasts with the occurrence of anhydrite and  
903 barite in samples from well cores located South of Fort St. John (within the inherited structure from  
904 the PRA collapse; Liseroudi et al., 2020), and which is interpreted as due to the contribution of  
905 externally sourced paleo-waters (i.e. sulfate-bearing fluids originated from the dissolution of  
906 underlying Devonian evaporites and migrated upwards through deep-seated faults/fractures to the  
907 Montney Fm).



908           The present survey does not question the occurrence of pyrobitumen or of primary kerogen  
909 in the Montney-Doig system throughout the WCSB and deciphering between migrated and in place  
910 hydrocarbons in these lithologies at basin scale is far beyond our scopes. Nevertheless, this study  
911 clearly suggests that, at least in the limited portion of the basin investigated, the advection of  
912 externally sourced paleo-waters (*sensu* McCaig and Knipe, 1990) was not recorded during the  
913 precipitation of calcites within the three fracture generations identified (Fig. 15). The lack of evidence  
914 for externally sourced paleo-waters is rather in agreement with a local closed-system behavior for  
915 the studied rocks, which would thus be in line with the hypothesis of a system that may have self-  
916 sourced and retained hydrocarbons. As the studied wells are located in between the HRSZ and the  
917 inherited structure from the PRA collapse (Fig. 1A), we may speculate that this area was primarily  
918 characterized by a closed-system hydrodynamic regime, at least during the Late Cretaceous to  
919 Paleogene evolution of the basin. This may have important implications for future exploration.

920

## 921       **6. CONCLUSIONS**

922           A multidisciplinary approach (including sedimentology, Rock-Eval pyrolysis, petrography, O-C-  
923 Sr isotope geochemistry and fluid inclusion microthermometry) has been applied for the first time to  
924 natural mineralized fractures (veins) hosted by mudrocks of the Lower-Middle Triassic Montney-Doig  
925 unconventional resource play from the Western Canada Sedimentary Basin.

926           Montney-Doig well core samples (ca. 2100-2500 m in depth) were collected from two wells in  
927 British Columbia. These rocks were deposited in shoreface to offshore environments and display  
928 variable mineralogy and organic content (TOC of 0.6 to 6.6 wt%).

929           Main analytical results of this survey allow for drawing the following three-fold conclusions:

- 930       1) Fracture occurrence was chiefly controlled by host-rock facies and only to a minor extent by  
931       host-rock diagenesis. Horizontal (bed-parallel) fractures are preferentially localized in very

932 fine facies (siltstones and siltstones with clays) of the offshore environment and where high  
933 frequency facies changes induced vertical heterogeneities. Vertical to high angle fractures  
934 are more abundant in the coarser facies of the offshore transition environment (coarse  
935 siltstones and fine sandstones), though they may also occur in offshore facies (e.g.  
936 calcispheric dolosiltstones) which have undergone burial dolomitization. Interestingly, host-  
937 rock TOC seems not to have controlled fracture occurrence.

938 2) Three generations of calcite cemented fractures were identified:

939 - The first generation (vertical veins), cemented by calcite precipitated at  $\sim 110^{\circ}\text{C}$  from brines  
940 carrying oil inclusions, records that oil generation had begun; they possibly opened when  
941 vertical movements of the foreland were limited and rapid sedimentation of the Colorado  
942 Group occurred (Late Cretaceous).

943 - The second generation (horizontal veins), cemented by calcite carrying  $\text{CH}_4 \pm \text{CO}_2$  inclusions,  
944 originated during peak HC generation; they possibly opened just before maximum burial  
945 (Late Cretaceous - Early Paleogene) as a result of overpressures induced by  $\text{CH}_4$  generation,  
946 assisted by compressive horizontal stress. Fairly distributed fluid overpressures throughout  
947 the succession would explain why they principally developed in the more brittle lithologies,  
948 rather than in the OM-rich intervals.

949 - The third generation (vertical veins), cemented by calcite also containing  $\text{CH}_4 \pm \text{CO}_2$  inclusions,  
950 possibly opened during basin uplift (Middle-Late Paleogene) due to horizontal stress  
951 decrease, assisted by residual fluid (gas) overpressure.

952 3) UV-light and CL response of the calcite cements together with their isotope geochemistry  
953 signature are quite consistent, irrespective of the vein generation and well core of belonging,  
954 pointing at lack of evidences for externally sourced paleo-waters. This suggests a closed  
955 system behavior for the studied rocks during Late Cretaceous – Paleogene times, at least in  
956 the limited portion of the basin investigated.

957 Overall the study underlines the relevance and usefulness of the applied methodological  
958 approach to characterize fracture controls, paleo-fluid circulation and openness of the fluid system in  
959 naturally fractured unconventional plays, with implications for exploration.

960

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973

## 974 **FIGURE CAPTIONS**

975

976 **Fig. 1.** Geographic and geological setting of the study area. **A.** Geological map of the Canadian  
977 segment of the Cordillera including the main belts and terranes, and the Western Canada  
978 Sedimentary Basin (WCSB). The purple color shows the subcrop and outcrop areas where Triassic  
979 deposits are still preserved (simplified from Rohais et al., 2018 and references herein). The Hay River  
980 Shear Zone (HRSZ) and the inherited structure from the Paleozoic Peace River Arch (PRA) collapse are

981 highlighted as main structural elements below the Triassic strata. (1) and (2) refer to the wells  
982 investigated, localized in between these two main structural domains. The black dotted line  
983 represents the trace of a cross-section (illustrated in B). **B.** Cross-section of the WCSB illustrating the  
984 main stratigraphic intervals and the faults (simplified from Ducros et al., 2017).

985

986 **Fig. 2.** Simplified sedimentary architecture and main stratigraphic sequences of the Lower and  
987 Middle Triassic of the WCSB (modified after Crombez et al., 2019).

988

989 **Fig. 3.** Simplified stratigraphy of the two investigated wells including the gamma ray response and  
990 sequences, the TOC and MinC from Rock-Eval (by merging data from this study and from Crombez et  
991 al., 2016), the facies associations (in yellow) and their interpretation in terms of depositional  
992 environments (in green), as well as the location of the observed fractures with their orientation. See  
993 Table 1 for the legend of numerical codes used to define the different facies associations and  
994 depositional environments.

995

996 **Fig. 4.** Reconstructed depositional profile for the Montney-Doig interval with main facies associations  
997 and microfacies petrographic images (in PPL) for some of them. 1. Lag; 2. Massive fine-grained  
998 sandstones; 3. Massive very fine-grained sandstones to siltstones; 4. Sand-dominated alternation of  
999 very fine-grained sandstones to siltstones; 5. Silt-dominated alternation of very fine-grained  
1000 sandstones to siltstones; 6. Massive siltstones; 7. Organic-rich massive siltstones; 8. Bioclastic  
1001 siltstones; 9. Laminated siltstones with mm-thick sand laminae (low density turbidite); 10. Laminated  
1002 siltstones.

1003

1004 **Fig. 5.** Frequency distribution of TOC values (in wt. %) from the two investigated wells according to  
1005 two different datasets: 1) continuous sampling of cores and cuttings from Crombez et al. (2016), in  
1006 black; 2) core samples hosting fractures from this study, in grey.

1007

1008 **Fig. 6.** Petrographic photos illustrating the main carbonate diagenetic phases occurring in the studied  
1009 samples of the Montney-Doig Fms. **A.** Euhedral to sub-hedral D1 crystals with dull orange CL totally  
1010 or partially replacing detrital dolomite grains (sample 16-17-21). **B.** D2 crystals with bright red CL  
1011 locally forming overgrowths around detrital grains replaced by D1 (sample 16-17-23). **C.** Facies with  
1012 calcispheres from the offshore environment (sample 16-17-21). PPL. **D.** Same image of C. under CL.  
1013 Calcispheres are replaced by the ferroan and non-luminescent D3. **E.** C2 crystals with bright orange  
1014 CL occurs as cement in intergranular pores or as replacement of calcispheres (sample 1236-C3-6). **F.**  
1015 C2 with bright orange CL occurs as cement in intergranular pores or replaces previous carbonate  
1016 phases (sample 1617-15). PPL= plane polarized light, CL= cathodo luminescence.

1017

1018 **Fig. 7.** Paragenetic sequence reconstructed for the Montney-Doig Fms in this study as compared with  
1019 the one from Vaisblat et al. (2017b). In black the phases that were observed in both studies; in blue  
1020 the phases that were observed only by Vaisblat et al. (2017b); in red the phases that were observed  
1021 only in this study. Black arrows indicate the four most abundant carbonate diagenetic phases.

1022

1023 **Fig. 8.** Well core pieces containing macroscopic mineralized fractures (i.e. veins). **A.** Horizontal vein  
1024 from well 12-36 at 2364.3 m of depth. **B.** Vertical vein from well 16-17 at 2350 m of depth.

1025

1026 **Fig. 9.** Frequency distribution of the properties FRACTURE, SEQUENCE, ENVIRONMENT and TOC from  
1027 16-17 and 12-36 well cores based on the sampled points reported in the spreadsheets of Tables S3a  
1028 and S3b (see Supplementary Material). The red line represents the distribution of the whole sampled  
1029 points where fractures occur (i.e. only the points with codes 2 and 3 of the property FRACTURE). The  
1030 black bars in the upper (A) and lower (B) histograms represent the sampled points corresponding  
1031 respectively to the occurrence of the horizontal and vertical to high angle fractures. The “777” values  
1032 are attributed to cells of the spreadsheets where information on a given property was not available.  
1033 In the histograms to the right (property TOC) the “777” sampled points represent ~80% of the total.  
1034 See Table 1 for code explanation.

1035

1036 **Fig. 10.** Photomicrographs of cemented fractures (i.e. veins) from the 16-17 and 12-36 wells. All  
1037 images are oriented with the upper part of the photos corresponding to the stratigraphic up  
1038 direction. PPL= plane polarized light, CPL= crossed-polarized light, CL= cathodoluminescence. **A.**  
1039 Vertical vein filled by blocky calcite crystals (sample 16-17-23). PPL. **B.** Same image of A. under CL.  
1040 The blocky calcite shows uniform, bright orange luminescence. **C.** Horizontal vein filled by fibrous to  
1041 elongated blocky calcite crystals with antitaxial arrangement. Median planes are highlighted by  
1042 yellow dotted lines. Crystal deformation suggests mixed mode I-mode II vein opening (sample 12-36-  
1043 C3-2). CPL. **D.** Same image of C. under CL. The crystals have dominantly bright orange luminescence.  
1044 Some of them may evolve from bright to dull orange during growth. **E.** Horizontal vein filled by  
1045 elongated blocky to blocky calcite crystals displaying incremental growth by crack-seal episodes,  
1046 highlighted (yellow arrows) by strings of wall rock inclusions, demonstrating mode I vein opening  
1047 (sample 16-17-2). PPL. **F.** Same image of E. under CL. The crystals show uniform bright orange  
1048 luminescence. **G.** Vertical vein filled by blocky calcite crystals replacing blocky (locally elongated  
1049 blocky) calcite crystals filling an horizontal vein as highlighted by the yellow dotted line (sample 12-

1050 36-C2-2B). CPL. **H.** Same image of G. under CL. The calcite crystals filling the different veins show  
1051 undistinguishable uniform bright orange luminescence.

1052

1053 **Fig. 11.** O-C stable isotope composition of the fracture sealing calcite cements. The vein orientation  
1054 (horizontal or vertical to high angle) and the stratigraphic sequence of belonging are reported. The  
1055 O-C isotope composition of calcites precipitated in equilibrium from Early Triassic seawater (from  
1056 Veizer et al., 1999) is represented by the light blue squares. **A.** Calcite veins from well 16-17. **B.**  
1057 Calcite veins from well 12-36.

1058

1059 **Fig. 12.** Photomicrographs illustrating some petrographic features of the analyzed FIs at room  
1060 temperature. **A.** Assemblages of possibly primary Type I FIs (yellow arrows) occurring in the core of  
1061 calcite crystals (vertical vein, sample 16-17-23). **B.** Details of assemblage of Type I FIs from the same  
1062 sample illustrated in A. **C.** Assemblage of possibly primary Type II FIs occurring in the calcite crystal  
1063 cores (vertical vein, sample 16-17-23). **D.** Same image of C. under UV-light revealing a bright green  
1064 fluorescence for the Type II FIs. **E.** Assemblage of possibly primary Type IV FIs displaying shapes from  
1065 negative crystal to crystallographically controlled (horizontal vein, sample 12-36-C2-2B). **F.** Same  
1066 image of E. under UV-light revealing a weak green fluorescence for some of the Type IV FIs.

1067

1068 **Fig. 13.** Results of microthermometry for primary FIs in calcite cements from differently oriented  
1069 veins. **A.** Frequency distribution of Th values for Type I and Type II FIs (sample 16-17-23, vertical  
1070 vein). The red bar refers to possibly reequilibrated Type I FIs. **B.** Frequency distribution of Tmi values  
1071 for Type I FIs (sample 16-17-23, vertical vein). **C.** Frequency distribution of Th values for Type IV FIs  
1072 (sample 12-36-C2-2B, horizontal and vertical veins). **D.** Frequency distribution of Th values for Type II

1073 and Type IV FIs (sample 12-36-C1-4, horizontal vein). Light and dark grey bars refer to horizontal and  
1074 vertical veins respectively.

1075

1076 **Fig. 14.** Results of this study (facies associations, depositional environments, TOC, location and  
1077 orientation of fractures) for the three cores (core 1, core 2, core 3) of well 12-36 are reported  
1078 together with data from literature: mineralogy (from British Columbia Oil & Gas Commission) and  
1079 brittleness (from Chatellier et al., 2018). High values of brittleness are in red, low values in blue,  
1080 intermediate values are in green. Mineralogy data are available for core 1 and core 3 only. Red  
1081 horizontal lines highlight the horizontal fractures. See figure 3 for location of the cores along well 12-  
1082 36.

1083

1084 **Fig. 15.** Thermal evolution of the Montney-Doig Fms, from deposition to present day, modeled with  
1085 TemisFlow™ by considering 3300 m of eroded sediments during Tertiary uplift (Ducros et al., 2017).  
1086 The main stages of the basin evolution are reported together with the possible time-temperature  
1087 frame for the three generations of veins: first vertical generation (blue arrow), second horizontal  
1088 generation (red arrow), third vertical generation (yellow arrow).

1089

1090



1091 **TABLE CAPTIONS**

1092 **Table 1.** Tables summarizing the codes attributed to the main properties (i.e. sequence, facies  
1093 associations, depositional environments, fracture occurrence) recorded along the logs of the studied  
1094 well cores. These codes were used for frequency distribution analysis with EasyTrace™ software.

1095

1096 **Table 2.** Range and mean TOC and HI values from bulk-rock Rock-Eval measurements for the  
1097 investigated samples by sequence and well core of belonging.

1098

1099 **Table 3.** Results of FI study from wells 12-36 and 16-17. Sample name, fracture orientation, texture  
1100 of the sealing calcite phases, type and petrographic occurrence of fluid inclusions (FI), type of fluid  
1101 phases and their response under UV-light are reported, together with the range and mode values of  
1102 homogenization temperatures ( $T_h$ ), ice melting temperatures ( $T_{mi}$ ) and calculated salinity.

1103

1104 **SUPPLEMENTARY MATERIAL CAPTIONS**

1105 **Table S1a.** Details on investigated samples from well core 16-17, including sample name, depth,  
1106 sequence of provenance, depositional environment, number of fractures, their orientation and  
1107 kinematic aperture, texture of the sealing calcites and their O-C-Sr isotope composition. In some  
1108 samples, O-C isotope analysis were performed for different veins or for different calcite textures  
1109 occurring in the same vein.

1110

1111 **Table S1b.** Same as in Table S1a for the samples from well core 12-36.

1112

1113 **Table S2a.** Details on samples from well core 16-17, including sample name, depth, sequence of  
1114 provenance, depositional environment and the parameters obtained from Rock-Eval analysis  
1115 accomplished with the Shale Play Method<sup>®</sup> developed at IFP Energies nouvelles (Pillot et al., 2014;  
1116 Romero-Sarmiento et al., 2016a, 2016b): Sh0, Sh1, Sh2, Tmax, TOC, HI, OI, HCcont, MinC. TOC= total  
1117 organic carbon, HI= hydrogen index, OI= oxygen index, HCcont= content in hydrocarbons, MinC=  
1118 mineral carbon. Column names with –B and –E refer to analyses performed on bulk-rock and  
1119 extracted organic matter, respectively.

1120

1121 **Table S2b.** Same as in Table S2a for the samples from well core 12-36.

1122

1123 **Table S3a.** Spreadsheet for well core 16-17 with the evolution of different properties with depth  
1124 used for frequency distribution analysis with Easytrace<sup>™</sup> software. The properties considered are:  
1125 presence/absence of fractures (i.e. FRACTURE), the stratigraphic sequence (i.e. SEQUENCE), the  
1126 depositional environment (i.e. ENVIRONMENT), the TOC values (i.e. TOC). Each line of the table

1127 represents a point virtually sampled along the log of well 16-17 to which information on the different  
1128 properties were attributed via numerical codes (see Table 1). As requested by the software, the value  
1129 “777” was attributed to the cells where information for a given property was not available.

1130 **Table S3b.** Same as in Table S3a for well core 12-36.

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| <b>FACIES ASSOCIATION</b>            | <b>Code</b> | <b>SEQUENCE</b>     | <b>Code</b> | <b>TOC</b>             | <b>Code</b> |
|--------------------------------------|-------------|---------------------|-------------|------------------------|-------------|
| Lag                                  | 1           | Sequence 1          | 1           | <1                     | 1           |
| Massive fine-grained sandstones      | 2           | Sequence 2          | 2           | between 1 and 2        | 2           |
| Massive vf-grained ss. to siltstones | 3           | Sequence 3          | 3           | between 2 and 3        | 3           |
| Sand-dominated alternation           | 4           | Sequence 4          | 4           | between 3 and 4        | 4           |
| Silt-dominated alternation           | 5           |                     |             | between 4 and 5        | 5           |
| Massive siltstones                   | 6           | <b>ENVIRONMENT</b>  | <b>Code</b> | between 5 and 11       | 6           |
| Organic-rich siltstones              | 7           |                     |             |                        |             |
| Bioclastic siltstones                | 8           | Lag                 | 1           | <b>FRACTURE</b>        | <b>Code</b> |
| Laminated siltstones (turbidite)     | 9           | Shoreface           | 2           |                        |             |
| Laminated siltstones                 | 10          | Offshore transition | 3           | no fracture            | 1           |
|                                      |             | Offshore            | 4           | horizontal             | 2           |
|                                      |             |                     |             | vertical to high angle | 3           |

**Table 1**

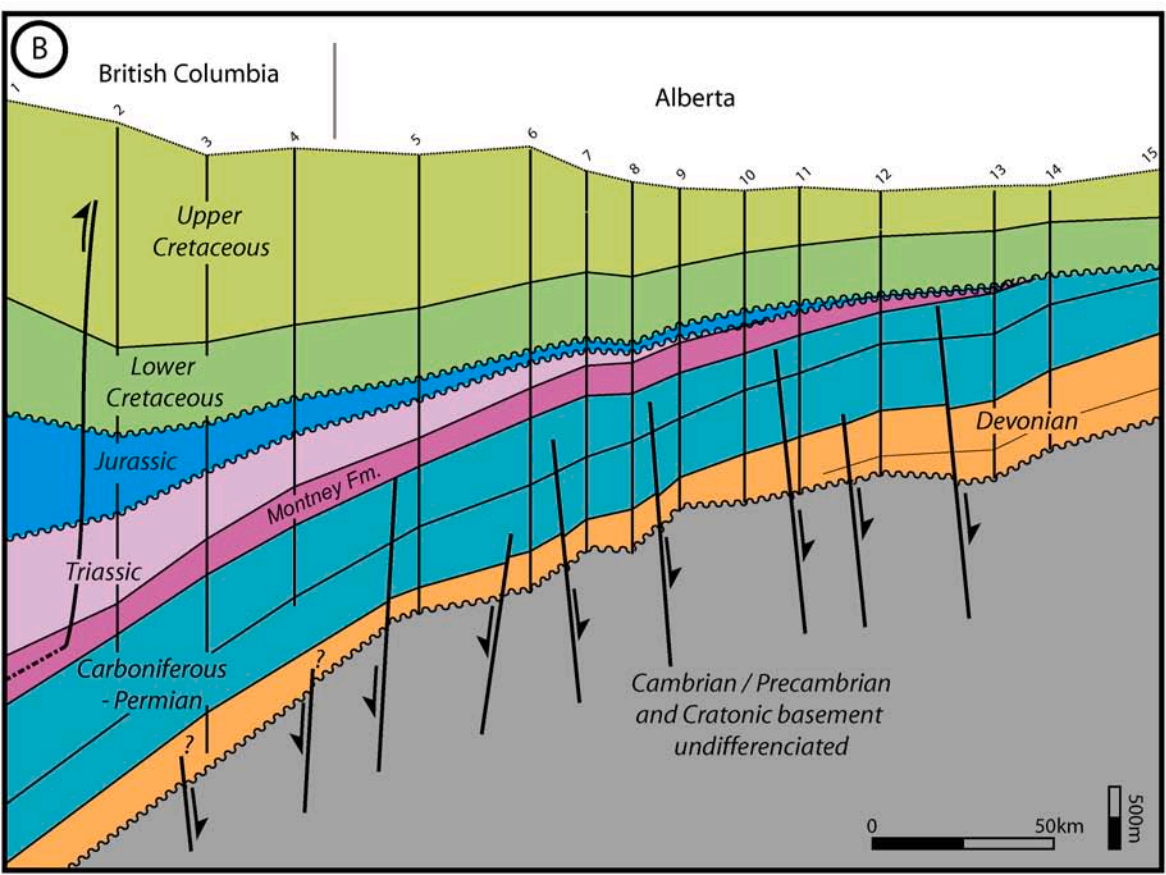
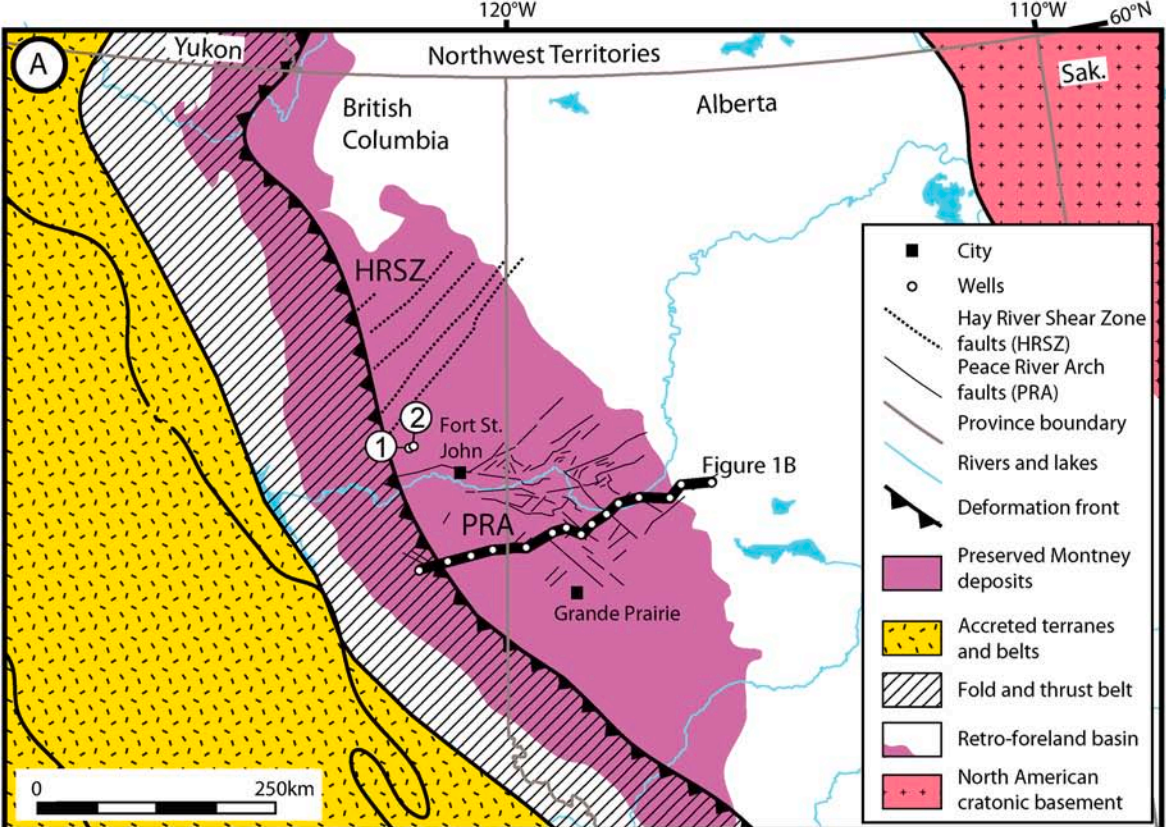
| Stratigraphic<br>sequence | Core 16-17 |      |                |      | Core 12-36 |      |                |      |
|---------------------------|------------|------|----------------|------|------------|------|----------------|------|
|                           | TOC (%)    |      | HI (mgHI/gTOC) |      | TOC (%)    |      | HI (mgHI/gTOC) |      |
|                           | range      | mean | range          | mean | range      | mean | range          | mean |
| 4                         | 1.1-6.6    | 3.7  | 11.0-27.0      | 19.3 | -          | -    | -              | -    |
| 3                         | 1.3-1.4    | 1.3  | 14.0-15.0      | 14.5 | 1.4-3.3    | 2.2  | 29.0-39.0      | 34.0 |
| 2                         | 0.8-1.6    | 1.2  | 8.0-10.0       | 9.0  | 0.9-2.5    | 1.2  | 16.0-24.0      | 20.0 |
| 1                         | 0.8-3.3    | 1.6  | 5.0-9.0        | 7.3  | 0.6-3.3    | 1.5  | 9.0-23.0       | 14.7 |

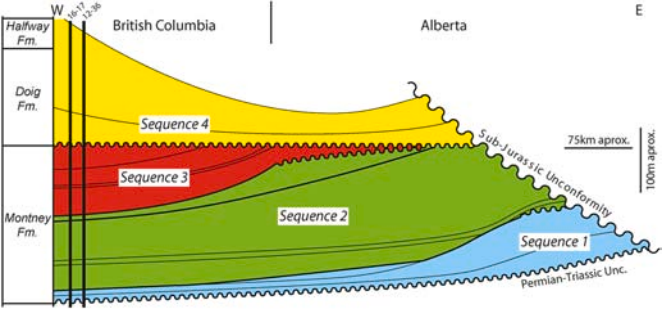
Table 2



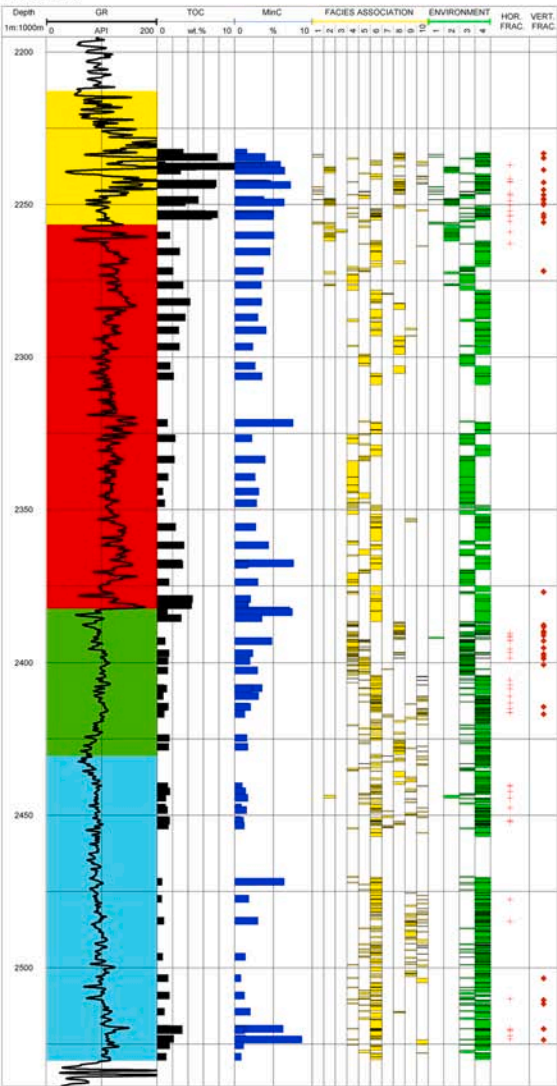
| Sample     | Vein orientation | Mineral phase/<br>texture    | FI type | Occurrence | Fluid phases                             | F         | UV-light response     | Th (°C)     |           | Tm <sub>i</sub> (°C) |            | Salinity (eq. NaCl wt%) |           |
|------------|------------------|------------------------------|---------|------------|--|-----------|-----------------------|-------------|-----------|----------------------|------------|-------------------------|-----------|
|            |                  |                              |         |            |  |           |                       | range       | mode (n)  | range                | mode (n)   | range                   | mode (n)  |
| 1617-23    | vertical         | calcite/<br>blocky           | I       | C/T/I      | L+V                                      | 0.89-0.95 | none                  | 78.0/118.0  | 101 (50)  | -16.0/-19.3          | -16.5 (15) | 19.4/21.9               | 19.8 (15) |
| 1617-23    | vertical         | calcite/<br>blocky           | II      | C/I        | O+G                                      | 0.85/0.90 | green                 | 36.5/45.0   | 41.5 (13) | -                    | -          | -                       | -         |
| 1617-23    | vertical         | calcite/<br>blocky           | III     | T          | O  | 1         | green                 | -           | -         | -                    | -          | -                       | -         |
| 1617-12    | horizontal       | calcite/<br>blocky           | III     | C/I/T      | O  | 1         | green                 | -           | -         | -                    | -          | -                       | -         |
| 1617-12    | horizontal       | calcite/<br>blocky           | IV      | C/T        | L<br>(CH <sub>4</sub> ±CO <sub>2</sub> ) | 1         | yellow/green<br>/none | -           | -         | -                    | -          | -                       | -         |
| C2-1236-2B | horizontal       | calcite/<br>elongated blocky | IV      | C/I        | L<br>(CH <sub>4</sub> ±CO <sub>2</sub> ) | 1         | yellow/green<br>none  | -69.0/-76.0 | -70 (10)  | -                    | -          | -                       | -         |
| C2-1236-2B | horizontal       | calcite/<br>blocky           | III     | C/T/I      | O  | 1         | green                 | -           | -         | -                    | -          | -                       | -         |
| C2-1236-2B | vertical         | calcite/<br>blocky           | III     | C/T/I      | O  | 1         | green                 | -           | -         | -                    | -          | -                       | -         |
| C2-1236-2B | vertical         | calcite/<br>blocky           | IV      | C/T        | L<br>(CH <sub>4</sub> ±CO <sub>2</sub> ) | 1         | green/yellow<br>none  | -62.0/-74.0 | -69 (8)   | -                    | -          | -                       | -         |
| C1-1236-4  | horizontal       | calcite/<br>elongated blocky | II      | Z/C        | O  | 0.80/0.92 | green                 | 24.0/58.0   | 31 (39)   | -                    | -          | -                       | -         |
| C1-1236-4  | horizontal       | calcite/<br>elongated blocky | IV      | Z/C        | L  | 1         | none                  | -66.0/-71.0 | -70 (9)   | -                    | -          | -                       | -         |

Table 3

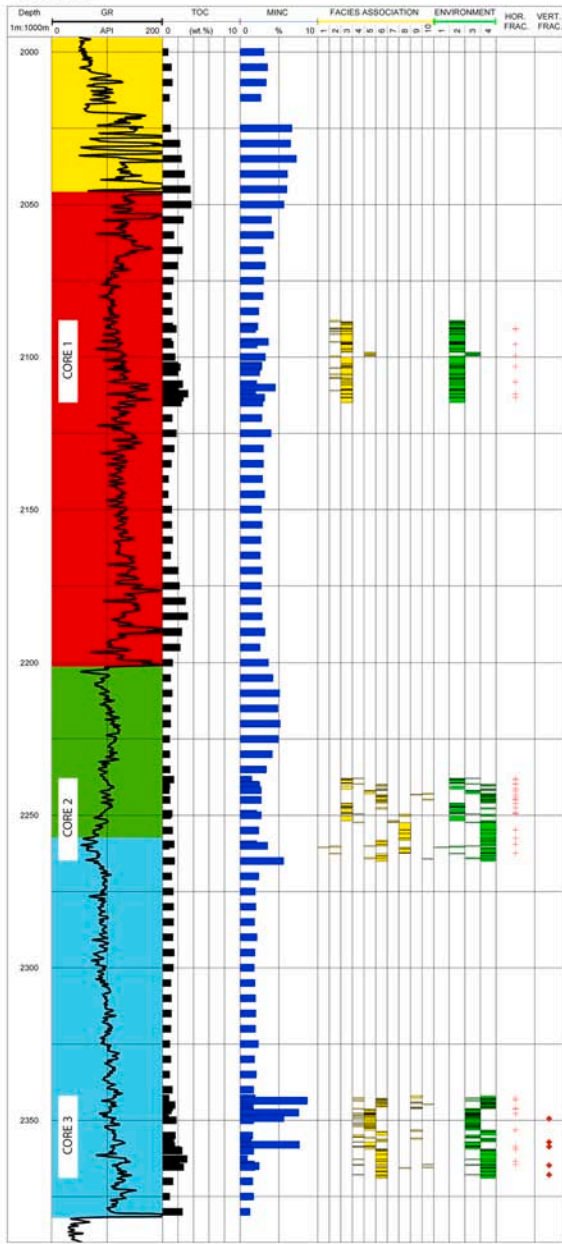


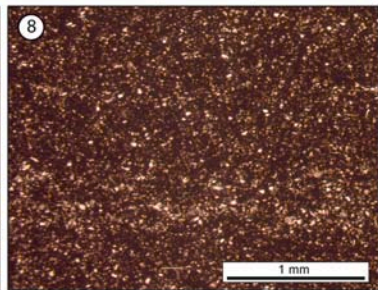
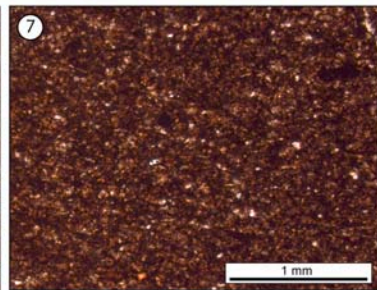
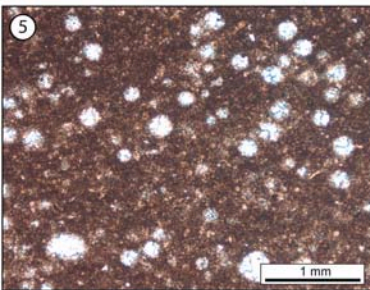
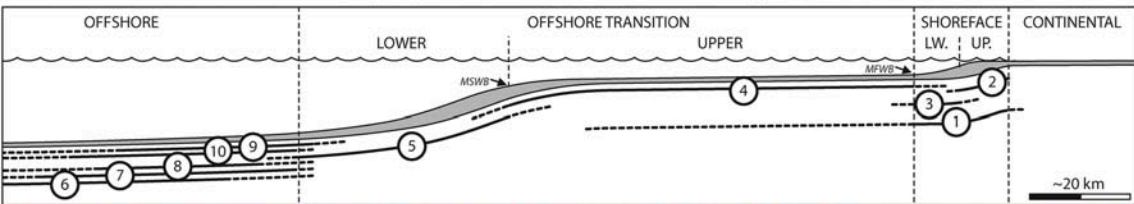
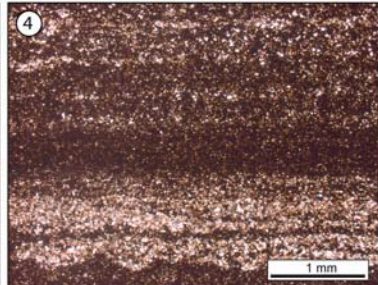
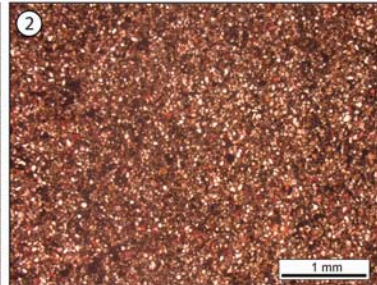
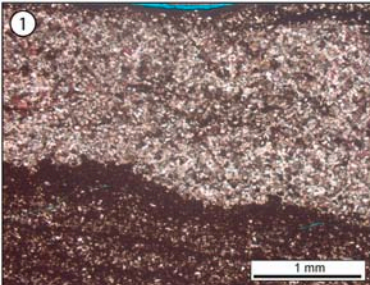


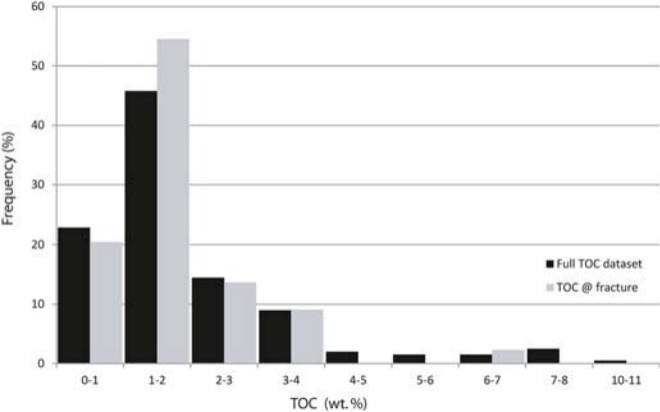
B. Well 16-17

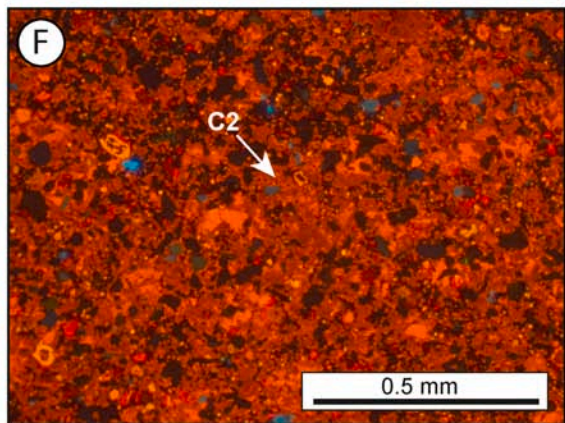
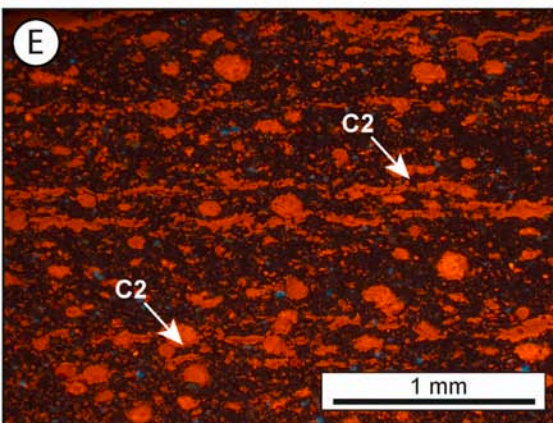
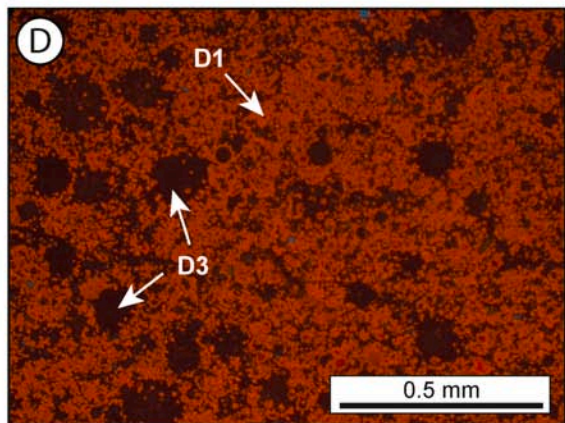
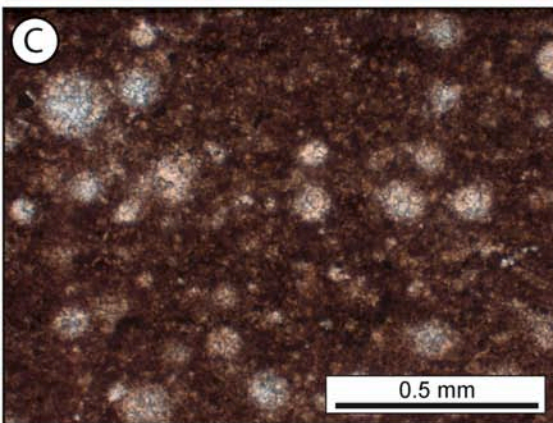
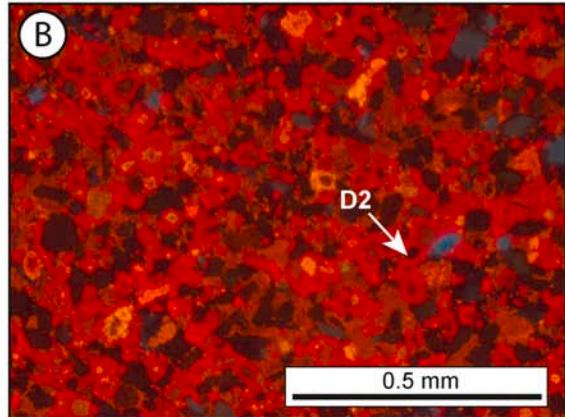
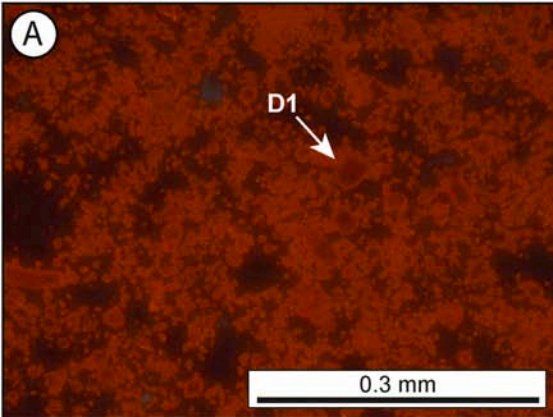


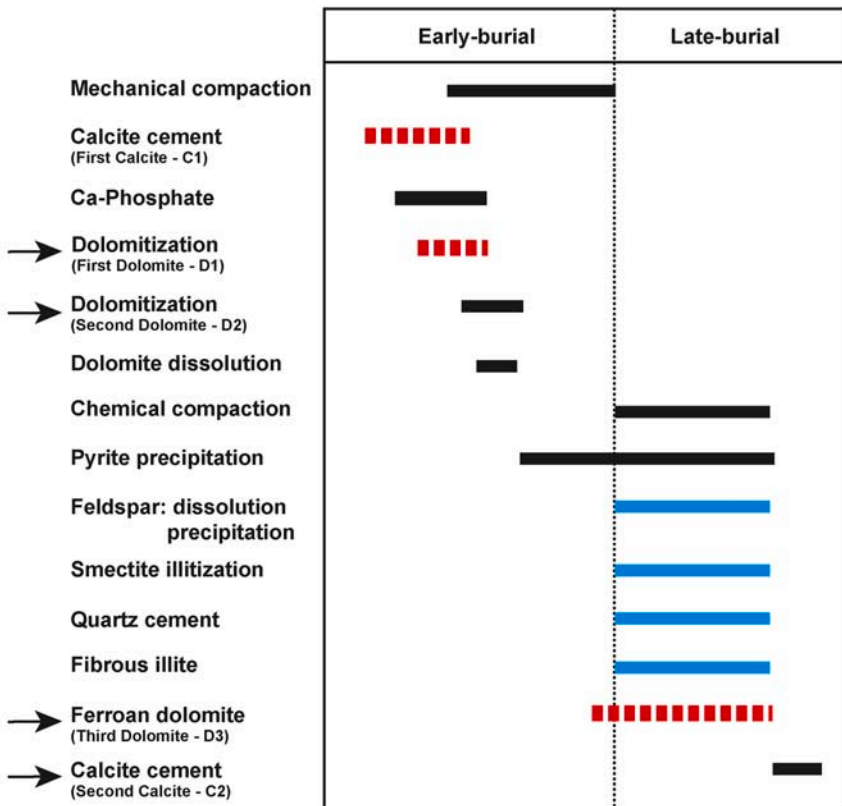
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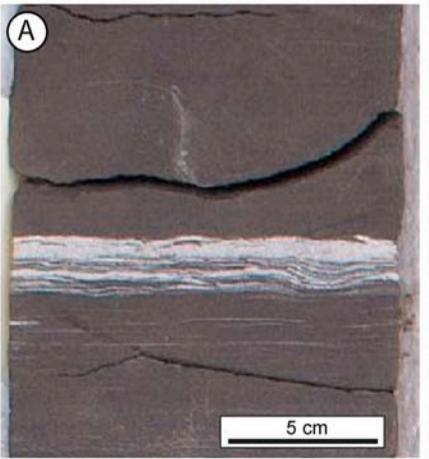




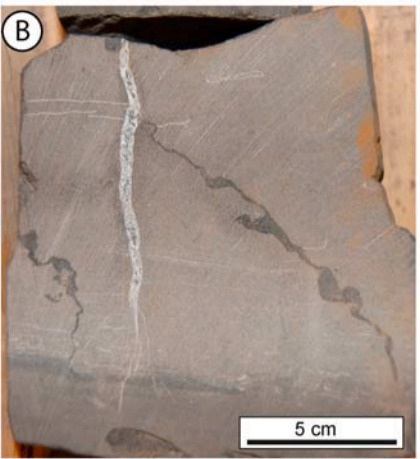




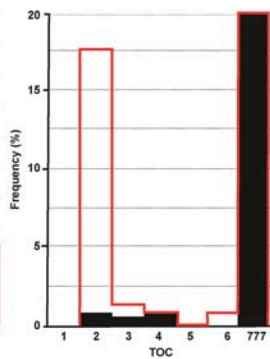
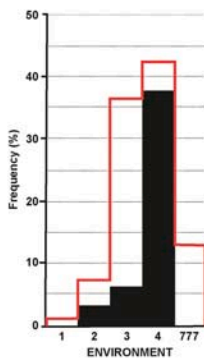
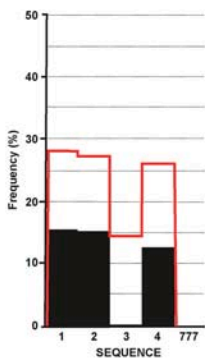
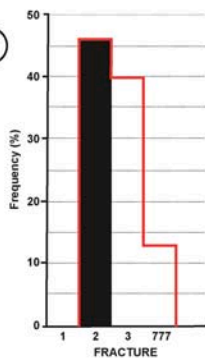
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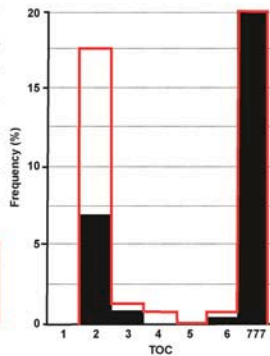
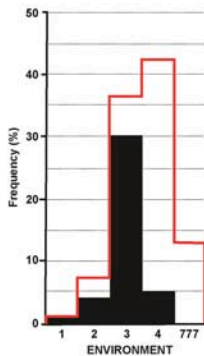
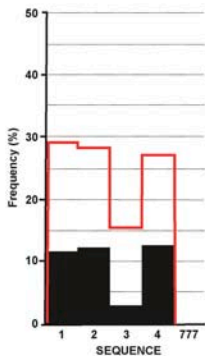
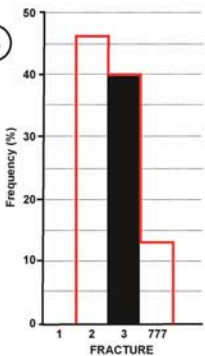
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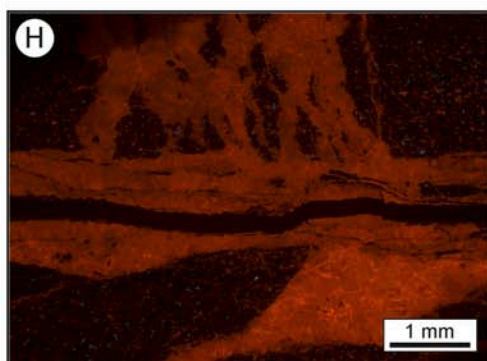
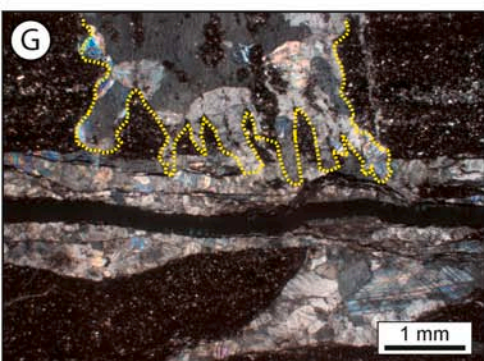
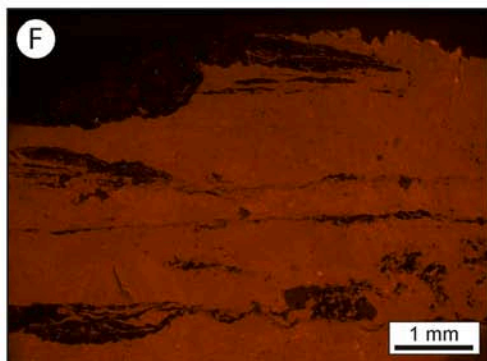
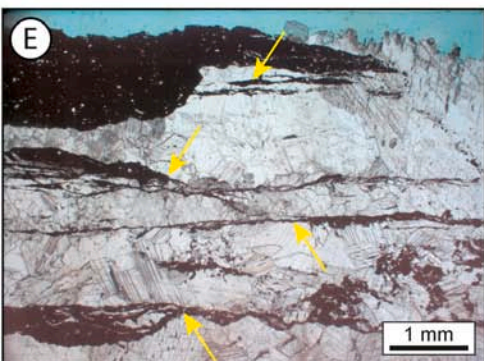
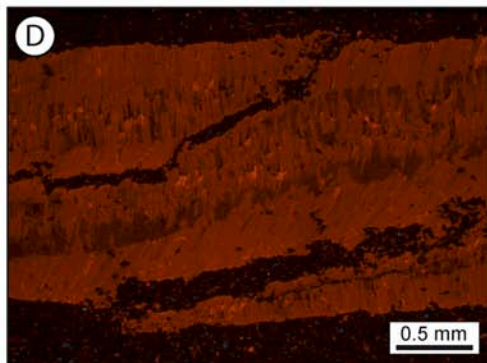
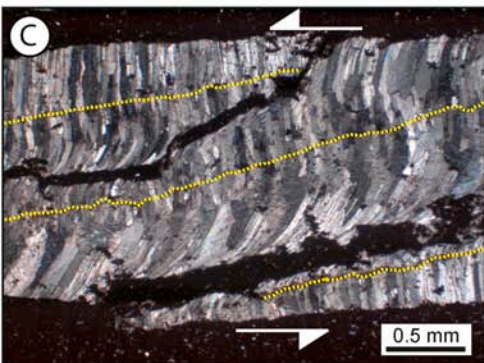
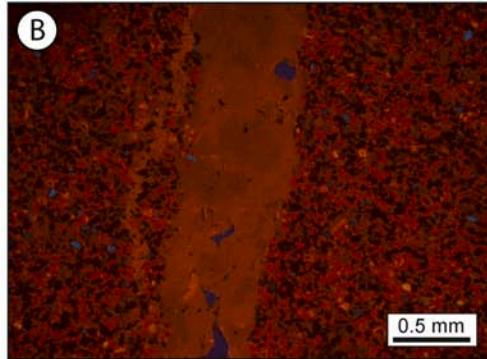
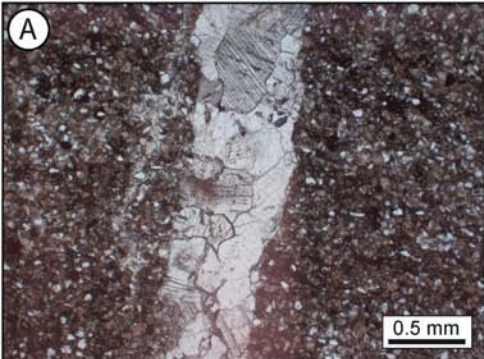


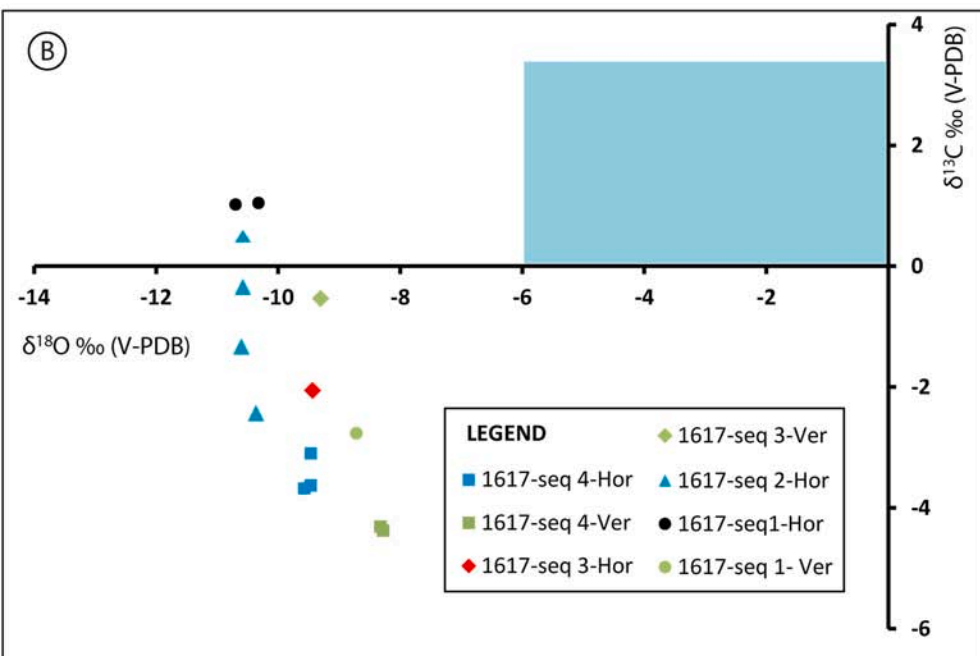
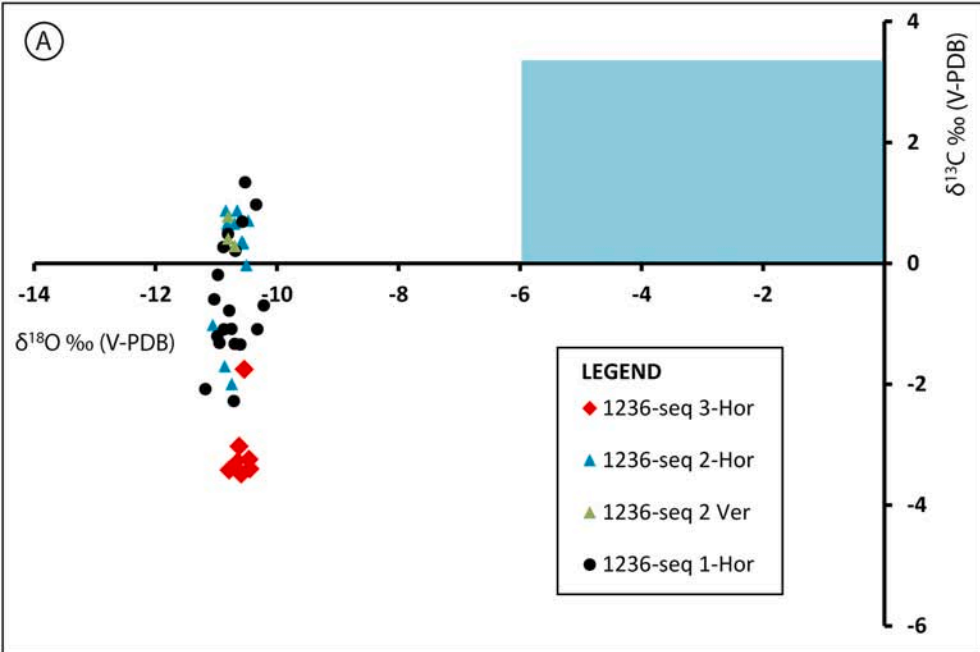
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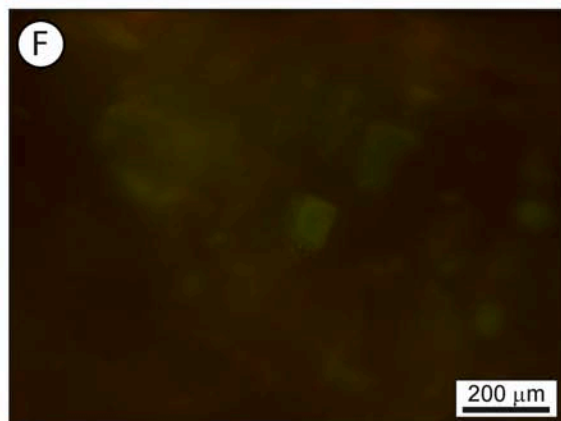
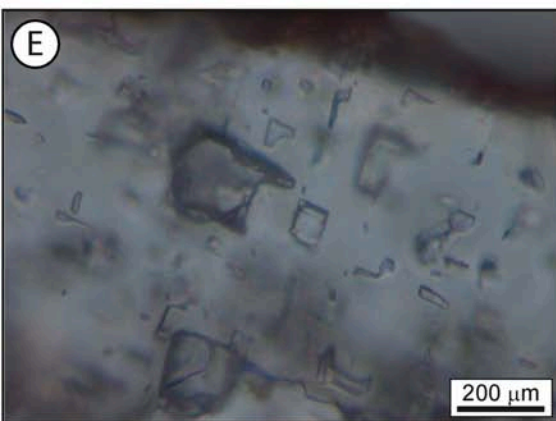
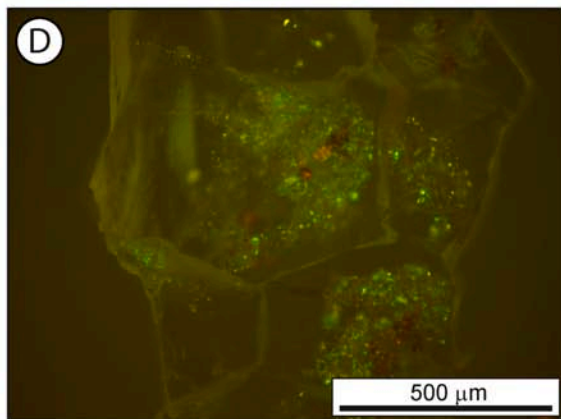
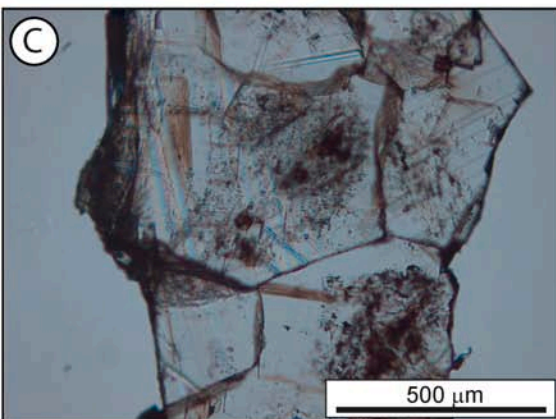
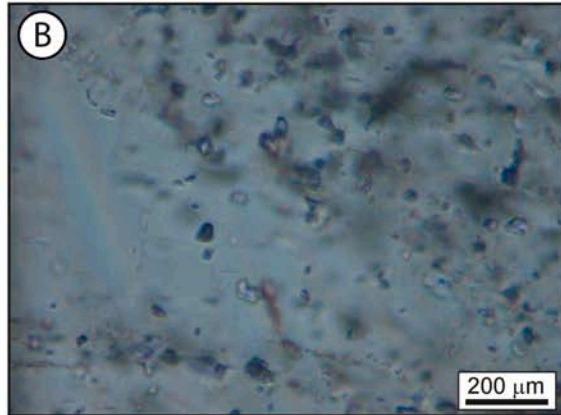
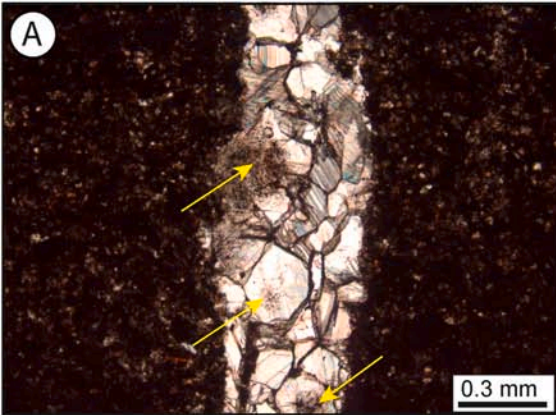


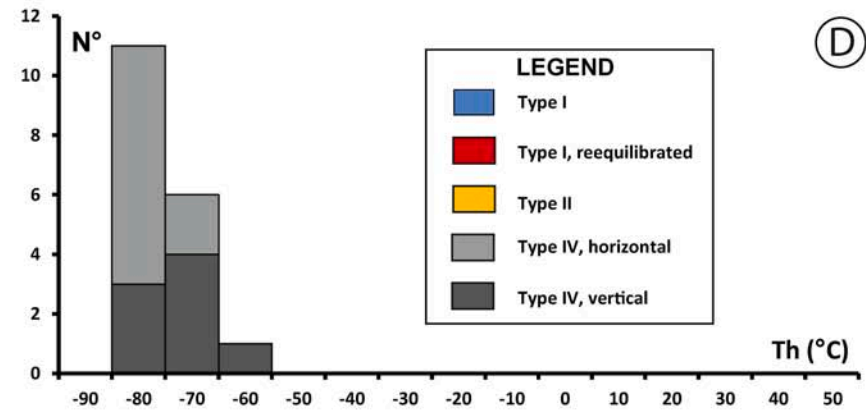
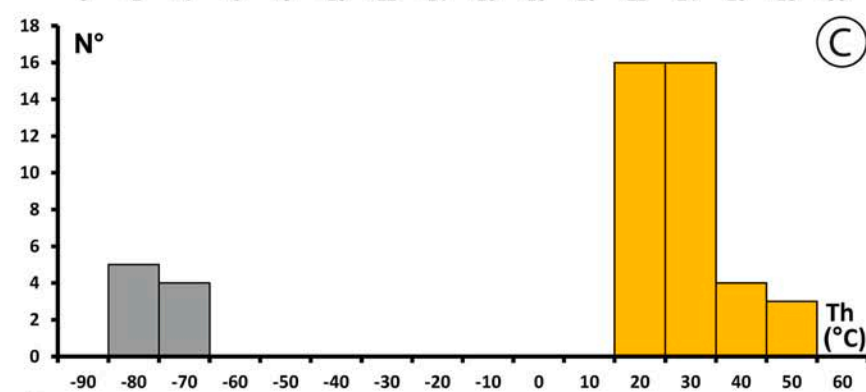
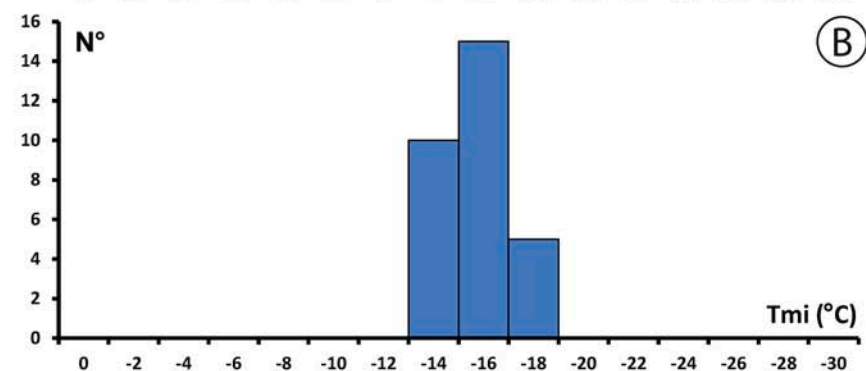
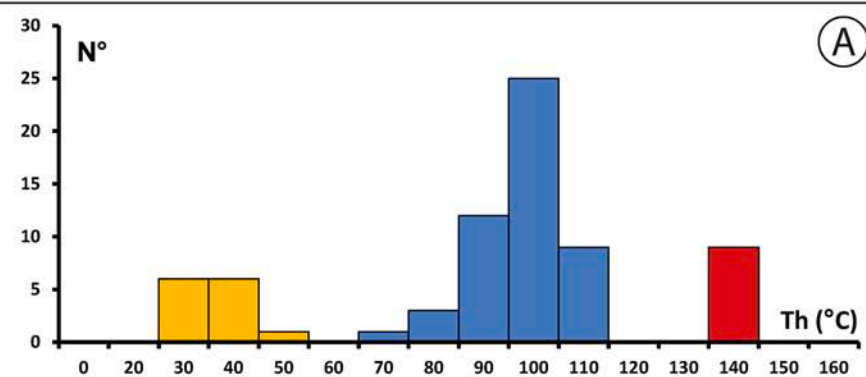
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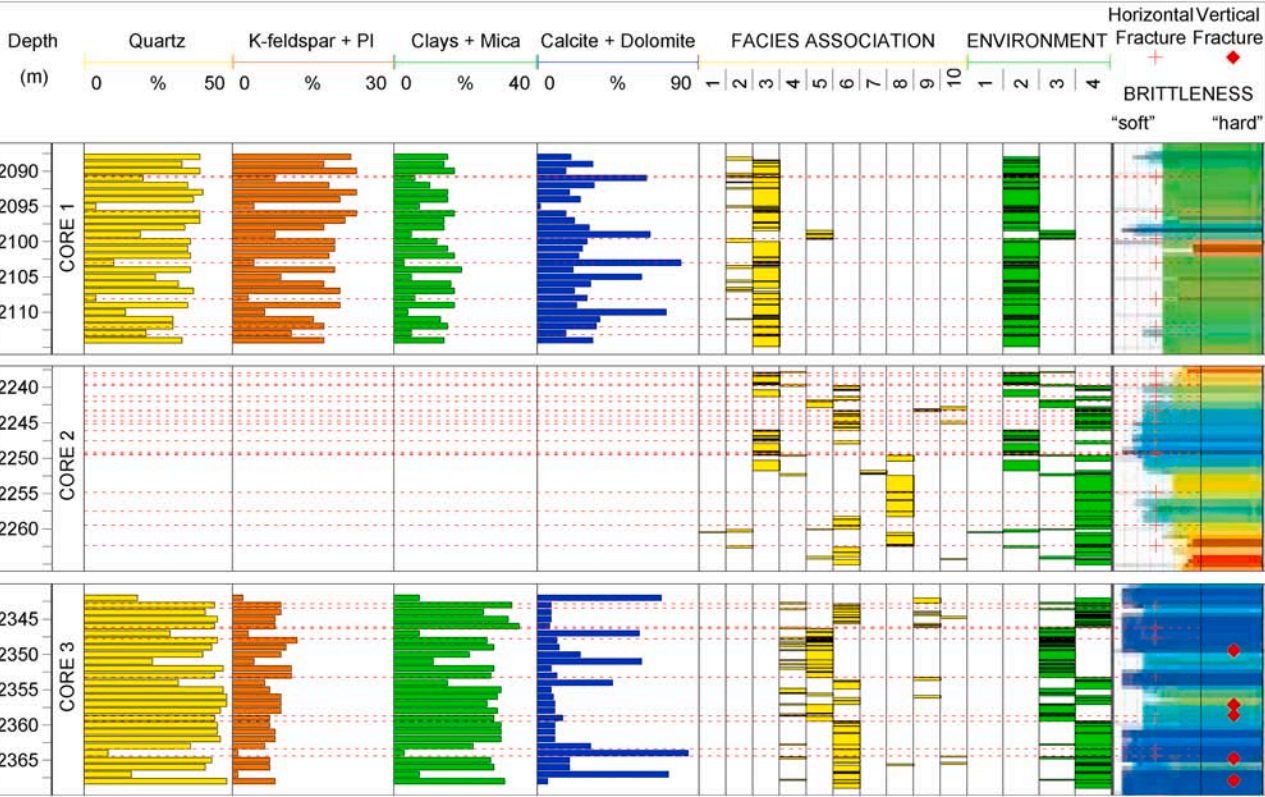


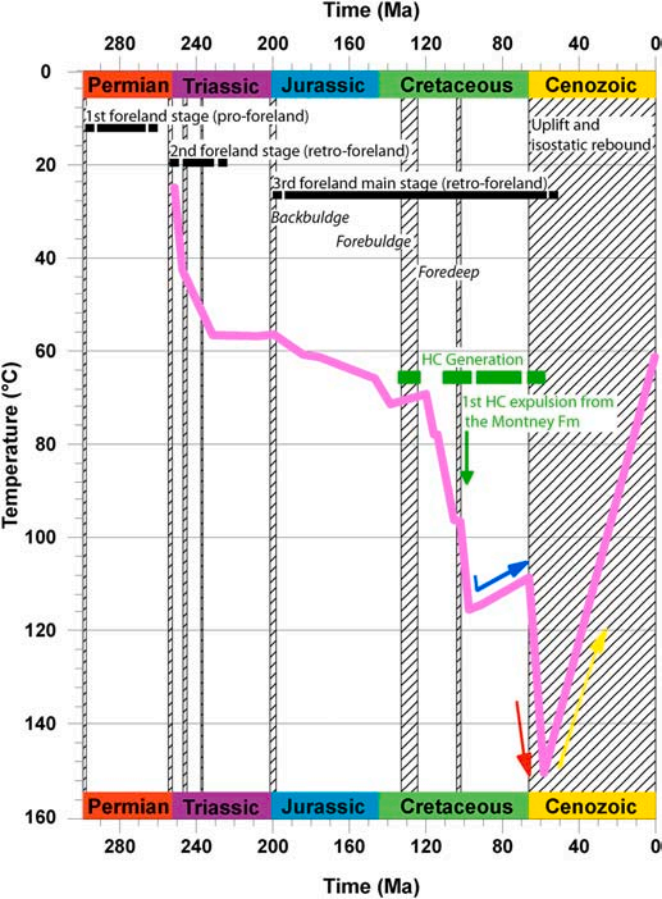














## Highlights

- Fracture diagenesis of Montney-Doig Fms in sub-surface cores was characterized
- Lithology-related factors controlled fracture occurrence
- Timing of fracture opening and role of circulating paleo-fluids were constrained
- The Montney-Doig Fms behaved like a closed hydraulic system through time

Journal Pre-proof

**Declaration of interests**

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

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