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L. Carbillet, L. Griffiths, Michael J. Heap, H. Duwiquet, P. Baud, M. Violay,
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1 **The Influence of Micro- and Macrocracks on the Permeability of Granite**

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3 Carbillet L.^{1*}, Griffiths L.^{1,2}, Heap M.J.^{1,3,4}, Duwiquet, H.⁵, Baud, P.¹, Violay, M.E.S.⁴, Reuschlé, T.¹, and Guillou-
4 Frottier, L.^{6,7}

5

6 ¹*Institut Terre et Environnement de Strasbourg, Université de Strasbourg, CNRS UMR 7063, Strasbourg, France*

7 ²*Norwegian Geotechnical Institute, Oslo, 0806, Norway*

8 ³*Institut Universitaire Français, Paris, France*

9 ⁴*Laboratory of Experimental Rock Mechanics, École Polytechnique Fédérale de Lausanne, Lausanne,*
10 *Switzerland*

11 ⁵*Geothermal Division, ENGIE, Paris la Défense, France*

12 ⁶*BRGM, Av. C. Guillemin, BP 36009 45060 Orléans Cedex 2, France*

13 ⁷*Univ. Orléans, CNRS, BRGM, ISTO, UMR7327, 45071 Orléans, France*

14

15 *Corresponding author: L. Carbillet (lcarbillet@unistra.fr)

16

17 **ORCID**

18 Carbillet L.: 0000-0001-6245-3679

19 Griffiths L.: 0000-0002-7546-2571

20 Heap M.J.: 0000-0002-4748-735X

21 Duwiquet, H.: 0000-0003-0278-3005

22 Baud, P.: 0000-0002-4728-7649

23 Violay, M.E.S.: 0000-0002-7402-8263

24 Reuschlé, T.: 0000-0002-0490-5072

25 Guillou-Frottier, L.: 0000-0002-2698-1447

26

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28

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44

45 **Author Contributions**

46

47 Study conceptualization and design were performed by Lucille Carbillet and Michael J. Heap. Material
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53 the manuscript. All authors read and approved the final manuscript.

54

55 **Highlights**

56

- 57 • Microcracks start forming in Lanhélin granite around 100°C, making the rock samples progressively
58 more porous and permeable but weaker up to 700°C.
- 59 • Under compression, the permeability of microcracked samples initially drops due to closing
60 microcracks and then increases to a residual level when a large fracture forms.
- 61 • During pre-failure deformation in the brittle regime, the permeability of our microcracked granite
62 decreases whereas that of initially intact granite increases.

63

64 **Abstract**

65

66 Damage zones exist around crustal faults, where micro- and macrocracks coexist and form a complex conduit
67 network for fluid flow. The permeability of these zones can change rapidly as the crack network evolves during
68 deformation. To investigate the relative influence of micro- and macrocracks on permeability, we performed an
69 experimental study on Lanhélin granite, consisting of three steps: (1) monitoring thermal microcracking using
70 high-temperature experiments, (2) measuring the evolution of physical properties following thermal stressing, and
71 (3) measuring the permeability of thermally-stressed samples during triaxial deformation. By monitoring acoustic
72 emission activity and P-wave velocity during heating, we find that thermal microcracking starts at ~100 °C and
73 accumulates up to the maximum temperature of 700 °C. Porosity and permeability increase and P-wave velocity,
74 uniaxial compressive strength, and thermal conductivity and diffusivity decrease as thermal stressing temperature
75 increases from room temperature to 700 °C. The axial permeability of thermally-stressed samples decreases by
76 about one order of magnitude during triaxial loading to the peak stress, due to the closure of pre-existing
77 microcracks, and then increases following the formation of a macroscopic shear fracture. Permeability then
78 remains more-or-less constant as strain is accommodated by

79 on the resultant shear fracture. Our results show that the permeability of microcracked granite evolves differently
80 to intact granite, for which permeability increases, during pre-failure deformation in the brittle regime. Such

81 results have important implications for fluid flow in crustal fault systems and their potential for geothermal energy
82 exploitation, which we explore using a simple numerical simulation.

83

84 **Keywords:** Microcrack; Macrocrack; Permeability; Granite; Acoustic emissions

85 1. Introduction

86

87 Cracks exist on a variety of scales, from microcracks on the millimetric scale (Kranz 1983) to crustal-scale faults
88 that can be over 1000 km long (e.g., the San Andreas Fault, USA; Anderson 1971). Micro- and macrocracks are
89 capable of significantly influencing the physical and mechanical properties of rock and rock-masses. Microcracks
90 can decrease the strength and stiffness (David et al. 2012; Wang et al. 2013; Chen et al. 2017; Li et al. 2020; Xu et
91 al. 2023), increase permeability (David et al. 1999; Darot and Reuschlé 2000; Chaki et al. 2008; Feng et al. 2021),
92 decrease P-wave velocity (Reuschlé et al. 2006; Nasserri et al. 2007; Chen et al. 2017; Pimienta et al. 2019), and
93 decrease thermal conductivity (Kant et al. 2017; Zhao et al. 2018) of granite. For example, Griffiths et al. (2017)
94 found that the uniaxial compressive strength of a fine-grained granite decreased from ~290 to ~120 MPa as the
95 number of microcracks per unit area increased from ~80 to ~160 mm⁻². David et al. (1999) found that
96 mechanically- and thermally-induced microcracks can increase the permeability of initially intact granite by up to
97 four orders of magnitude. In terms of macroscopic cracks, Villeneuve et al. (2018) showed that the presence and
98 number of macrocracks can significantly reduce the strength of a granitic rock. Mitchell and Faulkner (2008) and
99 Nara et al. (2018) found, respectively, that the presence of a macroscopic shear and tensile fracture can increase the
100 permeability of granite by up to three to four orders of magnitude.

101

102 Permeability—the ease at which fluid can flow through a porous medium—influences the distribution of pore fluids
103 and pore fluid pressure and therefore exerts a significant influence over crustal faulting and earthquakes (Sibson
104 1996; Jónsson et al. 2003; Wang and Manga 2021). As a result, the permeability of fault rocks and the permeability
105 distribution in fault zones have been closely scrutinised (Caine et al. 1996; Rawling et al. 2001; Faulkner et al.
106 2010). The permeability of crustal rocks and rock-masses also dictates the efficiency of hydrothermal circulation
107 (Hicks et al. 1996; Guillou-Frottier et al. 2013; Vallier et al. 2019; Duwiquet et al. 2019; Bauer et al. 2019) and
108 therefore the productivity of geothermal resources (Huenges and Ledru 2011). Pore pressure perturbations can
109 also create unwanted induced seismicity during geothermal production (Goertz-Allmann et al. 2011; Terakawa et
110 al. 2012) and the flooding and failure of underground engineered spaces (Yin et al. 2015).

111

112 Many of the experimental studies that have sought to measure the permeability of granite during triaxial
113 deformation, such as Mitchell and Faulkner (2008), provide permeability data up to or immediately following the
114 formation of a macroscopic shear fracture. Experiments have also typically been performed on intact granite
115 samples that initially contain very few microcracks while, in many crustal settings, such as fault zones (Mitchell
116 and Faulkner 2012) and geothermal reservoirs (Genter and Traineau 1996), granite can contain abundant
117 microcracks. In this study, we provide a comprehensive experimental dataset for the evolution of physical
118 properties of microcracked granite – representative of rocks found in damage zones adjacent to faults – under
119 elevated pressure and temperature representative of crustal conditions. Therefore, we present a suite of triaxial
120 deformation experiments, performed at different effective pressures, in which we measured the permeability of
121 microcracked granite samples to high strains in the brittle field. We complement these data with experiments in
122 which we measured the output of acoustic emissions, considered a proxy for microcrack growth (Lockner 1993),
123 and the P-wave velocity of granite during heating and cooling, and experiments in which we measure the physical
124 properties (porosity, permeability, P-wave velocity, thermal properties, and uniaxial compressive strength) of

125 granite following exposure to high temperatures (up to 700 °C). Finally, we examined fluid flow characteristics
126 within a crustal fault system through a simple numerical simulation parameterised using our experimental results.
127

128 **2. Materials and methods**

129

130 2.1 Experimental material: Lanhélin granite

131

132 We used Lanhélin granite for our experiments, a coarse-grained, blue-grey granodiorite from Brittany (France).
133 Lanhélin granite contains predominantly quartz, feldspar, and mica, and has an average crystal size of ~1 mm (Fig.
134 1). It was selected here due to its previous use in studies of rock physical properties and rock deformation
135 (Siratovich et al. 2015; Chandler et al. 2017; Wadsworth et al. 2017; Yu et al. 2021). In addition, Lanhélin granite
136 is particularly interesting to study the influence of cracks on permeability because it is prone to forming high-
137 aperture microcracks during thermal stressing (Griffiths et al., 2017). All the samples used in this study (13 in
138 total) were prepared from the same block of Lanhélin granite.

139

140 2.2 Physical property measurements of thermally-stressed samples

141

142 A suite of Lanhélin granite samples, 20 mm in diameter and nominally 40 mm in length, were thermally-stressed in
143 a furnace at room pressure up to target temperatures of 100 (sample LG2), 200 (LG3), 300 (LG4), 400 (LG5), 500
144 (LG6), 600 (LG7), and 700°C (LG8). For each target temperature, one sample was heated up at a rate of 1 °C/min
145 and left at the target temperature for two hours, and finally cooled down to room temperature at 1 °C/min. We then
146 measured the connected porosity, P-wave velocity, permeability, thermal diffusivity, thermal conductivity, and
147 uniaxial compressive strength (UCS) of the thermally-stressed samples at room temperature. The evolution of
148 physical properties in granite with temperature was not the focus of this study and therefore permeability, UCS and
149 thermal properties were voluntarily measured only for selected thermal stressing temperatures. For each target
150 temperature, one thermally-stressed sample served to measure all the physical properties. We also measured the
151 physical properties of an intact sample (i.e., a sample that was not thermally-stressed, sample LG1). In total, 8
152 granite samples were used for these measurements.

153

154 Connected porosity was measured at room pressure and temperature using the skeletal volume measured by a
155 helium pycnometer (AccuPyc II 1340) and the bulk sample volume measured by digital callipers. P-wave velocity
156 was measured using a digital oscilloscope and a pulser/receiver at room pressure and temperature along the length
157 of the samples. P-wave velocities were derived using the length of the rock samples and the travel time of a single
158 period sinusoidal signal with a frequency of 700 kHz through the samples, where P-wave arrival times were picked
159 manually on the oscilloscope. To ensure reproducibility, P-wave velocities were measured under an axial load of
160 300 N. Permeability was measured using a benchtop gas permeameter (see Heap and Kennedy (2016);
161 Farquharson et al. (2016) for a schematic of the experimental device) using either the pulse-decay method (for low
162 permeability samples) or the steady-state method (for high permeability samples). Permeability measurements
163 were made at room temperature, under a confining pressure of 1 MPa, using nitrogen as both the permeant and
164 confining gas. When required, Darcian permeability values were corrected for the effects of fluid-flow related

165 artefacts, i.e., turbulent flow, known as the Forchheimer effect, and gas slip along flow channel walls, known as the
166 Klinkenberg effect (see Heap et al. (2017) for a detailed presentation of the equation used). Thermal diffusivity and
167 thermal conductivity were measured at room pressure and temperature using a Hot Disk® TPS 500 Thermal
168 Constants Analyser using the transient plane source method (Gustafsson 1991; Gustavsson et al. 1994; Harlé et al.
169 2019; Heap et al. 2020, 2022a). Uniaxial compressive strength (UCS) was measured at room pressure and
170 temperature using a uniaxial loadframe (schematics provided in Heap et al. 2014). For each UCS test, the load was
171 applied through an axial steel piston advanced onto the tested sample at a constant displacement rate
172 corresponding to a strain rate of 10^{-5} s^{-1} . The displacement of the piston, monitored using an external linear variable
173 differential transducer (LVDT), was used to derive axial strain. The axial force was measured by a load cell and
174 converted into axial stress using the cross-sectional area of the tested sample. Samples were deformed until
175 macroscopic failure. UCS values correspond to the axial stress at the peak of the stress-strain curves obtained.

176

177 2.3 Output of acoustic emissions and P-wave velocity evolution during heating and cooling

178

179 In the following, we will use *in-situ* to refer to measurements performed live, directly in the experimental
180 apparatus during the heating-cooling cycles. These measurements differ from those made at room temperature on
181 samples that have been heated to a target temperature and cooled back to room temperature, referred to as
182 thermally-stressed samples.

183

184 In-situ measurements of acoustic emissions (AE) were performed on a granite sample (20 mm in diameter and
185 nominally 40 mm in length, sample LG9) during a heating-cooling cycle to 700 °C. The sample was placed in a
186 uniaxial loading frame specifically designed for high-temperature AE measurements (described in detail in
187 Griffiths et al. (2018)) between two vertical stainless-steel pistons, and a tube furnace (GSL 1100X) was mounted
188 around the assembly. The use of a uniaxial load frame served not to deform the samples but to apply a small load
189 ensuring good coupling between the measured sample and the piston and transducers while also improving the
190 repeatability of the measurements. In this apparatus, a broadband acoustic sensor (a micro80 miniature sensor
191 from Physical Acoustics with a bandwidth of 200–900 kHz) is embedded into the upper piston, the top of which is
192 attached to a load cell. The temperatures of the cell and the AE transducer are kept at room-temperature using a
193 specifically designed air-cooling system to avoid damage due to exposure to high temperatures. During the
194 thermal stressing cycle, a load of 100 N was applied using the servo-controlled upper piston to ensure a constant
195 coupling between the granite sample and the AE sensor. As the temperature of the tube furnace was elevated from
196 room conditions to 700°C and back, AE hits were recorded using a USB AE Node from Physical Acoustics
197 connected to the AE sensor. The software AWin was used to monitor the AE activity, with a detection threshold
198 for an AE hit set to 40 dB.

199

200 The same experimental setup (equipped with different platens and transducers) was used to measure in-situ P-wave
201 velocity on another granite sample (20 mm in diameter and nominally 40 mm in length, sample LG10) during a
202 heating-cooling cycle to 400 °C. For these measurements, a high-temperature acoustic sensor (S9215 from
203 Physical Acoustics with a resonant frequency of 100 kHz and an operating frequency of 80–560 kHz) was placed
204 at the end of the upper piston in contact with the sample, to serve as the source of the acoustic signal, and another

205 one was placed at the end of the lower piston in contact with the sample to serve as the receiver. The maximum
206 temperature for in-situ P-wave velocity measurements was limited to 400 °C due to the operating range of the
207 acoustic transducers used in the furnace.

208

209 To summarize, P-wave velocities and the output of AE activity were measured during heating and cooling, as
210 described here above, whilst measurements of the physical properties (e.g., porosity, thermal properties, etc.) were
211 measured on room-temperature samples that had been heated to and cooled from a target thermal stressing
212 temperature (see section 2.2).

213

214 2.4 Permeability evolution during deformation in the brittle regime

215

216 Three cylindrical samples of Lanhélin granite, 37 mm in diameter and 75–80 mm in length (samples LG11-13),
217 were prepared for permeability measurements during deformation in compression under triaxial conditions. To
218 create a network of microcracks within the sample, and to ensure that the permeability was measurable on
219 reasonable laboratory timescales, we first thermally-stressed these three samples to 700 °C at room pressure.
220 Similar to the experiments described above, these samples were first heated to 700 °C at a rate of 1 °C/min, left at
221 700 °C for two hours, and finally cooled down to room temperature at 1 °C/min.

222

223 To ensure that the shear fractures which formed in the samples during the triaxial deformation experiments
224 (described below) were oriented at the same angle with respect to the maximum principal stress and facilitate the
225 comparison between the data from the different experiments, we cut two 1 mm-wide and 4 mm-deep notches on
226 opposite sides of the sample. These notches, orientated at 30° to the maximum principal stress, were made using a
227 precision saw (Secotom Precision Cutting Machine from Struers) and were then filled with a liquid silicone rubber
228 that hardens at room temperature to prevent the confining pressure from puncturing the Viton© jacket.

229

230 Deformation experiments were performed on these samples using the FIRST (Fluid Induced eaRthquake
231 SimulaTor; see Noël et al. 2021; Heap et al. 2022b) triaxial apparatus at the Laboratory of Experimental Rock
232 Mechanics (Ecole Polytechnique Fédérale de Lausanne, Lausanne, Switzerland). Prior to deformation, the
233 samples were vacuum-saturated with deionised water. For each triaxial test, the sample to-be-deformed was placed
234 between two 5 mm-thick pore-fluid spreader plates, slid inside a Viton© jacket, and the whole assembly was
235 mounted inside a pressure vessel (schematic provided in the Supplementary Information). Two separated pairs of
236 servo-controlled pumps were used to control the pressure of the confining fluid (oil), P_c , and that of the pore fluid
237 (deionised water), P_p . The experiments were performed at effective pressures P_{eff} of 10, 30, and 50 MPa, with a
238 constant pore pressure of 10 MPa, and assuming a simple effective pressure law $P_{eff} = P_c - P_p$.

239

240 At the beginning of each test, the confining pressure and the pore pressure were increased to their target values
241 using the servo-controlled pumps and the system was left to equilibrate. Following microstructural equilibration,
242 axial loading was applied by lowering an axial piston on the sample at a constant displacement rate, corresponding
243 to an axial strain rate of 10^{-5} s^{-1} . The position of the piston was controlled and recorded using an external LVDT
244 attached to its top. Two internal LVDTs and the initial sample length were used to calculate axial strain, while the

245 pressure acting on the piston and the initial sample cross-section were used to calculate axial stress. Differential
246 stress was corrected for the friction at the interface between the axial piston and the walls of the apparatus column.
247 During deformation, only one pore pressure pump was used to maintain a pressure of 10 MPa, while the other
248 pump was disconnected from the circuit in order to ensure accuracy in the measurement of the pore volume
249 change. The porosity change was calculated from the change in pore volume (given by the displacement of the
250 piston in the pore pressure intensifier during the experiment) and the initial bulk sample volume. During
251 deformation, fully drained conditions were ensured at the imposed strain rate given the permeability and
252 dimensions of the samples.

253

254 During each triaxial test, the permeability of the sample was measured first under hydrostatic conditions (before
255 any axial loading was applied) and then intermittently during deformation. The permeant fluid for these
256 permeability measurements was the pore fluid, i.e., deionised water. For each permeability measurement, the
257 loading was halted (except for the first measurement under hydrostatic conditions) and both the upstream and
258 downstream pore fluid pumps were connected to the circuit. The upstream pore pressure pump was set at 10.1 MPa
259 while the downstream one was set at 9.9 MPa so that a pore pressure difference of 0.2 MPa was imposed between
260 the two ends of the sample. The resulting movement of water through the sample was tracked and the
261 corresponding volumetric flow rate was used to calculate permeability using Darcy's law. Steady-state flow, i.e.,
262 when the displacement of the upstream and downstream pore pressure pumps was constant, was typically achieved
263 in several minutes. After several tens of minutes at constant flow rate, one of the pore pressure pumps was isolated
264 from the pore pressure circuit while the other was set back to a constant pore pressure of 10 MPa and loading was
265 continued. During all three triaxial tests, permeability was measured using this method up to an axial strain of
266 0.09–0.10. A detailed example of the method used to compute permeability is presented in the Supplementary
267 Information (as an accompanying Excel file). Permeability measurements were conducted along the axial
268 direction of the cylindrical samples only (not along the fracture plane generated during triaxial compression) and
269 considered the change in sample length. The permeability values obtained correspond to the equivalent
270 permeability of the bulk samples which, at certain stages of the deformation, can include the contributions of the
271 potentially more permeable fracture and less permeable matrix. In the following, we will refer to our bulk axial
272 permeability data simply as *permeability*. It should be noted that the measured permeability represents a lower
273 bound permeability since microcracks in the rock samples likely undergo relaxation when loading is halted to
274 perform the measurements.

275

276 3. Results

277

278 3.1 Petrophysical and microstructural evolution during thermal stressing

279

280 The porosity, P-wave velocity, permeability, uniaxial compressive strength, thermal diffusivity, and thermal
281 conductivity of Lanhélin granite are plotted as a function of thermal-stressing temperature in Fig 2. These data
282 show that porosity and permeability increase, whereas P-wave velocity, uniaxial compressive strength, thermal
283 diffusivity, and thermal conductivity decrease as a function of thermal-stressing temperature (Table 1). All of the

284 measurements in this section were performed at room temperature on samples that had been heated to and cooled
285 from a target thermal stressing temperature.

286

287 Increases in porosity start at a thermal-stressing temperature above 200 °C (Fig. 2a). Porosity is increased from
288 ~ 0.005 to ~ 0.01 as thermal-stressing temperature is increased from 200 to 500 °C. Increases in porosity are much
289 greater at temperatures above 500 °C and porosity increases from ~ 0.01 at 500 °C to ~ 0.03 at 700 °C. P-wave
290 velocity decreases more-or-less linearly as thermal-stressing temperature is increased from room temperature up
291 to 700 °C (Fig. 2b). P-wave velocity is reduced from ~ 6 km/s at room temperature down to ~ 1.5 km/s at a thermal-
292 stressing temperature of 700 °C. Permeability increases exponentially as a function of thermal-stressing
293 temperature (Fig. 2c). Permeability increases from $\sim 10^{-20}$ m² at room temperature to $\sim 10^{-14}$ m² at a thermal-
294 stressing temperature of 700 °C, an increase by about six orders of magnitude, at ambient pressure. Uniaxial
295 compressive strength remains quasi constant at a thermal-stressing temperature of 300 °C (uniaxial compressive
296 strength is ~ 150 – 160 MPa) (Fig. 2d). At a thermal-stressing temperature of 700 °C, the uniaxial compressive
297 strength is decreased from ~ 150 – 160 MPa to ~ 65 MPa. Mechanical data for the UCS tests are available in
298 Supplementary Information. Thermal conductivity and thermal diffusivity remain almost unchanged to a thermal-
299 stressing temperature of 300 °C (Figs. 2e and 2f). Above a thermal-stressing temperature of 300 °C, thermal
300 conductivity and thermal diffusivity decrease up to the maximum thermal-stressing temperature of 700 °C.
301 Thermal diffusivity decreases from ~ 2 mm²·s⁻¹ at temperatures of 25–300 °C to ~ 1 mm²·s⁻¹ at a thermal-stressing
302 temperature of 700 °C. Thermal conductivity decreases from ~ 3 – 3.5 W·m⁻¹·K⁻¹ at temperatures of 25–300 °C to
303 ~ 1.75 W·m⁻¹·K⁻¹ at a temperature of 700 °C.

304

305 Cumulative AE hits and P-wave velocity, measured under in-situ conditions, as a function of temperature during
306 the thermal-stressing of samples of Lanhélin granite are shown in Fig. 3 (these data were collected on different
307 samples, but we show them here together for ease of comparison). We note that the output of AE activity during the
308 heating of the sample starts at ~ 100 °C (solid red line). The number of AE hits increases as a function of
309 temperature above ~ 100 °C, more rapidly above ~ 450 °C and then with a noticeable increase at a temperature of
310 ~ 550 – 600 °C. During the cooling of the sample, there is a steady increase in the number of AE hits from 700 °C
311 down to room temperature (solid blue line). In-situ P-wave velocity decreases during the heating of the sample, the
312 rate of which increases as the sample is heated above ~ 100 °C (dashed red line). During heating to 400 °C, the P-
313 wave velocity of the sample is reduced from ~ 6 km/s down to ~ 3 km/s. During cooling, the P-wave velocity of the
314 sample increases from ~ 3 km/s at a temperature of 400 °C to ~ 3.5 km/s at room temperature (dashed blue line).

315

316 We also determined the microcrack surface area per unit volume and the 2D microcrack density for samples of
317 Lanhélin granite thermally-stressed to 300 and 700 °C, as well as for an intact sample. Optical microscope images
318 of intact Lanhélin granite, and samples thermally-stressed to 300 and 700 °C, are shown in Figs. 4a, 4b, and 4c,
319 respectively. Qualitatively, we observe an increase in the number of microcracks as the sample is thermally-
320 stressed to 300 and 700 °C. We calculate the microcrack surface area and the 2D microcrack density for each of
321 these images using the automated segmentation procedure for optical micrographs presented and described in
322 Griffiths et al. (2017). These results are presented in Fig. 4d. We find that the microcrack surface area increases
323 from ~ 0.8 mm⁻¹ at room temperature to ~ 6 mm⁻¹ at a thermal stressing temperature of 700 °C (white circles). 2D

324 microcrack density increases from ~ 0.025 to ~ 0.17 as thermal-stressing temperature is increased from room
325 temperature to $700\text{ }^{\circ}\text{C}$ (blue diamonds).

326

327 3.2 Mechanical data and permeability changes in the brittle regime

328

329 In the following, we adopt the convention that compressive stress and compactive strain are positive. We will first
330 describe the evolution of porosity and permeability under hydrostatic conditions (up to 50 MPa), before describing
331 the mechanical data and the evolution of porosity and permeability as a function of axial strain (at effective
332 pressures of 10, 30, and 50 MPa).

333

334 The evolution of porosity and permeability as a function of effective pressure is shown in Figs. 5a and 5b,
335 respectively (data available in Table 2). We find that porosity and permeability decrease non-linearly as a function
336 of increasing effective pressure. In both cases, the decreases in porosity and permeability per unit effective
337 pressure are larger at low effective pressure. Over the entire pressure range (up to 50 MPa), porosity decreased by
338 ~ 0.008 and permeability decreased from $6.9 \times 10^{-16}\text{ m}^2$ at an effective pressure of 2 MPa to $5.3 \times 10^{-18}\text{ m}^2$ at an
339 effective pressure of 50 MPa.

340

341 The stress-strain curves and porosity change as a function of axial strain for the three triaxial experiments (at
342 effective pressures of 10, 30, and 50 MPa) performed on thermally-stressed samples of Lanhélin granite are shown
343 in Figs. 6a and 6b, respectively. The three experiments can all be classed as brittle: the stress increases to a
344 maximum (the peak stress), followed by a stress drop associated with macroscopic sample failure (the formation of
345 a shear fracture). Following fracture formation, the stress remained more-or-less constant as deformation was
346 accommodated by sliding on the formed fracture. While differential stress increases almost linearly to a maximum,
347 porosity change decreases (compaction) and then switches to increasing, which marks the onset of dilatancy. The
348 differential stress at which this onset of dilatancy occurs is referred to as C' (Paterson and Wong 2005). We find
349 that (1) the peak stress increases with increasing effective pressure, from ~ 200 to ~ 500 MPa as effective pressure is
350 increased from 10 to 50 MPa, (2) the initial nonlinear portion of the stress-strain curves, usually attributed to
351 microcrack closure, is more pronounced at 10 MPa, and (3) the residual differential stress, the frictional sliding
352 stress, increases as a function of effective pressure. The porosity change curves show that porosity first decreases
353 (compaction) during the initial stages of loading, before increasing (dilatancy) as the sample approaches the peak
354 stress. The differential stress values for the onset of inelastic deformation, referred to as C' , are 42, 76, and 127
355 MPa at effective pressures of 10, 30 and 50 MPa, respectively. We measure a large increase in sample porosity
356 following the formation of the macroscopic shear fracture; porosity then increases during sliding on the formed
357 fracture. We note that the decrease in porosity during the initial stages of loading is higher for the sample deformed
358 at 10 MPa, and that the increase in porosity following shear fracture formation decreases as effective pressure
359 increases.

360

361 The evolution of permeability as a function of axial strain for the three triaxial experiments shown in Fig. 6 are
362 provided in Fig. 7 (the mechanical data, permeability data and pictures of the deformed samples are available in the
363 Supplementary Information), alongside the stress-strain curves. We first note that the starting permeability (under

364 hydrostatic conditions, before any axial loading is applied, i.e., at a differential stress of 0 MPa) is 1.1×10^{-16} , $1.2 \times$
365 10^{-17} , and $5.3 \times 10^{-18} \text{ m}^2$ at effective pressures of 10, 30, and 50 MPa, respectively. For all three experiments, the
366 permeability decreases as stress is increased up to the peak stress. Following the formation of a macroscopic shear
367 fracture, the permeability increases, before remaining more-or-less constant as the strain is accommodated by
368 sliding on the formed fracture. The permeabilities of the samples at the end of the experiments, following
369 deformation to an axial strain of ~ 0.1 , are 1.4×10^{-17} , 3.3×10^{-18} , and $9.9 \times 10^{-19} \text{ m}^2$ at effective pressures of 10, 30,
370 and 50 MPa, respectively. Therefore, the permeabilities of the samples at the end of the experiments are lower than
371 at the start of the experiments.

372

373 4. Discussion

374

375 In the following, the evolution of petrophysical properties and microstructure during heating and cooling is first
376 discussed. We then discuss the permeability evolution induced by triaxial compression up to large axial strains in
377 thermally cracked Lanhélin granite and compare these data to the permeability evolution during inelastic
378 deformation of other crystalline and non-crystalline rocks. We relate our results to the evolution of the basement
379 and damage zone adjacent to crustal faults, where past or present activity on the fault promotes stress,
380 displacement and/or strain increases that, in turn, can lead to fracturing. Based on our results, we discuss how the
381 elevated pressure and temperature conditions within these zones might influence permeability. Finally, we use our
382 experimental results to constrain a numerical simulation of fluid flow within a granitic crustal fault system to
383 discuss their potential as geothermal reservoirs.

384

385 4.1 The influence of thermal-stressing on the petrophysical properties and microstructure of granite

386

387 The in-situ measurements provide additional information to measurements performed periodically on samples
388 during heating and cooling, to help us to better interpret the microstructural and petrophysical changes associated
389 with thermal stressing. For example, the AE and P-wave velocity in-situ measurements performed on Lanhélin
390 granite samples allow us to provide a more precise onset temperature for thermal microcracking and to gain
391 insights into how microcracking progresses throughout heating and cooling. We find that the onset of AE activity,
392 considered a proxy for microcrack growth (Lockner 1993), in Lanhélin granite occurs at $\sim 100 \text{ }^\circ\text{C}$ and that the
393 number of AE hits continually increases as the granite is heated to $700 \text{ }^\circ\text{C}$ (Fig. 3). Our data also show that P-wave
394 velocity decreases as temperature increases (P-wave velocity data are only available up to $400 \text{ }^\circ\text{C}$ due to the
395 maximum working temperature of the P-wave transducers). The onset of AE activity during the heating of granite
396 is observed to be $70\text{--}120 \text{ }^\circ\text{C}$ (Yong and Wang 1980; Hall and Bodnar 1989; Griffiths et al. 2018), in accordance
397 with our onset temperature of $\sim 100 \text{ }^\circ\text{C}$ for Lanhélin granite. The continuous AE activity as the granite is heated
398 above the onset temperature for thermal microcracking is also in line with previous studies that measured the
399 output of AE during the heating of granite (e.g., Glover et al. 1995; Griffiths et al. 2018). We interpret the AE
400 activity during the heating of Lanhélin granite to be the result of the initiation of growth of thermal microcracks, as
401 observed in our thin section analysis (Fig. 4), and as discussed in the published works mentioned above. Thermal
402 microcracks initiate and grow due to the stresses resulting from the differential thermal expansion of the mineral
403 constituents of the granite (quartz, feldspar, and mica; Figs. 1 and 4, Fredrich and Wong (1986)). This conclusion

404 is corroborated by the measured decrease in P-wave velocity during heating (Fig. 3) and following exposure to
405 high temperature (Fig. 2). Large decreases in the P-wave velocity of granite have been previously observed to
406 result from an increase in thermal microcracking (Reuschlé et al. 2006; Nasserri et al. 2007; Griffiths et al. 2017).
407 The decrease in P-wave velocity as Lanhélin granite is heated to 450 °C, from ~6 to ~3 km.s⁻¹, is similar to the
408 decrease measured in Westerly granite using the same setup and method (Griffiths et al. 2018). Griffiths et al.
409 (2018) found that the P-wave velocity of granite is reduced from ~5 to ~2.5 km.s⁻¹ as temperature is increased from
410 room temperature to 450 °C. The noticeable increase in AE activity at ~550–600 °C is considered to be the result
411 of the α/β transition of quartz that occurs at ~573 °C and results in the large volumetric expansion of quartz and
412 therefore an increase in crystal-to-crystal stresses (Glover et al. 1995). Decrepitation of fluid inclusions – cracking
413 at grain-fluid contacts due to thermal stressing – potentially also contributed minorly to the AE activity measured
414 during heating (Hall and Bodnar 1989). For a more systematic investigation of microstructural evolution of granite
415 during thermal stressing, we direct the reader to Meredith et al. (2001).

416

417 The cooling of Lanhélin granite is associated with a continuous increase in AE activity and an increase in P-wave
418 velocity (Fig. 3). An increase in P-wave velocity during the cooling of granite was also observed by Griffiths et al.
419 (2018) and was interpreted as the closing of newly formed thermal microcracks that were propped open at high-
420 temperature. However, we also record a continuous increase in AE activity during cooling, as previously observed
421 in experiments performed on granite (Griffiths et al. 2018) and volcanic rocks (Heap et al. 2014; Browning et al.
422 2016; Daoud et al. 2020; Griffiths et al. 2024). Considering that an increase in thermal microcracking is well-
423 known to reduce the P-wave velocity of granite (Reuschlé et al. 2006; Nasserri et al. 2007; Griffiths et al. 2017), it is
424 very unlikely, based on the P-wave velocity data of Fig. 3, that thermal microcracking is occurring during the
425 cooling of the sample. As discussed in Griffiths et al. (2018), we consider that the majority of the AE hits recorded
426 during the cooling of Lanhélin granite is not associated with newly formed or propagating microcracks, but with
427 the friction on the surfaces of microcracks which partially close up as the rock cools and contracts. We highlight
428 that contemporaneous P-wave velocity measurements are important for the interpretation of AE activity during the
429 heating and cooling of rock samples.

430

431 In addition to the observations made using the in-situ measurements, we find that the porosity and permeability of
432 Lanhélin granite increase, whereas P-wave velocity, uniaxial compressive strength, thermal diffusivity, and
433 thermal conductivity decrease as a function of thermal-stressing temperature (Fig. 2). In detail, porosity, strength,
434 thermal diffusivity and conductivity are more-or-less constant below 300°C and then decrease drastically up to
435 700°C. These changes are interpreted as the consequence of the increase in the number of thermal microcracks as a
436 function of thermal stressing temperature (Fig. 4). Thermal microcracks have previously been observed to
437 decrease the strength and stiffness (Alm et al. 1985; David et al. 2012; Wang et al. 2013; Griffiths et al. 2017; Chen
438 et al. 2017), increase the permeability (Darot et al. 1992; David et al. 1999; Darot and Reuschlé 2000; Chaki et al.
439 2008), decrease the P-wave velocity (Reuschlé et al. 2006; Nasserri et al. 2007; Chen et al. 2017), and decrease the
440 thermal conductivity (Kant et al. 2007; Zhao et al. 2018) of granite.

441

442 4.2 The evolution of permeability during brittle faulting: comparison to other crystalline rocks

443

444 We find that the porosity and permeability of thermally-stressed Lanhélin granite decrease as a function of
445 hydrostatic pressure (Fig. 5). These data are in agreement with published data for thermally-stressed granites
446 (Darot et al. 1992; Darot and Reuschlé 2000; Nasseri et al. 2009), and naturally microcracked (Vinciguerra et al.
447 2005; Fortin et al. 2011) and thermally-stressed basalts (Nara et al. 2011). The reduction in porosity and
448 permeability as a function of effective pressure measured here is interpreted as the result of the progressive closure
449 of thermal microcracks (generated during the thermal treatment to 700 °C; see Fig. 4c).

450

451 Under triaxial compression at low confining pressure and room temperature, the permeability of microcracked
452 Lanhélin is found to decrease up to the peak stress and then to increase back to a residual value that remains
453 constant during sliding on the formed shear fracture (Fig. 7). The difference in permeability between intact and
454 faulted samples has been previously measured on low-porosity crystalline rocks in the laboratory. For initially
455 intact La Peyratte granite, Acosta and Violay (2020) reported a decrease in apparent permeability from 6.0×10^{-20}
456 to 1.3×10^{-20} m² after hydrostatic loading up to a confining pressure of 25 MPa, and then an increase to 9.0×10^{-20}
457 m² following macroscopic brittle failure. Zoback and Byerlee (1975) measured the evolution of the permeability of
458 Westerly granite during triaxial compression ($P_C = 25$ MPa, $P_p = 11$ MPa) and showed that the permeability of
459 the deformed sample was twice the permeability of the intact sample. Oda et al. (2002) reported a larger increase in
460 permeability for samples of Inada granite loaded up to brittle failure, where the permeability of the sample post-
461 brittle failure was about two to three orders of magnitude larger than that of the intact granite under the same
462 confining pressure ($P_C = 140$ MPa).

463

464 Critical stress-path dependent permeability variations can also result from microstructural changes that occur in
465 the pre-failure stage of deformation, in particular during the interval between the onset of inelastic deformation
466 C' and macroscopic brittle failure (Faulkner and Armitage 2013). In contrast to our findings on microcracked
467 Lanhélin granite, the permeability of crystalline rocks that initially contain very few microcracks can be
468 significantly higher after failure, by up to several orders of magnitude, compared to the pre-failure state (Oda et al.
469 2002; Chen et al. 2014; Kluge et al. 2021). In detail, the permeability decreases a little as a result of microcrack
470 closure during the early stages of loading (i.e. at low stress), and then remains almost constant until it starts
471 increasing after the differential stress reaches the critical value required for the onset of inelastic deformation. At
472 stresses higher than the onset of dilatancy, microcracks start forming and propagating and, as they start to coalesce
473 to form a macroscopic shear fracture, large increases in permeability are observed, particularly because
474 permeability is measured in the direction of the crack growth. The permeability of a sample of Odenwald granite
475 deformed at an effective pressure of 20 MPa, for example, was initially equal to 1×10^{-18} m², decreased to less than
476 10^{-18} m² as the sample is loaded at low values of stress, before increasing to more than 10^{-17} m² when the stress
477 exceeded the critical value for the onset of dilatancy (Kluge et al. 2021). The permeability of the sample of
478 Odenwald granite then increased to more than 10^{-16} m² following the stress drops associated with sample failure
479 (Kluge et al. 2021). Similar results were reported by Chen et al. (2014) for Beishan granite samples, whose
480 permeability increased by up to three orders of magnitude during the growth of a cross-cutting shear fracture under
481 triaxial loading at confining pressures of 5, 8, and 10 MPa. In detail, they showed that the relative evolution of
482 permeability during brittle deformation was the same for the different confining pressures but that, at a given
483 differential stress, the higher the confining pressure, the lower the permeability. These results are in agreement

484 with those of Mitchell and Faulkner (2008) for Westerly granite, the permeability of which increased by up to two
 485 orders of magnitude as the microfracture density (estimated using image analysis) increased by more than a factor
 486 of five during cyclic loading to macroscopic failure.

487

488 In comparison, the permeability evolution during triaxial deformation of crystalline rocks initially populated by
 489 microcracks is different. For initially microcracked samples, the permeability post-failure is not significantly
 490 higher than the pre-failure permeability. In their study, Chen et al. (2017) conducted triaxial deformation tests on
 491 Beishan granite samples that were heated to temperatures ranging from 100 to 800 °C prior to deformation. Upon
 492 initial loading, the permeability of the samples decreased, and the permeability reduction was more pronounced in
 493 samples exposed to higher temperatures (Chen et al. 2017). This may be explained by the fact that the density of
 494 thermally induced microcracks increases with the maximum temperature of exposure (Homand-Etienne &
 495 Houpert 1989) and, therefore, that the reduction in permeability caused by microcrack closure during the initial
 496 loading phase was greater in Beishan granite samples subjected to higher temperatures. As a result of brittle
 497 faulting, the permeability of Beishan granite samples thermally-stressed to 100 °C increased by an order of
 498 magnitude, whereas the permeability of samples thermally-stressed to 500 °C increased by a factor of two.
 499 Inversely, for samples heated to even higher temperatures (600 to 800 °C), the permeability after failure was
 500 approximately half of that before failure (Chen et al. 2017). The latter results are similar to those we report here for
 501 Lanhélin granite thermally-stressed to 700 °C, which show a permeability decrease following deformation in the
 502 brittle regime (Fig. 7). The permeability increase (to a residual value) observed in our samples upon the formation
 503 of shear fractures is likely limited by the fact that the fractures formed cut the samples at an angle (30° to the
 504 maximum principal stress) and do not connect the top and bottom end of the samples (where the pore fluid flows in
 505 and out of the samples). Although the fracture created acts as a preferential conduit for the pore fluid within a
 506 limited portion of the overall sample volume, the pore fluid is also required to travel through the remaining
 507 undeformed sections of the sample (result of strain localisation, Fig. 6). Consequently, the bulk permeability of the
 508 samples, which is governed by the permeability of the different layers within the anisotropic deformed rock
 509 (Farquharson and Wadsworth, 2018), did not exhibit a significant increase following the development of the shear
 510 fractures.

511

512 Our results show the evolution of permeability under compressive loading, including the interval from the onset of
 513 dilatancy (when porosity starts to increase, i.e., when the porosity change increases, Figs. 6 and 7) to the peak
 514 stress at failure. Typically, this phase is associated with the growth, propagation, and coalescence of microcracks
 515 that eventually lead to failure along a cross-cutting shear band (Paterson and Wong 2005). To gain further insight
 516 into the dependency of the permeability evolution to the initial microstructural attributes during this interval, we
 517 plot the evolution of permeability prior to the formation of a macroscopic shear fracture, termed the “permeability
 518 reduction factor” ε (Zhu and Wong 1997; Wong and Zhu 1999), as a function of initial porosity and permeability,
 519 where:

520

$$\varepsilon = \frac{k_{peak}}{k_C} - 1 \quad (1)$$

521

522 k_{peak} and $k_{C'}$ in Equation (1) are the permeabilities of the sample at the peak stress and at the stress for the onset of
523 inelastic deformation C' , respectively. Our data for microcracked Lanhélin are plotted alongside compiled data
524 from the literature (for granites, gabbros, sandstones, limestones, and Volvic trachyandesite) in Fig. 8. A value of
525 ε greater than zero indicates that the permeability of the rock increased as it approached macroscopic failure, and a
526 value of ε less than zero indicates that the permeability decreased. Previous studies have defined a porosity
527 threshold of 0.15 for the switch from permeability-increasing to permeability-decreasing behaviour in the brittle
528 regime (Zhu and Wong 1997; Wong and Zhu 1999). Our new data show that a negative permeability reduction
529 factor is possible at a porosity below 0.15 when the porosity consists of a network of microcracks. Fig. 8b shows
530 the permeability reduction factor as a function of the permeability at the onset of inelastic deformation $k_{C'}$.
531 According to this representation, the switch from a permeability-increasing to permeability-decreasing behaviour
532 occurs at a permeability threshold of 10^{-18} m^2 . However, more experimental are now required to verify this
533 potential limit and draw firm conclusions on the evolution of permeability during triaxial deformation of initially
534 microcracked rocks. The permeability decreasing behaviour observed for granite samples with initial porosities
535 lower than 0.15 in this study could be explained by (1) the relatively high initial permeability of the microcracked
536 Lanhélin granite compared to other low-porosity crystalline rocks (Fig. 8b) and (2) the fact that the macroscopic
537 fractures formed under triaxial compression developed at an angle and do not connect both ends of the samples
538 (Fig. 7). In comparison to other low-porosity crystalline rocks, the void space of our thermally-stressed granite
539 samples consists of an extensive network of microcracks (Fig. 4) which serves as a relatively efficient pathway for
540 fluid flow despite its relatively small volume. Indeed, our microcracked granite samples exhibit initial
541 permeability values higher than 10^{-18} m^2 – the permeability threshold defined previously – whereas the
542 permeability reported for initially intact low-porosity crystalline rocks such as Inada granite (Oda et al. 2002) or
543 Creighton gabbro (Trimmer et al. 1980) are more than one order of magnitude lower. Despite the thermal
544 microcracks accounting for a small volume within the bulk volume of the sample, their efficiency as fluid
545 pathways results in the sample exhibiting a low porosity but a relatively high permeability. However, the measured
546 increase in permeability due to the development of a macroscopic fracture (Fig. 7) is constrained by the fact that
547 the fracture does not connect the top and bottom of the sample. Therefore, the measured permeability increase does
548 not apparently counterbalance and overcome the decrease in permeability caused by the closure of the pre-existing
549 microcrack network under progressive loading, which is known as one of the most effective processes for
550 permeability reduction (Keaney 1998a).

551

552 Overall, our results highlight the importance of initial porosity and permeability and of microcrack density on the
553 evolution of permeability during inelastic deformation prior to failure at low effective pressure (in the brittle
554 regime). During pre-failure inelastic deformation, axial permeability can be significantly reduced in low-porosity
555 crystalline rocks initially populated by thermal microcracks whereas a permeability increase is observed in
556 initially intact low-porosity crystalline rocks.

557 4.3 Implications for fluid flow and geothermal energy exploitation

558

559 In crustal fault systems, micro- and macrocracks that populate the region surrounding the fault core, known as the
560 “damage zone”, play a significant role in fluid flow dynamics (Chester and Logan 1986; Wibberley et al. 2008;
561 Faulkner et al. 2010; Mitchell and Faulkner 2012). Our study sheds light on the influence of these micro- and
562 macrocracks on the permeability of granite and therefore serves to study fluid flow in microcracked damage zones
563 typically found around permeable faults and fractures in crustal fault systems.

564

565 Fluid flow and convective heat transfer in fault zones are controlled by a combination of geological and physical
566 factors. The permeability ratio between the fault and the surrounding basement is a crucial variable governing the
567 dynamics of convection (López and Smith 1995), which directly affects the intensity, depth, and spatial extent of
568 temperature anomalies (Duwiquet et al. 2019, 2021; Guillou-Frotier et al. 2020). Various processes within fault
569 damage zones enhance permeability compared to the intact basement rock. Tectonic comminution, fluid pressure,
570 and basement alteration are expected to generate fracture networks at different scales, thereby promoting fluid
571 flow (Hagemann et al. 1992; Cassard et al. 1994; Bejan and Lorente 2006; Launay 2019). In our study, the
572 permeability measurements performed after thermal stressing and at different stages of deformation enable us to
573 quantify the potential impact of such processes. To further understand the implications of our findings, we
574 consider a dynamic numerical model with an idealised 3D fault/basement geometry (Fig. 9). The Comsol
575 Multiphysics software was used to model a synthetic vertical fault zone with a thickness of 400 m in a cubic
576 homogeneous basement with 5.5 km-long sides. Thermal insulation was imposed to the vertical faces of the
577 basement and fluid circulation was blocked. A temperature of 20°C and a pressure of 10^5 Pa were imposed to the
578 upper horizontal face. Finally, a heat flux of $100 \text{ mW}\cdot\text{m}^{-2}$ (representing the sum of the mantle heat flow and that
579 resulting from the decay of radioactive elements in the underlying crust) was imposed at the base of the model (see
580 Duwiquet et al. 2022 for more details on the simulations). The physical properties measured on the intact,
581 thermally-stressed and deformed Lanhélin granite samples were used to parameterise the simulations (Table 3).

582

583 Two simulations were conducted, employing comparable model parameters (Table 3) with the exception of the
584 permeability of the basement surrounding the permeable fault. Fig. 9a illustrates the result obtained when an intact
585 basement surrounding the fault is considered, characterised by a permeability of $kb = 10^{-20} \text{ m}^2$ (Table 1). In
586 contrast, Fig. 9b depicts the results for a damaged basement containing both micro- and macrocracks, exhibiting a
587 permeability of $kb = 5 \times 10^{-18} \text{ m}^2$ (Table 3). Without any external heat source, the initial temperature distribution used in the
588 simulation follows the average geothermal gradient of the European crust ($30 \text{ }^\circ\text{C}/\text{km}$) (Fink et al. 2022).

589

590 The characteristics of the simulated fluid flow within the fault depends on the fault system configuration: a
591 permeable fault surrounded by an intact basement or by a fractured basement. In both cases, where fault
592 permeability is 10^{-14} m^2 , the convective motion is observed, with downwelling and upwelling motions contributing
593 to fluid flow. The velocities of the downwelling fluid reach $18 \times 10^{-9} \text{ m/s}$, while at the top of the upwelling
594 motions, the velocities are $4 \times 10^{-9} \text{ m/s}$ in the case of a damage zone whereas the velocities reach 19×10^{-9} and $2 \times$
595 10^{-9} m/s , respectively, in the case of an intact basement. Driven by buoyancy forces, these convective flows

596

597 significantly impact the thermal distribution within the fault. The upwelling motion leads to the rise of the 150 °C
598 isotherm to shallower depths, such as 1.1 km in the middle of the fault in the case of a fractured basement, and 2.2
599 and 1.1 km on the left and right sides respectively, in the case of an intact basement. Conversely, the downward
600 fluid motion concentrates the 150 °C isotherm at depths of 5.0 km on the left side in the case of a damage zone, and
601 at depths of 3.9 and 4.7 km in the case of an intact basement. In summary, the existence of a fractured basement
602 populated with micro- and macrocracks surrounding the permeable fault enhances convection within the fault
603 system due to its higher permeability, in contrast to a fault system comprised solely of a permeable fault directly
604 surrounded by an intact basement. Since the rocks adjacent to large fault systems are likely cracked on a number of
605 scales, we recommend that future modelling incorporates values for micro- and macrofractured rocks.

606

607 These simplified model results show how crustal fault zones with microcracked granitic damage zones around the
608 permeable fault can represent economically viable targets for high-temperature geothermal energy in amagmatic
609 systems. However, it is important to acknowledge that fault zones are subjected to various processes, such as
610 mineralisation, which can reduce permeability and limit fluid flow. Future studies could consider permeability as a
611 dynamic parameter that evolves in time and space, adapted to fractured media, and investigate its variation in
612 relation to triaxial deformation (Fig. 7).

613

614 5. Conclusions

615

616 In this study, we conducted a comprehensive investigation of the petrophysical and mechanical properties of
617 Lanhélin granite following thermal stressing, and the permeability evolution of thermally microcracked Lanhélin
618 granite during triaxial deformation. Our findings provide insights into the behaviour of granite under various
619 conditions and have implications for understanding fluid flow dynamics in fault zones and the exploitation of
620 geothermal resources.

621

622 First, our in-situ measurements during heating and cooling allowed us to observe the onset and progression of
623 thermal microcracking, as indicated by acoustic emission activity and a decrease in P-wave velocity. By observing
624 and analysing their microstructure, we confirmed that thermal microcracks initiate and grow due to the differential
625 thermal expansion of mineral constituents, leading to increases in porosity and permeability and decreases in P-
626 wave velocity, uniaxial compressive strength, thermal diffusivity, and thermal conductivity. Our experiments
627 demonstrated that the porosity and permeability of thermally-stressed granite decrease with increasing hydrostatic
628 pressure, reflecting the closure of thermal microcracks, in agreement with previous studies on microcracked rocks.
629 Second, under triaxial compression at low effective pressure (in the brittle regime), the permeability of
630 microcracked Lanhélin granite decreases following deformation beyond the peak stress. In detail, permeability
631 decreases during loading up to the peak stress and then increases back to a constant residual value upon the
632 development of a shear fracture. Our data revealed that, when the porosity consists of a network of microcracks, a
633 negative permeability reduction factor can be observed at a porosity below 0.15. However, it should be noted that
634 the increase in axial permeability resulting from the growth of shear fractures in our deformed samples was limited
635 by the fact that the shear bands did not extend across the full length of the samples and connect their ends. This
636 limitation, combined with the closure of pre-existing microcracks under progressive loading, led to an overall

637 decrease in permeability. Our results underscore the importance of initial porosity, permeability, and microcrack
 638 density in controlling the permeability evolution during inelastic deformation. The behaviour differs for
 639 microcracked rocks compared to initially intact low-porosity crystalline rocks, where the latter often exhibit a
 640 significant increase in permeability following inelastic deformation and failure in the brittle regime.

641

642 Our findings have implications for fluid flow dynamics in crustal fault zones. The presence of micro- and
 643 macrocracks in the damage zone can significantly impact permeability, influencing convective heat transfer and
 644 temperature anomalies. Our numerical simulations, incorporating the measured petrophysical properties of
 645 Lanhélin granite, demonstrated the enhanced convective flow within a fault system when surrounded by a
 646 fractured basement compared to an intact basement. Therefore, considering the presence of small deformation
 647 features such as microcracks and larger shear fractures surrounding a crustal fault could be crucial for identifying
 648 and modelling economically viable targets for geothermal energy in amagmatic systems. Future work should
 649 therefore help gathering more experimental data to inform on the evolution of permeability in rocks found in the
 650 surrounding of crustal fault zones at high pressure and high temperature.

651

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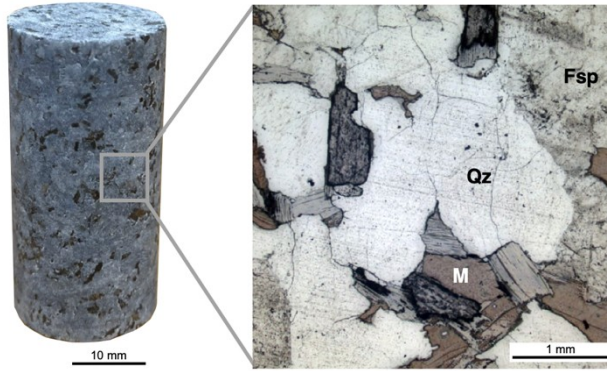
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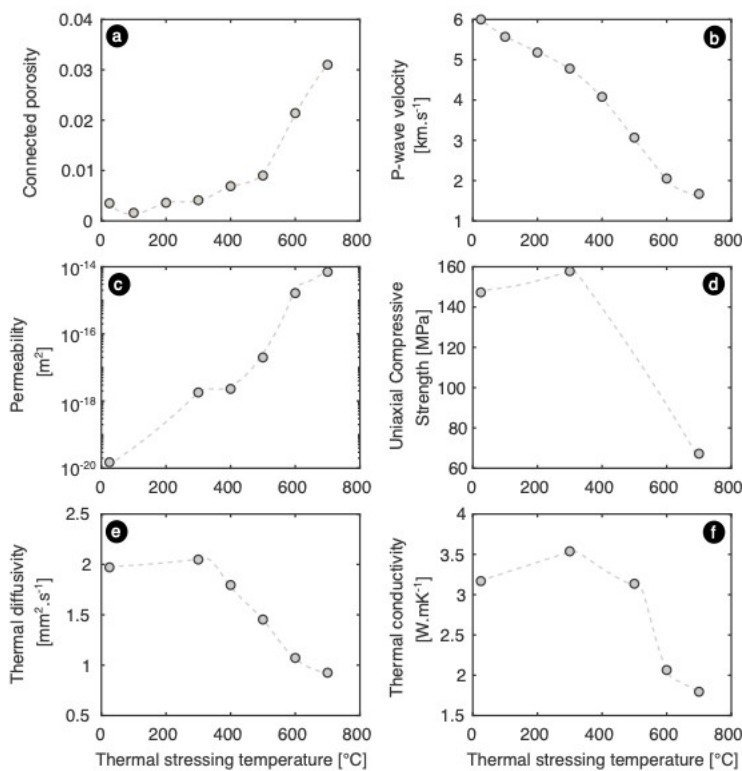
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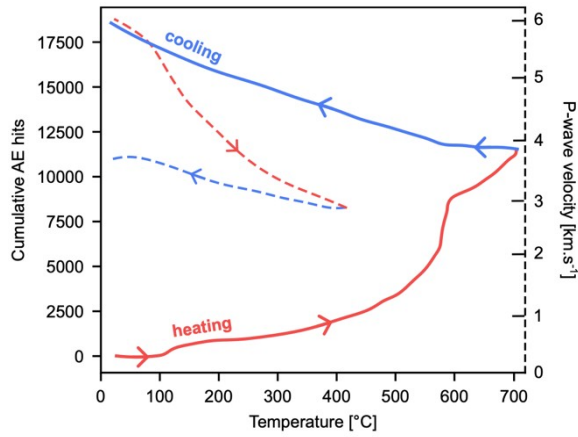
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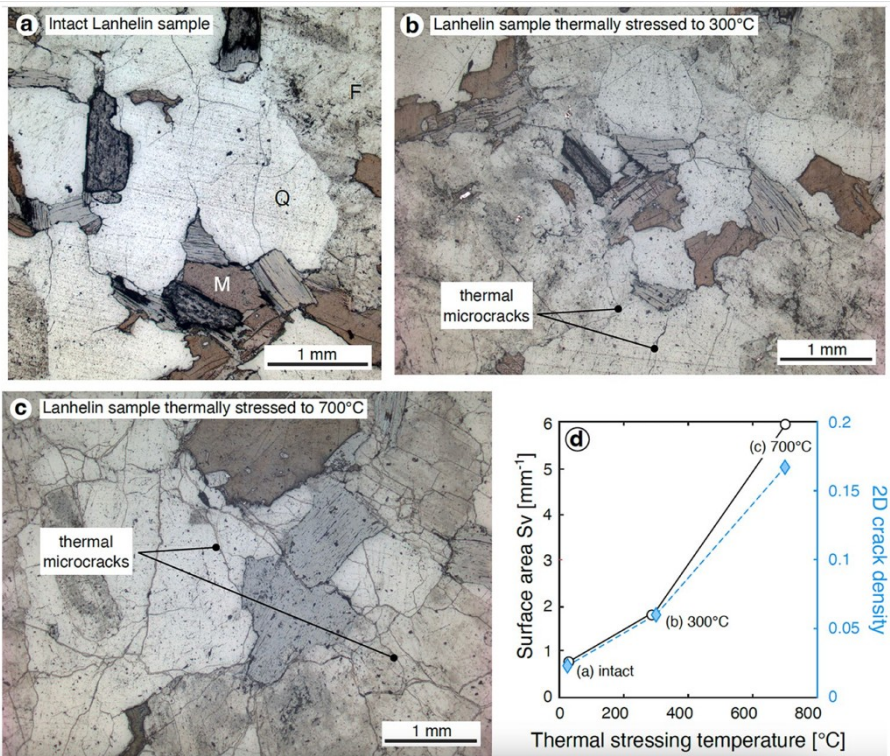
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 975 **Fig. 1** Photograph of a cylindrical (20 mm in diameter) Lanhélin granite sample and optical microscope image
 976 (taken in reflected light) of the intact microstructure, where quartz (Qz), feldspar (Fsp), and mica (M) crystals can
 977 be observed.



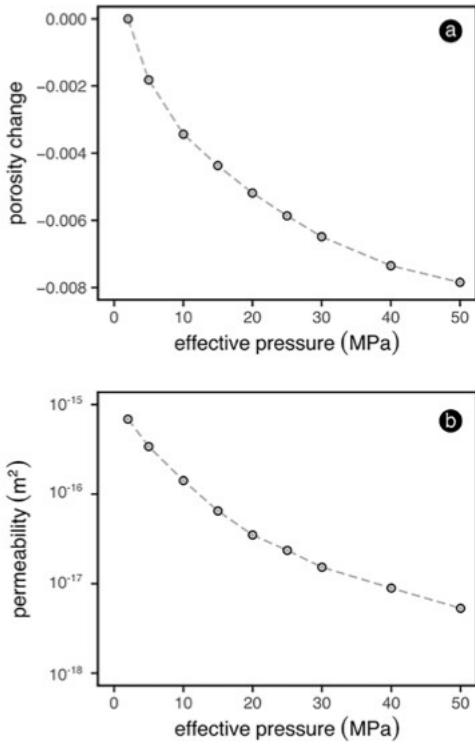
978
 979 **Fig. 2** Evolution of petrophysical properties following the thermal stressing of Lanhélin granite to 700 °C
 980 (measurements made at room temperature). (a) Connected porosity, (b) P-wave velocity, (c) permeability, (d)
 981 uniaxial compressive strength, (e) thermal diffusivity, and (f) thermal conductivity are given as a function of
 982 thermal-stressing temperature. Little variations are attributed to natural variability between the different granite
 983 samples used for the measurements at each temperature. Data are given in Table 1.



984 **Fig. 3** Acoustic emissions (AE) (solid lines) during a heating-cooling cycle from 20 to 700 °C and in-situ P-wave
 985 velocity (dashed lines) during a heating-cooling cycle from 20 to 450 °C for Lanhélin granite. Heating and cooling
 986 curves are shown in red and blue, respectively. AEs were recorded continuously using a 40 dB detection threshold.
 987 P-wave velocities were measured every 10 seconds during the heating-cooling cycle. While AE could be measured
 988 using sensors outside the furnace, the transducers for velocity measurements were placed inside the furnace, which
 989 ensured good quality in the data but limited the maximum temperature that could be used.
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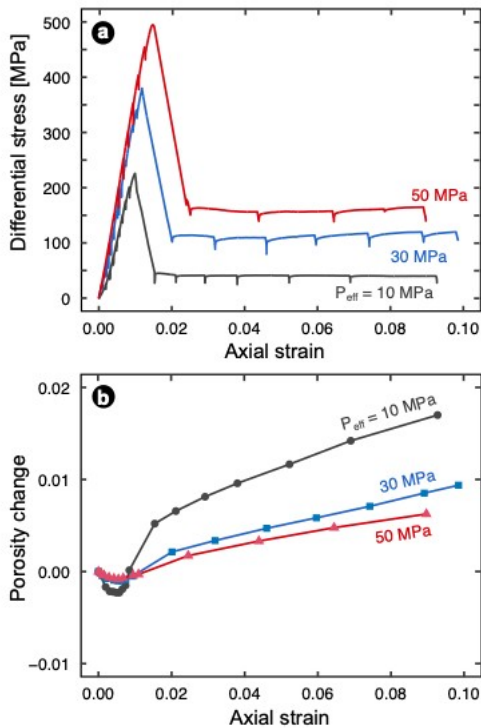


992 **Fig. 4** Microstructural evolution of Lanhélin granite following thermal stressing to 700 °C. An optical microscope
 993 image (taken in reflected light) of (a) the intact microstructure, (b) the microstructure following exposure to 300
 994 °C, and (c) the microstructure following exposure to 700 °C. These images were analysed using the crack-counting
 995 method presented in Griffiths et al. (2017) to determine (d) the evolution of surface area per unit volume (open
 996 circles) and 2D crack density (blue diamonds) as a function of thermal-stressing temperature.
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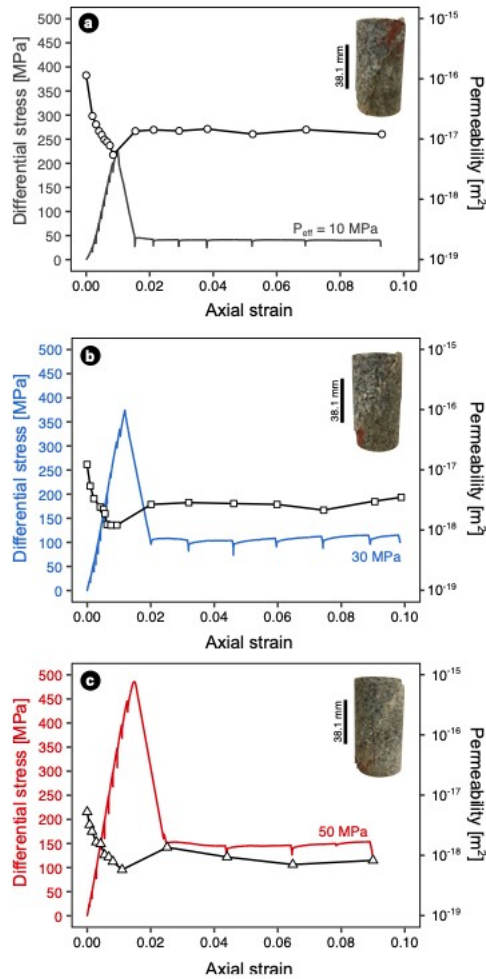
Fig. 5 Porosity change (a) and permeability (b) as a function of effective pressure during the hydrostatic pressurisation of thermally-stressed (to 700 °C) Lanhélin granite. Data are given in Table 2.



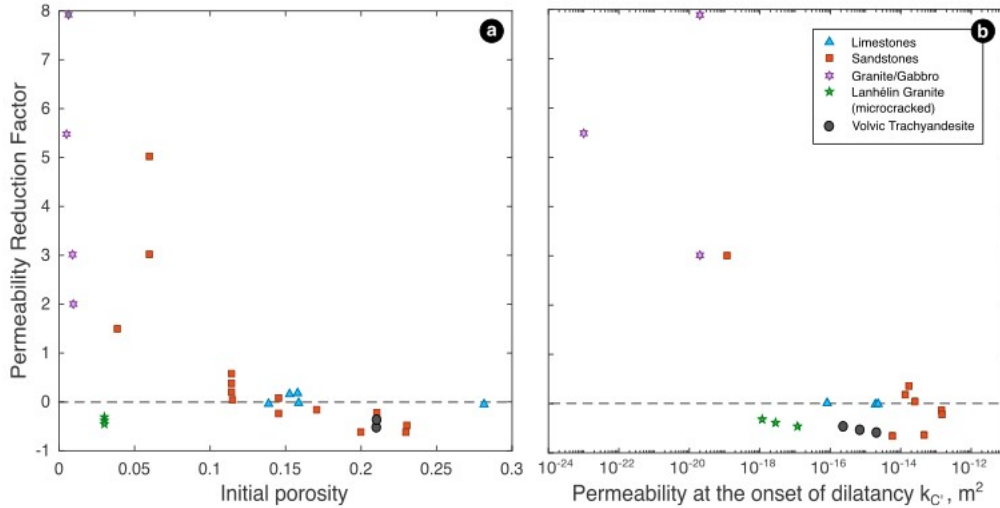
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Fig. 6 Mechanical data for the triaxial compression of thermally-stressed (to 700 °C) Lanhélin granite in the brittle regime. (a) Differential stress and (b) porosity change are presented as a function of axial strain for tests performed at effective pressures of 10 MPa (black lines), 30 MPa (blue lines), and 50 MPa (red lines). Several stages can be observed on the stress-strain curves. First, the elastic deformation of the sample. Then, the inelastic deformation: dilatancy takes place by opening and propagation of microcracks and ultimately, as the microcracks coalesce into a macroscopic fracture, brittle failure occurs as stress reaches its peak value. Lastly, mechanical data show a stage

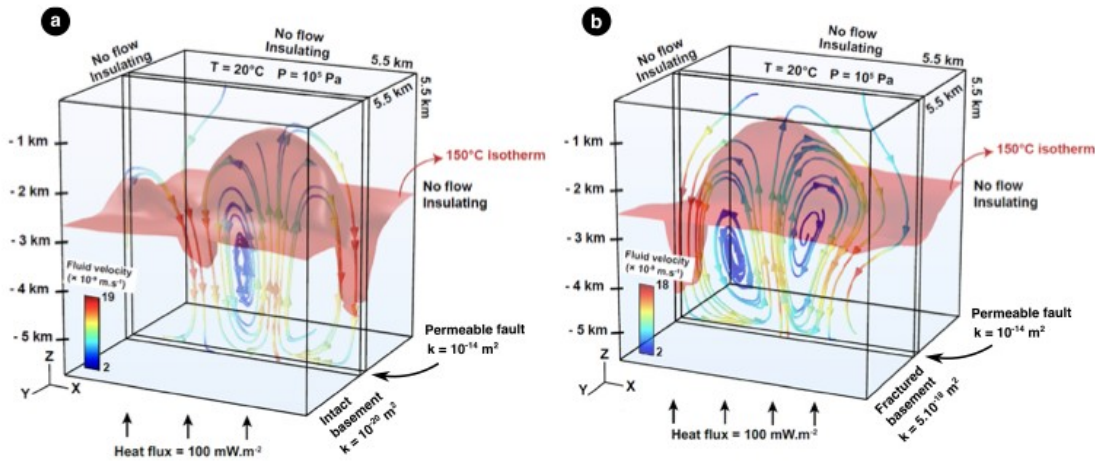
1010 during which stress remains more-or-less constant as deformation is accommodated by sliding on the formed
 1011 fracture. An Excel file containing all the mechanical data is provided in the Supplementary Information.
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1013
 1014 **Fig. 7** Evolution of the permeability of thermally-stressed (to 700 °C) Lanhélin granite during deformation in the
 1015 brittle regime. Mechanical data (solid lines) and permeability (dashed lines) are presented as a function of axial
 1016 strain for triaxial experiments performed at effective pressures of (a) 10 MPa, (b) 30 MPa, and (c) 50 MPa.
 1017 Photographs of the post-deformation samples are included in the top right corner of each panel. Larger version of
 1018 these photographs and an Excel file containing all the mechanical data and permeability data are provided in the
 1019 Supplementary Information.
 1020



1021 **Fig. 8** Permeability reduction factor (see Equation 1) as a function of (a) initial porosity and (b) permeability at the
 1022 onset of inelastic deformation for microcracked Lanhélin granite (green stars) and compiled data for limestones
 1023 (blue triangles; data from Meng et al. 2019), sandstones (red squares; data from (Zhu and Wong 1997; Keaney et
 1024 al. 1998; Wong and Zhu 1999; Baud et al. 2012), granites and gabbros (purple stars; data from (Zoback and
 1025 Byerlee 1975; Trimmer et al. 1980), and trachyandesite (black circles; data from Heap et al. (2022)).
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1028 **Fig. 9** Simulated fluid flow within a crustal fault zone system consisting of (a) a permeable fault surrounded by
 1029 intact basement rock and (b) a permeable fault surrounded by a fractured basement, populated with micro- and
 1030 macrocracks. The numerical simulation results represent the steady state condition, after 10 000 years. The 150 °C
 1031 isotherm is highlighted by the pink surface. Fluid flow is marked by streamlines, where the red colour corresponds
 1032 to the highest velocity and the blue colour to the lowest. A detailed description of the method used for the
 1033 numerical simulations is given in Duwiquet et al. (2022).
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Sample	Thermal stressing temperature (°C)	Connected porosity	P-wave velocity (km.s ⁻¹)	Permeability (m ²)	UCS (MPa)	Thermal diffusivity (mm ² .s ⁻¹)	Thermal conductivity (W.mK ⁻¹)
LG1	RT	0.0035	6.00	1.50×10^{-20}	147	1.97	3.17
LG2	100	0.0016	5.57	-	-	-	-
LG3	200	0.0036	5.18	-	-	-	-
LG4	300	0.0041	4.78	1.80×10^{-18}	158	2.05	3.54
LG5	400	0.0069	4.08	2.30×10^{-18}	-	1.80	2.52
LG6	500	0.0090	3.07	1.99×10^{-17}	-	1.45	3.14
LG7	600	0.0214	2.05	1.64×10^{-15}	-	1.07	2.06
LG8	700	0.0310	1.67	7.05×10^{-15}	67	0.92	1.79

1037

1038 **Table 1** Connected porosity, P-wave velocity, permeability, uniaxial compressive strength (UCS), thermal
1039 diffusivity, and thermal conductivity data for Lanhélin granite as a function of thermal-stressing temperature.

1040 Permeability, UCS and thermal properties were voluntarily measured only for selected thermal stressing
1041 temperatures. RT - room temperature. Mechanical data for the UCS tests are included in Supplementary
1042 Information.

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Effective pressure (MPa)	Porosity change	Permeability (m ²)
2	0	6.9×10^{-16}
5	-0.00182	3.4×10^{-16}
10	-0.00344	1.4×10^{-16}
15	-0.00437	6.5×10^{-17}
20	-0.00518	3.5×10^{-17}
25	-0.00587	2.4×10^{-17}
30	-0.00649	1.5×10^{-17}
40	-0.00735	8.9×10^{-18}
50	-0.00785	5.3×10^{-18}

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1046 **Table 2** Porosity change and permeability data as a function of effective pressure during the hydrostatic
 1047 pressurisation of thermally-stressed (to 700 °C) Lanhélin granite. The pore pressure used in the experiment was 10
 1048 MPa.

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Parameters	Symbols	Values	Units
Density	ρ_s	2700	kg/m ³
Fault porosity	φ_f	0.10	-
Basement porosity	φ_b	0.05	-
Fault permeability	k_f	10^{-14}	m ²
Intact basement permeability	k_b	10^{-20}	m ²
Fractured basement permeability	k_d	5×10^{-18}	m ²
Thermal conductivity	λ_s	3.54	W/(m.K)
Thermal capacity	C_{pa}	1000	K/(kg.K)
Thermal expansion coefficient	β	7×10^{-6}	1 /K
Fluid variables and parameters			
Fluid density	ρ_f	$1036.5 - 0.14167 \times (T-273.15) - 0.0022381 \times (T - 273.15) \times (T - 273.15)$	kg/m ³
Thermal conductivity	λ_f	0.60	W/(m.K)
Thermal capacity	C_{pf}	4100	J/(kg.K)
Thermal expansion coefficient	α	10^{-4}	1/K
Fluid dynamic viscosity	μ	$2.414 \times 10^{-5} \exp(570/((T-273.15)+133))$	Pa.s

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1053 **Table 3** Rock and fluid parameters used in the numerical simulations, derived from measurements made on intact,

1054 thermally-stressed, and deformed Lanhélin granite in this study.

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