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1 Petrophysical properties, mechanical behaviour, and failure  
2 modes of impact melt-bearing breccia (suevite) from the Ries  
3 impact crater (Germany)

4

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18

## 19 **Abstract**

20 The physical properties and mechanical behaviour of impactites are an  
21 important parameter in fluid flow models and slope stability and landscape evolution  
22 assessments for heavily impacted planetary bodies. We first present porosity,  
23 permeability, Young's modulus, and uniaxial compressive strength measurements for  
24 three suevites from the Ries impact crater (Germany). Porosity ranges from 0.18 to

25 0.43, permeability from  $5.8 \times 10^{-16}$  to  $5.1 \times 10^{-14}$  m<sup>2</sup>, Young's modulus from 1.4 to 8.1  
26 GPa, and uniaxial compressive strength from 7.3 to 48.6 MPa. To explore their  
27 mechanical behaviour, we performed triaxial deformation experiments on these  
28 samples at a range of confining pressures. The brittle–ductile transition for the lowest  
29 (0.25) and highest (0.38) porosity suevite samples was at a confining pressure of ~30  
30 and ~10 MPa, respectively (corresponding to, for example, depths respectively of ~1  
31 and ~4 km on Mars). Microstructural observations show that the dominant deformation  
32 micromechanism during brittle deformation is microcracking, and during ductile  
33 deformation is cataclastic pore collapse. We show that a theoretically grounded  
34 permeability model for welded granular media accurately captures the permeability of  
35 the studied suevites, and we use micromechanical models to glean insight as to their  
36 mechanical behaviour. Finally, we upscale our laboratory measurements to provide  
37 physical property values for lengthscales more relevant for large-scale models, and we  
38 compare these data with those for basalt (a lithology representative of the surface of the  
39 inner Solar System bodies). These analyses show how macroscopic fractures serve to  
40 increase the permeability and decrease the strength and Young's modulus of suevite;  
41 for example, we find that basalt can be a factor of 2–5 stronger than suevite in the  
42 shallow crust. Our study suggests, therefore, that the rock masses comprising older,  
43 bombarded crusts are substantially weaker and more porous and permeable than the  
44 younger plains material on these bodies. These findings should be considered in large-  
45 scale fluid flow modelling and when providing crustal strength estimates or slope  
46 stability assessments for planetary bodies on which protracted records of impact  
47 bombardment are preserved.

48

49 **Keywords:** suevite; porosity; strength; permeability; alteration; Young's modulus

## 50 **1 Introduction**

51           Impact craters are a geological feature ubiquitous to the solid-surface planetary  
52 bodies of our Solar System (e.g., Neukum et al., 1975; Melosh, 1989; Phillips et al.,  
53 1992; Wood et al., 2010; Barnouin et al., 2012; Robbins and Hynek, 2012; Robbins et  
54 al., 2017). The preservation of impact craters on Earth is rare, largely due to erosion  
55 and tectonic activity (e.g., Grieve and Shoemaker, 1994). However, impact  
56 bombardment of bodies in the inner Solar System is all but certain; for example, nearly  
57 400,000 >1 km-diameter craters have been identified on Mars, a few hundred of which  
58 are >100 km in diameter (Robbins and Hynek, 2012), and Mercury is similarly marked  
59 (e.g., Fassett et al., 2012a). Well-preserved craters on the terrestrial planets and the  
60 Moon are associated with impactites, rocks that have either been created or modified  
61 by an impact (e.g., Osinski et al., 2011; Stöffler et al., 2018).

62           On Earth, where we can readily access and assess such deposits (even if they  
63 are comparatively rare), proximal impactites have been classified into three groups:  
64 shocked target rocks, impact breccias, and impact melt rocks (e.g., Stöffler and Grieve,  
65 2007; Osinski and Pierazzo, 2013; Stöffler et al., 2018). Impact breccias are  
66 characterised by their clast contents, the most distinctive of which are impact melt-  
67 bearing breccias (“suevites”). Impactites as observed on Earth are also often variably  
68 altered (typically argillic alteration and zeolitisation) by the hydrothermal system that  
69 forms following the impact (e.g., Osinski et al., 2013; Kirsimäe and Osinski, 2013) or  
70 by supergene alteration (Muttik et al., 2010, 2011).

71           The petrophysical properties of impactites, such as porosity, permeability,  
72 Young’s modulus, and strength, are valuable input parameters for a wide range of  
73 modelling. For example, the porosity and permeability of impactites are required in  
74 models designed to understand fluid flow within the hydrothermal systems associated

75 with impact craters (e.g., Rathbun and Squyres, 2002; Abramov and Kring, 2004, 2005;  
76 Sanford, 2005; Versh et al., 2006; Abramov and Kring, 2007; Barnhart et al., 2010).  
77 Further, for heavily impacted crusts such as that of Mars and Mercury, the porosity and  
78 permeability of impactites may provide useful constraints for estimates of crustal  
79 porosity and permeability. Porosity and permeability estimates of the Martian crust, for  
80 example, are required for the modelling of outflow channel formation (e.g., Gulick,  
81 1998; Manga, 2004; Hanna and Phillips, 2005; Andrews-Hanna and Phillips, 2007;  
82 Musiol et al., 2011), crustal hydrothermal convection (e.g., Travis et al., 2003),  
83 hydrothermal crustal cooling (e.g., Parmentier and Zuber, 2007), methane release into  
84 the atmosphere (e.g., Lyons et al., 2005; Etiope et al., 2013), and groundwater flow  
85 (e.g., Grimm and Painter, 2009; Harrison and Grimm, 2009). Additionally, knowledge  
86 of the mechanical behaviour and strength of impactites is important for our  
87 understanding of impact crater collapse (e.g., Melosh and Ivanov, 1999; Wünnemann  
88 and Ivanov, 2003), the modelling of impact crater geometry (e.g., Collins, 2004; Potter  
89 et al., 2012), and values of uniaxial compressive strength of crustal rocks are needed  
90 for characterising the strength and stability of rock masses using tools such as rock  
91 mass rating (RMR) or the geological strength index (GSI) (e.g., Schultz, 1993, 2002;  
92 Neuffer and Schultz, 2006; Nahm and Schultz, 2007; Okubo et al., 2011; Brunetti et  
93 al., 2014; Klimczak, 2015; Klimczak et al., 2015). Further, strength values for  
94 impactites could help inform studies that provide rock strength estimates from rover  
95 drilling and abrasion data (e.g., Thomson et al., 2013; Peters et al., 2018), as well as  
96 improving physical weathering rate estimates (e.g., Viles et al., 2010; Eppes et al.,  
97 2015).

98         Despite the need for robust input parameters for a variety of modelling studies,  
99 there are, to our knowledge, only two studies that provide laboratory-measured values

100 for the permeability of impact melt-bearing breccia (Parnell et al., 2010; Heap et al.,  
101 2020a), and but one that provides values of uniaxial compressive strength (Heap et al.,  
102 2020a). Other studies have, however, provided other physical property values for  
103 impactites. For example, values of density, porosity, thermal conductivity, elastic wave  
104 velocities, and magnetic susceptibility are available for impactites from the Chicxulub  
105 impact crater (Mexico) (e.g., Mayr et al., 2008a; Elbra and Pesonen, 2011; Christeson  
106 et al., 2018). Parnell et al. (2010) found that the permeability of impact breccias from  
107 the Haughton (Canada), Ries (Germany), and Rochechouart (France) impact craters to  
108 be between  $\sim 1 \times 10^{-16}$  and  $\sim 1 \times 10^{-15}$  m<sup>2</sup>. Heap et al. (2020a) measured the porosity,  
109 permeability, and uniaxial compressive strength of samples of impact melt-bearing  
110 breccia collected from two locations at the Ries impact crater. They found average  
111 connected porosities of 0.24 and 0.25, and average uniaxial compressive strengths of  
112 34.2 and 29.2 MPa, for samples taken from the quarries at Seelbronn and Altenbürg,  
113 respectively. The average permeability of these impact melt-bearing breccias was found  
114 to be  $\sim 2 \times 10^{-15}$  m<sup>2</sup> (Heap et al., 2020a). However, notwithstanding these few studies,  
115 petrophysical measurements are currently unavailable for altered impactites and thus a  
116 full assessment of the mechanical behaviour, failure modes (brittle or ductile), and  
117 deformation micromechanisms of impactites is absent. The paucity of studies that  
118 report petrophysical and mechanical properties of impactites reflects, in part, the poor  
119 preservation potential of impact craters and their impactites on Earth (Grieve and  
120 Shoemaker, 1994). As a result, the Ries impact crater (Figure 1), one of only several  
121 impact craters on Earth with well-preserved and easily accessible impact melt-bearing  
122 breccias (Osinski et al., 2011), serves as an ideal location to sample and study  
123 impactites.

124           The Ries crater (Figure 1) is a double-layer rampart crater with an inner- and  
125 outer-ring diameter of about 12 and 24 km, respectively (e.g., Wünnemann et al., 2005;  
126 Figure 1), that shares similarities with Martian craters (e.g., Sturm et al., 2013). The  
127 crater formed following an impact into a layer of sedimentary rocks 600–700 m thick  
128 overlying a crystalline basement (e.g., Pohl et al., 1977; Stöffler et al., 2013) about 15  
129 million years ago (e.g., Rocholl et al., 2018). Types of impactite identified at Ries crater  
130 include (von Engelhardt, 1990): shocked target rocks; breccia derived mainly from the  
131 sedimentary cover (the “Bunte breccia”, e.g., Hörz et al., 1983); polymict crystalline  
132 breccia; impact melt-bearing breccias (“suevite”, e.g., von Engelhardt et al., 1995,  
133 1997); and impact melt rocks (e.g., Osinski, 2004; Pohl et al., 1977). We focus here on  
134 providing rock physical and mechanical properties (porosity, permeability, Young’s  
135 modulus, compressive strength, and failure mode) for samples of suevite, a type of  
136 impact melt-bearing breccia found at the Ries impact crater. There are three types of  
137 suevite at Ries crater: (1) “crater suevite”, which is up to 300 m thick and covered by  
138 Miocene lake sediments; (2) “outer suevite”, which is only  $\leq 25$  m thick and outcrops  
139 up to 22 km from the centre of the structure; and (3) dyke suevite, found in the  
140 parautochthonous crater basement and in ejected crystalline megablocks (e.g., Stöffler  
141 et al., 2013). These poorly sorted impact melt-bearing breccias formed as a result of the  
142 welding of glass/melt and lithic fragments deposited by fluidised granular density  
143 currents (e.g., Siegert et al., 2017; see also von Engelhardt, 1997; Newsom et al., 1990;  
144 Stöffler et al., 2013) generated following the impact. Suevites at Ries contain  
145 centimetre-sized fragments of sedimentary and crystalline rocks (e.g., Graup, 1981; von  
146 Engelhardt, 1990, 1997) and angular fragments and aerodynamically shaped bombs of  
147 impact melt glass (Stähle, 1972; Osinski, 2003). These clasts are hosted within a matrix  
148 of impact melt glass, rock and mineral fragments, crystallites, and clay minerals (and

149 sometimes zeolites) that formed as a result of post-deposition hydrothermal alteration  
150 and weathering (Newsom et al., 1986; von Engelhardt, 1990, 1997; Osinski, 2004,  
151 2005; Muttik et al., 2008, 2011; Sapers et al., 2017). High-pressure minerals, such as  
152 coesite (Stöffler, 1966; Stähle et al., 2008) and diamond (Schmitt et al. 2005; Stöffler  
153 et al., 2013), have also been identified. Coesite, a high-pressure silica phase, is the  
154 mineral that confirmed the impact origin of these deposits in the 1960s (Shoemaker and  
155 Chao, 1961).

156         Here, we present a laboratory study designed to first report the petrophysical  
157 properties (porosity, permeability, Young's modulus, and uniaxial compressive  
158 strength) of variably altered suevites (impact melt-bearing breccia) from Reis crater  
159 (building on data recently published by Heap et al., 2020a). We then present the results  
160 of a suite of triaxial deformation experiments designed to better understand and contrast  
161 the mechanical behaviour of relatively unaltered and altered suevites. We complement  
162 these mechanical data with microstructural work to determine the operative  
163 deformation micromechanisms. We then provide permeability and micromechanical  
164 modelling to better understand the fluid flow and mechanical properties of these  
165 samples. Since upscaled values are of most importance for large-scale models, we use  
166 techniques often employed in geotechnical engineering to provide values of strength,  
167 Young's modulus, and permeability relevant for longer lengthscales. Finally, we  
168 outline the implications of these data for planetary bodies that record protracted periods  
169 of impact bombardment.

170

## 171 **2 Experimental materials and methods**

172         Blocks of suevite (impact melt-bearing breccia) were collected from three  
173 different quarries located within or adjacent to the Ries impact crater in Nördlingen:



174 the Seelbronn quarry (one block: SUE), the Altenbürg quarry (one block: ALT), and  
175 the Aumühle quarry (three blocks: AU1, AU2, and AU3) (Figure 1). These blocks were  
176 selected due to their visible differences in alteration (colour). The blocks from all three  
177 quarries are poorly sorted clastic rocks that contain millimetre- and centimetre-sized  
178 clasts of glass (aerodynamically shaped bombs and angular fragments), crystalline  
179 rocks (Variscan granites and pre-Variscan gneisses and amphibolites), and sedimentary  
180 rocks within a fine-grained matrix (as described in von Engelhardt, 1990, 1997; Stöfler  
181 et al., 2013); photographs of representative 25 mm-diameter core samples are shown in  
182 Figure 2.

183 The mineral content of the five blocks was quantified using X-ray powder  
184 diffraction (XRPD). Powdered aliquots of each block, containing an internal standard  
185 of 20 wt.% ZnO, were ground for 8 min with 10 ml of isopropyl alcohol in a McCrone  
186 micronising mill using zirconium dioxide cylinder elements. The XRPD analyses were  
187 performed on powder mounts using a Bruker D8 advance eco X-ray diffractometer  
188 (CuK $\alpha$ , automatic divergence slit, 15 mm irradiated sample length, anti-air scatter  
189 screen, 2.5° primary and secondary soller slits, step-scan from 2 to 75° 2 $\Theta$  with 0.01°  
190 increments per second, counting time one second per increment, 40 kV, 25 mA). The  
191 phases in the whole rock powders were quantified using the Rietveld program BGMN  
192 (Bergmann et al., 1998) and Profex 4.0.2 graphical user interface (Döbelin and  
193 Kleeberg, 2015). To identify the clay minerals, <2  $\mu$ m fractions were also separated by  
194 gravitational settling. Oriented mounts containing the <2  $\mu$ m fraction were X-rayed in  
195 an air-dried and ethylene-glycolated state.

196 Thin sections were prepared from offcuts taken from blocks from each of the  
197 three quarries (SUE, ALT, and AU1) and their microstructure was examined using a  
198 Tescan Vega 2 XMU system backscattered scanning electron microscope (SEM).

199 To assess the pore-throat size distribution of our blocks, we performed mercury  
200 injection porosimetry on offcuts (between 3 and 5 g) of blocks SUE, ALT, and AU1  
201 using a Micromeritics Autopore IV 9500. The mercury equilibration time and filling  
202 pressure were 10 s and ~3860 Pa, respectively, and the evacuation time and evacuation  
203 pressure were 5 min and 50  $\mu$ mHg, respectively. The pressure range was ~690 Pa to  
204 ~414 MPa.

205 Cylindrical samples were cored from each block in the same orientation with  
206 diameters of 20 or 25 mm (Figure 2). These samples were then precision-ground to a  
207 nominal length of 40 or 60 mm, respectively. The samples were cored so as to avoid  
208 centimetre-sized clasts. The prepared cylindrical samples were then washed with water  
209 to remove any water-soluble grinding fluid and vacuum dried at 40 °C in an oven for  
210 at least 48 hours.

211 The connected porosity of each cylindrical sample was then determined using  
212 the skeletal (connected) volume of the sample measured by a helium pycnometer  
213 (Micromeritics AccuPyc II 1340), and the bulk volume of the sample calculated using  
214 the sample dimensions.

215 Permeabilities of the 25 mm-diameter cylindrical samples were measured with  
216 a nitrogen gas permeameter (Farquharson et al., 2016; Heap and Kennedy, 2016) under  
217 a confining pressure of 1 MPa. Permeability was measured using the steady-state  
218 method. Steady-state volumetric flow rates were determined (using a gas flowmeter)  
219 for several differential pressures (acquired with a pressure transducer). These data were  
220 used to calculate permeability using Darcy's law and to check whether the Forchheimer  
221 (Forchheimer, 1901) or Klinkenburg (Klinkenburg, 1941) corrections were required.  
222 These corrections were applied when necessary on a case-by-case basis.

223           The 25 mm-diameter samples were deformed uniaxially using a uniaxial load  
224 frame. Dry samples were deformed at an axial strain rate of  $10^{-5} \text{ s}^{-1}$  until macroscopic  
225 failure (that is, the formation of a macroscopic fracture). A lubricating wax was placed  
226 on the end-faces of the samples to avoid problems with friction between the sample and  
227 the pistons. During deformation, axial displacement and axial load were measured with  
228 a linear variable differential transducer (LVDT) and a load cell, respectively. The  
229 deformation of the loading column itself (including endcaps and piston) was removed  
230 from the axial displacement signal. Axial displacement and axial load were converted  
231 to axial strain and axial stress using the sample dimensions. The static Young's modulus  
232 was calculated using the slope of the strain–stress curve within the linear elastic region  
233 (e.g., Heap et al., 2020b).

234           Triaxial deformation experiments were performed on 20 mm-diameter samples  
235 (a detailed schematic of the device is provided in Farquharson et al. (2017)). Dry  
236 samples were deformed at a constant strain rate of  $10^{-5} \text{ s}^{-1}$  until macroscopic failure (for  
237 the brittle experiments) or until an axial strain of 6% was reached (for the ductile  
238 experiments). A strain of 6% was chosen as a cut-off because this value is large enough  
239 for observing the ductile behaviour and for recovering a sample for analysis, but not  
240 too large to require unreasonable experimental times. (The choice of this cut-off strain  
241 amount does not impact on our interpretation or conclusions.)

242           Experiments were performed at a range of constant confining pressures, from  
243 2.5 to 80 MPa. Within a given experiment, the confining pressure was held constant  
244 using a confining pressure intensifier. Axial displacement and axial load were measured  
245 during deformation using an LVDT and a load cell, respectively. Axial displacement  
246 (following removal of the deformation of the loading column) and axial load were  
247 converted to axial strain and axial stress, respectively, using the sample dimensions.

248 Additional triaxial experiments were performed in which the differential stress was held  
249 at a constant 2 MPa and the confining pressure was increased in steps of 3–15 MPa.  
250 The confining pressure was only increased once the axial strain had stabilised, i.e.,  
251 when the sample had equilibrated to the new confining pressure (with typical durations  
252 on the order of 30 minutes to an hour). The confining pressure was increased in this  
253 stepwise manner until the onset of inelastic behaviour, signalled by a large increase in  
254 axial strain. We assume a simple effective pressure law where the effective pressure  
255 ( $P_{\text{eff}}$ ) is the confining pressure minus the pore fluid pressure (zero for the experiments  
256 performed herein). We consider compressive stresses and strains as positive.

257

## 258 **3 Results**

259

### 260 3.1 Mineral content and microstructure

261 The mineral content for all the blocks collected is provided in Table 1. Blocks  
262 SUE and ALT are primarily composed of amorphous phases, plagioclase, smectite,  
263 calcite, and quartz, with minor quantities of coesite, K-feldspar, biotite, and Fe oxides,  
264 such as hematite or goethite (Table 1). The main difference between blocks SUE and  
265 ALT is that SUE contains more plagioclase and ALT contains more calcite (Table 1).  
266 The blocks from the Aumühle quarry (AU1, AU2, and AU3), which are all similar in  
267 composition, contain much fewer amorphous phases and much more smectite than the  
268 blocks from Seelbronn (SUE) and Altenbürg (ALT) (Table 1).

269 Representative backscattered SEM images of the SUE, ALT, and AU1 blocks are  
270 shown in Figure 3. All three blocks contain poorly sorted angular fragments (quartz,  
271 calcite, K-feldspar, plagioclase) within a fine-grained matrix (Figure 3). The SEM  
272 images clearly show that the block from the Aumühle quarry (AU1; Figure 3c) is much

273 more porous (black pixels) than those from the Seelbronn (SUE; Figure 3a) and  
274 Altenbürg (ALT; Figure 3b) quarries. We also note that blocks SUE and ALT are  
275 pervasively microcracked (Figures 3a and 3b, respectively).

276 The pore-throat size distributions for the SUE, ALT, and AU1 blocks are  
277 provided as Figure 4. These data show that AU contains a greater volume of large pore  
278 throats than SUE and ALT. For example, ~50% of the volume of pore throats within  
279 AU are greater than 0.1  $\mu\text{m}$ , whereas pore throats greater than 0.1  $\mu\text{m}$  only comprise 30  
280 and 35% of the volume of pore throats in ALT and SUE, respectively (Figure 4). The  
281 ALT sample contains a large volume of small pore throats: ~60% of the volume of pore  
282 throats within ALT is less than 0.01  $\mu\text{m}$ , whereas pore throats less than 0.01  $\mu\text{m}$  only  
283 comprise 40 and 15% of the volume of pore throats in SUE and AU, respectively.

284

### 285 3.2 Porosity and permeability

286 The average connected porosities of the blocks from the quarries at Seelbronn  
287 (SUE), Altenbürg (ALT), and Aumühle (AU) are 0.238, 0.237, and 0.390, respectively  
288 (Table 2). Their average permeabilities are  $1.95 \times 10^{-15}$ ,  $2.02 \times 10^{-15}$ , and  $8.20 \times 10^{-15}$   
289  $\text{m}^2$ , respectively (Table 2). Our data show that permeability increases as a function of  
290 increasing porosity. The samples from Aumühle (AU) are more porous and more  
291 permeable than those from Seelbronn (SUE) and Altenbürg (ALT) (Figure 5; Table 2)

292

### 293 3.3 Uniaxial compressive strength and static Young's modulus

294 Representative axial stress–strain curves for samples from each of the three  
295 quarries are provided in Figure 6a. First, axial stress is a non-linearly increasing  
296 function of axial strain, typically attributed to the closure of pre-existing microcracks  
297 (section “OA” labelled on the SUE curve in Figure 6a). This stage is followed by a

298 quasi-linear elastic stage, from which the static Young's modulus was determined  
299 (section "AB"; Figure 6a). Following the elastic deformation stage, the axial stress is a  
300 non-linearly decreasing function of axial strain, typically interpreted as due to the  
301 nucleation and growth of microcracks (section "BC"; Figure 6a). The peak stress (the  
302 uniaxial compressive strength of the sample) is followed by strain softening and a stress  
303 drop associated with the macroscopic failure of the sample (Figure 6a). The average  
304 uniaxial compressive strengths of the blocks from the quarries at Seelbronn (SUE),  
305 Altenbürg (ALT), and Aumühle (AU) are 34.2, 29.2, and 12.9 MPa, respectively (Table  
306 2). Our data show that uniaxial compressive strength decreases as a function of  
307 increasing porosity. The samples from Aumühle (AU) are weaker than those from  
308 Seelbronn (SUE) and Altenbürg (ALT) (Figure 6; Table 2).

309         Static Young's modulus as a function of connected porosity is shown in Figure  
310 7. The average Young's moduli are 6.5, 5.7, and 3.0 GPa for the blocks from the  
311 quarries at Seelbronn (SUE), Altenbürg (ALT), and Aumühle (AU), respectively (Table  
312 2). We find that Young's modulus decreases as a function of increasing porosity. The  
313 samples from Aumühle (AU) have a lower Young's modulus than those from Seelbronn  
314 (SUE) and Altenbürg (ALT) (Figure 7; Table 2)

315

#### 316 3.4 Triaxial deformation experiments

317         The stress–strain curves for the triaxial deformation experiments are shown in  
318 Figures 8a (for SUE) and Figure 8b (for AU1) (a mechanical summary of the triaxial  
319 experiments performed for this study is provided in Table 3). The suevites from  
320 Seelbronn and Aumühle were selected for triaxial testing because they represent the  
321 endmembers in terms of alteration, porosity, and strength (Table 2). For sample SUE,  
322 the experiments performed at  $P_{\text{eff}}$  of 0, 5, 10, and 15 MPa clearly show the post-peak

323 stress drop associated with brittle behaviour (i.e., the formation of a shear fracture; see  
324 Wong and Baud (2012) for a review on brittle and ductile behaviour of porous rocks).  
325 The differential stress at failure also increases as the effective pressure increases from  
326 0 to 15 MPa (Figure 8a; Table 3). Although the stress–strain curve for the sample  
327 deformed at 20 MPa appears to be different to those for samples deformed at pressures  
328  $\leq 15$  MPa, there is a subtle stress drop (following an axial strain of  $\sim 5\%$ ; Figure 8a).  
329 The sample deformed at 20 MPa can therefore be considered brittle (the post-  
330 deformation sample contained a shear fracture), but its style of failure is very close to  
331 the transition between brittle and ductile behaviour.

332         The stress–strain curves for the samples deformed at 30, 50, and 80 MPa are  
333 typical of ductile behaviour: there is no peak stress or stress drop associated with the  
334 formation of a shear fracture (Figure 8a; Wong and Baud, 2012). Further, the  
335 experiments performed at 50 and 80 MPa show strain-hardening behaviour (where  
336 differential stress increases as a function of axial strain: Figure 8a). For sample AU1,  
337 the experiments performed at  $P_{\text{eff}}$  of 0, 2.5, and 5 MPa deformed in a brittle manner,  
338 whereas the experiments performed at  $P_{\text{eff}}$  of 10, 15, and 20 MPa were ductile (Figure  
339 8b). The tests performed at  $P_{\text{eff}}$  of 15 and 20 MPa show strain-hardening behaviour  
340 (Figure 8b). The block from the quarry at Aumühle (AU1) therefore transitioned from  
341 brittle to ductile behaviour at a lower effective pressure (i.e., depth) than the block from  
342 Seelbronn (SUE).

343

### 344 3.5 Post-deformation microstructures

345         To better understand the operative micromechanical processes during brittle and  
346 ductile behaviour, thin sections were prepared of the samples of SUE deformed at  $P_{\text{eff}}$   
347 of 10 and 80 MPa. Backscattered SEM images of the sample deformed at 10 MPa (i.e.,

348 in the brittle regime) show a large shear fracture and associated off-fault microcracking  
349 (orientated sub-parallel to the maximum principal stress) (Figures 9a and 9b). The  
350 sample deformed at 80 MPa (i.e., in the ductile regime) did not contain a shear fracture.  
351 Instead, deformation was accommodated by the cataclastic collapse of pores (Figures  
352 9c and 9d). Microcracks are seen to radiate from the collapsed pores (Figure 9c and  
353 9d).

354

## 355 **4 Discussion**

356

### 357 4.1 The permeability of impact melt-bearing breccia

358 Although data in the literature are rare, the permeability of the suevites  
359 measured herein (Table 2) compare well with published measurements. For example,  
360 Parnell et al. (2010) determined the permeability of two samples from the Ries crater  
361 to be  $1.58 \times 10^{-15}$  and  $2.67 \times 10^{-15}$  m<sup>2</sup>. These authors also reported the permeability of  
362 samples from the impact craters at Rochechouart and Haughton as  $\sim 1 \times 10^{-15}$  m<sup>2</sup> and  
363 between  $\sim 1 \times 10^{-16}$  and  $\sim 1 \times 10^{-15}$  m<sup>2</sup>, respectively. The permeability of impact melt-  
364 bearing breccia from the Chicxulub impact crater was estimated as between  $\sim 1 \times 10^{-18}$   
365 and  $\sim 1 \times 10^{-15}$  m<sup>2</sup> (Mayr et al., 2008b). These measurements highlight a remarkable  
366 similarity worldwide in permeability of impact melt-bearing breccia from different  
367 impact craters formed within different lithologies. In strong contrast, the permeability  
368 of lavas of the same composition, erupted from the same volcano, can vary by as much  
369 as five or six orders of magnitude (e.g., Farquharson et al., 2015).

370 We compare our suite of data with a model for the permeability of partially  
371 welded granular media (Wadsworth et al., 2016) that has been validated against  
372 geological porous materials including welded volcanic rocks and sandstones of various



373 diagenetic maturity. This model starts from the assumption that the system was first a  
374 granular deposit of particles with mean radius  $\langle r \rangle$ . Because suevite-forming systems  
375 begin as droplets and particles deposited in density currents, this is a valid starting  
376 assumption, and consistent with similar cases such as the permeability of variably  
377 welded deposits of pyroclastic density currents. From that initial granular state, the  
378 model assumes that the particles weld together, reducing the internal pore surface areas.  
379 Typically, the Wadsworth et al. (2016) model would require a solution to relate the  
380 particle size distribution centred around  $\langle r \rangle$  to a distribution of sizes of characteristic  
381 pores centred around  $\langle a \rangle$ . However, here we have directly measured  $\langle a \rangle$  (Figure 4),  
382 negating the need for this interim step.

383 We take the  $\langle a \rangle$  from the SUE measurement of pore-throat sizes (Figure 4) as  
384 intermediate among the three measurements, and find that  $\langle a \rangle = 1.09 \mu\text{m}$  (we  
385 therefore assume that the pore-throat size is a good approximation of the pore size). We  
386 then assume that the pore surface area,  $s$ , is related by (Torquato, 2013):

387

$$388 \quad s = -\frac{3(1 - \phi) \ln(1 - \phi)}{\langle a \rangle} \quad (1).$$

389

390 This relation assumes that  $s$  is given by the surface area of a suite of overlapping  
391 spherical pores (Wadsworth et al., 2016). The permeability,  $k$ , is given by (Wadsworth  
392 et al., 2016; Martys et al., 1994):

393

$$394 \quad k = \frac{2[1 - (\phi - \phi_c)]}{s^2} (\phi - \phi_c)^n \quad (2),$$

395

396 where  $\phi_c$  is the percolation threshold below which all the pores have pinched off, and  
397 the permeability is zero, and  $n$  is a percolation exponent. On theoretical grounds,  $n =$   
398 4.4 was derived for porous overlapping sphere systems (Halperin et al., 1985), leaving  
399 only  $\phi_c$  unconstrained in this model.

400 We fit Equation (2) to our complete dataset, allowing  $\phi_c$  to vary freely using a  
401 root-mean-square error between model and data as a minimisation target. We find that  
402 the best results are found using  $\phi_c = 0.075$  and that to within 5% accuracy,  $0.06 \lesssim$   
403  $\phi_c \lesssim 0.1$  is valid. This range of values is consistent with the range of  $\phi_c$  found for  
404 sintering or welding angular particles (Wadsworth et al., 2016; Heap et al. 2019).  
405 Therefore, what we present is a theoretically grounded model for the permeability of  
406 welded granular media that is applicable to suevites across the range of porosities  
407 measured herein (Figure 10).

408

409 4.2 The uniaxial compressive strength and Young's modulus of impact melt-bearing  
410 breccia

411 To our knowledge, the values of uniaxial compressive strength for suevites  
412 reported here (Table 2) and in Heap et al. (2020a) are the only such values yet reported  
413 for this rock type. Nonetheless, we find that the samples we assay here have properties  
414 similar to other rock types found on Earth. For example, poorly sorted clastic volcanic  
415 rocks from Mt. Meager volcano, Canada—welded block-and-ash flow deposits—have  
416 a similar strength to the samples we report here. The uniaxial compressive strength of  
417 welded block-and-ash flow from Mt. Meager volcano is  $\sim 20$  MPa when the porosity is  
418 between 0.2 and 0.25 (i.e., the porosity range of the samples measured herein: Table 2)  
419 (Heap et al., 2015a); this strength value lies between those we find for the Altenburg  
420 (ALT) and Aumühle (AU) samples (Figure 6b).

421 Our data show that uniaxial compressive strength decreases as a function of  
422 increasing porosity (Figure 6b), in agreement with studies on sedimentary (e.g., Chang  
423 et al., 2006; Baud et al., 2014) and volcanic rocks (e.g., Al-Harathi et al., 1999; Heap et  
424 al., 2014a; Schaefer et al., 2015; Harnett et al., 2019). Although the relatively high  
425 porosity of the samples from Aumühle quarry could explain their relative weakness, we  
426 also note that studies have shown that alteration can result in a strength reduction (e.g.,  
427 del Potro and Hürlimann, 2009; Pola et al., 2012; Wyering et al., 2014; Frolova et al.,  
428 2014). The blocks from the Aumühle quarry are more altered than those from Seelbronn  
429 or Altenbürg: more of the amorphous phases have been altered to smectite (Table 1).  
430 Therefore, the relative weakness of samples from Aumühle (Figure 6b) can be  
431 explained by a combination of their higher porosity and greater degree of alteration.

432 The pore-emanating crack model of Sammis and Ashby (1986) is often used to  
433 better understand the mechanical behaviour of porous materials, including rocks (e.g.,  
434 Baud et al., 2014). This model describes a two-dimensional elastic medium populated  
435 by circular pores of uniform radius  $r$ . Tensile cracks emanate (in the direction of the  
436 maximum principal compressive stress) when the applied stress reaches the critical  
437 stress intensity factor,  $K_{IC}$  (sometimes called the fracture toughness). These cracks then  
438 propagate and coalesce to induce the macroscopic failure of the elastic medium. An  
439 analytical approximation of the pore-emanating crack model of Sammis and Ashby  
440 (1986) was derived by Zhu et al. (2010) in the case of uniaxial compression, to model  
441 the uniaxial compressive strength,  $\sigma_p$ , as a function of porosity,  $\phi$ :

442

$$443 \quad \sigma_p = \frac{1.325 K_{IC}}{\phi^{0.414} \sqrt{\pi r}} \quad (3).$$

444

445           The two unknowns in the equation are  $K_{IC}$  and  $r$ . Therefore, at this stage, it is  
446 prudent to first estimate the value or values of the term  $K_{IC}/\sqrt{\pi r}$  that best describe the  
447 experimental data. Our experimental data cannot be well described using a single value  
448 of  $K_{IC}/\sqrt{\pi r}$  (Figure 11). The data can be bracketed between two theoretical curves  
449 where  $K_{IC}/\sqrt{\pi r}$  equals 3.5 and 20 MPa (Figure 11). In the absence of measurements of  
450  $K_{IC}$  for impact melt-bearing breccia, we can use the value of  $K_{IC}$  identified to describe  
451 the welded block-and-ash flow deposits from Mt. Meager volcano:  $K_{IC} = 0.15 \text{ MPa}\cdot\text{m}^{-1/2}$   
452 (Heap et al., 2015a). This value of  $K_{IC}$  yields values of  $r$  between  $\sim 18$  and  $\sim 585 \mu\text{m}$ .  
453 Therefore, if  $K_{IC} = 0.15 \text{ MPa}\cdot\text{m}^{-1/2}$  is an appropriate value, our modelling suggests that  
454 strength of the suevites is controlled by the largest pores (see our microstructural  
455 analysis and our mercury injection data; Figures 3 and 4). Of note, the pore-emanating  
456 crack model of Sammis and Ashby (1986) describes a medium that contains pores of a  
457 uniform radius, which we know to be a simplification of our experimental materials  
458 (see Figures 3 and 4). We further emphasise that the  $K_{IC}$  of the block from Aumühle  
459 may be lower than that for the other blocks due to the greater alteration intensity (Table  
460 1). Finally, the blocks from Seelbronn and Altenbürg also contain microcracks, which  
461 are not considered in the Sammis and Ashby (1986) model. Nevertheless,  
462 micromechanical models such as that proposed by Sammis and Ashby (1986) can help  
463 glean information as to the first-order controls on the mechanical behaviour and  
464 strength of brittle materials.

465           To our knowledge, the values of Young's modulus for suevites reported here  
466 (Table 2) are the only such values reported. Our data show that Young's modulus  
467 decreases as a function of increasing porosity (Figure 7), in agreement with previous  
468 studies (e.g., Heap et al., 2020b). Chemical alteration is also known to influence the  
469 Young's modulus of rocks and can result in increases to Young's modulus when the

470 alteration is pore-filling (e.g., Frolova et al., 2014; Mordensky et al., 2018; Durán et al.,  
471 2019) and decreases to Young's modulus in the case of dissolution (e.g., Frolova et al.,  
472 2014). The low Young's modulus of samples from Aumühle (Figure 7) can therefore  
473 be explained by a combination of their higher porosity and greater degree of alteration.

474

#### 475 4.3 Failure envelopes for impact melt-bearing breccia

476 The mechanical behaviour of porous materials is best compared on a graph of  
477 differential stress,  $Q$ , as a function of the effective mean stress,  $P$  (Wong and Baud,  
478 2012).  $P$  is given by  $P = (\sigma_1 + 2\sigma_3)/3 - Pp$ , where  $Pp$  is the pore fluid pressure (0  
479 MPa in our experiments),  $\sigma_1$  is the maximum principal compressive stress (the axial  
480 stress in our experiments), and  $\sigma_3$  is the minimum principal compressive stress (the  
481 confining pressure in our experiments). The failure envelope of a material can be  
482 delineated by plotting the differential stress at failure as a function of the effective mean  
483 stress at failure. Deformation experiments must therefore be performed at a range of  
484 effective pressures to map the failure envelope.

485 Failure in the brittle regime is defined as the differential stress at the peak stress  
486 (as indicated in Figure 6a), and failure in the ductile regime is defined as the differential  
487 stress at the onset of inelastic compaction, termed  $C^*$  (Wong and Baud, 2012).  $P^*$  is  
488 the term given for the onset of inelastic compaction under hydrostatic conditions (i.e.,  
489 when  $Q = 0$ ) (e.g., Zhang et al., 1990; Wong and Baud, 2012).  $C^*$  was determined  
490 using the evolution of the volume of the sample during deformation, monitored by the  
491 confining pressure intensifier.  $C^*$  can be observed as an inflection point in the data of  
492 differential stress as a function of the change in confining fluid volume (i.e., sample  
493 volume).

494 In our experiments in which we held the differential stress at 2 MPa and  
495 increased the confining pressure,  $C^*$  was 2 MPa and  $P$  was calculated using the  
496 confining pressure required for inelastic compaction (monitored using the axial strain  
497 signal). A sample can be considered pre-failure if the state of stress plots inside the  
498 failure envelope, and failed if the state of stress plots outside the failure envelope. The  
499 sample will have failed in a brittle or ductile manner if the stress state falls outside the  
500 failure envelope to the left or right, respectively. The failure envelopes for the suevites  
501 from Seelbronn (SUE) and Aumühle (AU1) are shown in Figure 12 (with data available  
502 in Table 3). The failure envelope for SUE is much larger than that for AU1. In other  
503 words, SUE is pre-failure for much more of the stress space represented. The transition  
504 from brittle to ductile behaviour, represented by the “top” of the failure envelope, is at  
505 an effective pressure of  $\sim 30$  and  $\sim 10$  MPa for SUE and AU1, respectively (Table 3).  
506 As discussed above, the relative weakness of the blocks from Aumühle is likely due to  
507 a combination of higher porosity (Tables 2 and 3) and greater alteration intensity (Table  
508 1) than for the samples from Seelbronn or Altenbürg. The position of  $P^*$  for SUE and  
509 AU1 is estimated to be  $90 \pm 5$  and  $22 \pm 5$  MPa, respectively (Figure 12).

510

#### 511 4.4 Deformation micromechanisms

512 Our microstructural work indicates that the operative deformation  
513 micromechanisms of these suevite samples are microcracking and cataclastic pore  
514 collapse in the brittle and ductile fields, respectively (Figure 9a and 9b). Failure in the  
515 brittle regime is the result of the initiation and propagation of microcracks,  
516 preferentially aligned to the direction of maximum principal compressive stress (Figure  
517 9a and 9b), which coalesce to form a macroscopic, through-going shear fracture (see  
518 also Lockner et al., 1991; Aben et al., 2019). Porosity reduction in the ductile regime is

519 driven by the cataclastic collapse of pores (Figure 9c and 9d). This deformation  
520 micromechanism has been previously observed in porous limestones (e.g., Zhu et al.,  
521 2010; Vajdova et al., 2012) and porous volcanic rocks (e.g., Loaiza et al., 2012;  
522 Adelinet et al., 2013; Zhu et al., 2011; Heap et al., 2015b, 2015c; Zhu et al., 2016).

523 Of particular interest, because of their ability to reduce permeability and create  
524 permeability anisotropy, is whether such compaction is localised into bands.  
525 Compaction bands have been well documented in studies on porous sandstone, and  
526 form as a result of grain crushing (e.g., Baud et al., 2004, 2006; Fortin et al., 2006;  
527 Stanchits et al., 2009; Baud et al., 2015). Compaction bands have also been observed  
528 in porous limestones (e.g., Cilona et al., 2014; Baud et al., 2017; Huang et al., 2019)  
529 and porous volcanic rocks (e.g., Loaiza et al., 2012; Adelinet et al., 2013; Heap et al.,  
530 2015b, 2017a). We observed no evidence of compaction bands in our samples: pore  
531 collapse (Figure 9c and 9d) was distributed throughout the sample. In addition, we did  
532 not observe small stress drops in the ductile stress–strain curves (Figure 8), which have  
533 also been associated with the formation of compaction bands in laboratory experiments.  
534 Therefore, similar to porous tuff from Alban Hills, Italy (Zhu et al., 2010) and Whakaari  
535 volcano, New Zealand (Heap et al., 2015c), i.e., granular volcanic rocks with a similar  
536 genesis and texture to the suevites studied herein, inelastic compaction in suevite does  
537 not appear to localise into bands. The absence of compaction bands in these materials  
538 may be related to their wide grain size distribution (e.g., Figure 3), a factor known to  
539 suppress compaction localisation in sandstones (Cheung et al., 2012).

540

#### 541 4.5 Micromechanics of hydrostatic compaction

542 To investigate the micromechanics of hydrostatic compaction, we follow the  
543 approach outlined by Zhu et al. (2011). By considering the stresses required to yield a

544 single macropore embedded within an effective medium, and by adopting the Mohr–  
545 Coulomb failure criterion, the onset of inelastic hydrostatic compaction,  $P^*$ , is related  
546 to uniaxial compressive strength by (Zhu et al., 2011):

547

$$548 \quad P^* = 2\sigma_p/3 \quad (4).$$

549

550 Figure 13a shows  $P^*$  as a function of uniaxial compressive strength for the SUE and  
551 AU1 suevites and a collection of tuffs (volcanic rocks that share textural similarities to  
552 the studied suevites). All of these data plot above the theoretical line delineated by  
553 Equation (4). As discussed in Zhu et al. (2011), this distribution of samples is likely  
554 because the effective medium is less porous than the matrix of these rocks, i.e., the  
555 matrix contains numerous micropores, as exposed by our microstructural observations  
556 (Figure 3) and our mercury injection experiments (Figure 4). We can study the  
557 influence of micropores on the onset of inelastic hydrostatic compaction by combining  
558 Equations (3) and (4) (Zhu et al., 2011):

559

$$560 \quad S^* = \frac{K_{IC}}{(\phi_*/\phi)^{0.414} \sqrt{\pi a^*}} \quad (5),$$

561

562 where  $a^*$  is the average radius of a micropore,  $\phi_*$  is the porosity of the effective  
563 medium, and  $S^*$  captures the cooperative influence of the fracture toughness, micropore  
564 radius, and the partitioning of macro- and microporosity (Zhu et al., 2011). This model  
565 describes a dual porosity medium of a macropore surrounded by many micropores. We  
566 find that SUE and AU1 are best described by values of  $S^*$  of 70 and 14 MPa,



567 respectively (Figure 13b). We note that these values of  $S^*$  are similar to those predicted  
568 for tuffs (Figure 13b; Zhu et al., 2011).

569 The average micropore radius can then be estimated using  $a^* > (K_{IC}/S^*)^2/\pi$   
570 (Zhu et al., 2011). Assuming  $K_{IC} = 0.15 \text{ MPa}\cdot\text{m}^{-1/2}$  (Heap et al., 2015a), the average  
571 micropore radius for SUE and AU1 are 1.5 and 36.5  $\mu\text{m}$ , respectively. These pore radii  
572 predictions are in agreement with the pore throat radii determined by our mercury  
573 injection experiments (Figure 4). The modelling described above therefore highlights  
574 an important role for micropores in controlling the onset of inelastic hydrostatic  
575 compaction in the studied materials.

576

#### 577 4.6 Upscaling laboratory measurements for use in large-scale modelling

578 The laboratory experiments described herein were performed on samples with  
579 diameters of 20 or 25 mm. These measurements, therefore, do not consider the  
580 influence of macroscopic fractures, which are well known to influence permeability  
581 (e.g., Brace, 1984; Clauser, 1992; Neuman, 1994; Heap and Kennedy, 2016;  
582 Farquharson and Wadsworth, 2018), strength (e.g., Hoek et al., 2002), and Young's  
583 modulus (e.g., Hoek and Diederichs, 2006). Since the crusts of heavily impacted  
584 planetary bodies are likely highly fractured (e.g., Wiczorek et al., 2013; Heap et al.,  
585 2017b; Lognonné et al., 2020), these laboratory values must be upscaled before they  
586 can be used in large-scale models.

587 The strength of a fractured rock-mass can be estimated with the generalised  
588 Hoek–Brown failure criterion (Hoek et al., 2002):

589

590 
$$\sigma'_1 = \sigma'_3 + C_o \left( m_b \frac{\sigma'_3}{C_o} + s \right)^a \quad (6),$$

591

592 where  $\sigma'_1$  and  $\sigma'_3$  are the effective maximum and minimum principal stresses,  
593 respectively,  $C_o$  is the laboratory-measured uniaxial compressive strength, and  $m_b$ ,  $s$ ,  
594 and  $a$  are unitless fitting parameters (Hoek et al., 2002):

595

596 
$$m_b = m_i \exp\left(\frac{GSI - 100}{28 - 14D}\right) \quad (7),$$

597 
$$s = \exp\left(\frac{GSI - 100}{9 - 3D}\right) \quad (8),$$

598 
$$a = \frac{1}{2} + \frac{1}{6} \left[ \exp\left(-\frac{GSI}{15}\right) + \exp\left(-\frac{20}{3}\right) \right] \quad (9).$$

599

600 GSI is the Geological Strength Index, a unitless value between 0 and 100 that  
601 describes the rock-mass characteristics (Marinos et al., 2005). High values of GSI  
602 represent intact or massive rock masses, whereas low GSI values represent blocky and  
603 fractured rock masses with weathered fracture surfaces and clay-filled fractures  
604 (Marinos et al., 2005).  $D$  is a unitless disturbance factor related to anthropogenic  
605 blasting damage in large excavations (we therefore use  $D = 0$  in our study; the full  
606 equation is provided here for completeness). Finally, unitless constant  $m_i$  describes the  
607 shape of the failure envelope on a graph of  $\sigma'_1$  as a function of  $\sigma'_3$ , and is therefore  
608 related to the microstructural and mineralogical attributes of the rock (Eberhardt, 2012).

609 To provide strengths for suevite rock masses, we require estimates for  $C_o$ ,  $m_i$ ,  
610 and GSI. Values of  $C_o$  can be taken directly from our experimental data, and we take  
611 values of 44.7 and 14.5 MPa for the SUE and AU1 suevites, respectively (excluding  
612 outliers). Values of  $m_i$  can be estimated using our triaxial experimental data. We  
613 estimate  $m_i$ , using the data fitting function (modified cuckoo fit algorithm, basic error  
614 summation, and absolute error type) in RocData (Rocscience, 2017), to be 12 and 4.7

615 for the SUE and AU1 suevites, respectively (Figure 14). In the absence of GSI  
616 estimates, we use the suggested average for volcanic rock masses of 55 suggested by  
617 Heap et al. (2020b). However, because of the variability in GSI values for natural  
618 outcrops (estimates of GSI for volcanic rock outcrops, for example, typically range  
619 between disintegrated (GSI = 20-40) and blocky (GSI = 65-85); see Heap et al., 2020b  
620 and references therein), we provide rock mass strength estimates for GSI = 45-65 (i.e.,  
621 GSI =  $55 \pm 10$ ). We can now model the compressive strength of SUE and AU1 as a  
622 function of depth (using  $P = \rho gh$ , where  $P$  is the lithostatic pressure,  $h$  is the depth,  
623  $g$  is the acceleration due to gravity, and  $\rho$  is the bulk rock density (taken as  $2000 \text{ kg/m}^3$ ,  
624 guided by our laboratory density measurements)) (Figure 15). In the following analyses  
625 we provide strength estimates for rock masses on Mars (i.e.,  $g = 3.711 \text{ m/s}^2$ ).

626 We compare our strength estimations for SUE and AU1 with that for basalt  
627 (Figure 15), which is lithologically representative of much of the surface materials of  
628 the inner Solar System worlds (e.g., Byrne, 2019). To do so, we performed six uniaxial  
629 compressive strength experiments on dry samples of Volvic basalt (Chaîne des Puys,  
630 France). The connected porosity and uniaxial compressive strength of these samples  
631 were determined as described in the methods section. To estimate the strength of a  
632 basaltic rock-mass, we used a value of  $C_o$  of 92.6 MPa (Table 4), an  $m_i$  of 22 (see  
633 Schultz, 1995), and a GSI of  $55 \pm 10$  (Heap et al., 2020b). We increased the bulk rock  
634 density from 2000 to  $2300 \text{ kg/m}^3$ , as guided by our laboratory density measurements  
635 for Volvic basalt. We find that the upscaled compressive strength of suevite is much  
636 lower than that of basalt (Figure 15). For example, the rock-mass strength of SUE, AU1,  
637 and Volvic basalt are, respectively,  $\sim 11$ ,  $\sim 25$ , and  $\sim 45$  MPa at a depth of 0.5 km (Figure  
638 15). Furthermore, the basalt will remain brittle over the range of depth (up to 5 km)  
639 shown in Figure 15; the suevites from Seelbronn (SUE) and Aumühle (AU1) will be

640 ductile at depths of about 1 and 4 km, respectively (Figure 15). We note that, based on  
 641 the depth of impact craters (e.g., Boyce et al., 2005; Stepinski et al., 2009), impact melt-  
 642 bearing breccias are unlikely to exist below a couple of kilometres in the lithosphere.  
 643 We further note that we use data for Volvic basalt for illustrative purposes: the porosity  
 644 of basalt on Mars will certainly vary and its strength is very much porosity-dependent  
 645 (e.g., Al-Harhi et al., 1999).

646 The Young's modulus of a fractured rock-mass,  $E_{rm}$ , can be estimated using  
 647 the Hoek–Diederichs equation (Hoek and Diederichs, 2006):

648

$$649 \quad E_{rm} = E_i \left[ 0.02 + \left( 1 - \frac{D}{2} \left[ 1 + \exp \frac{(60 + 15D - GSI)}{11} \right]^{-1} \right) \right] \quad (10).$$

650

651 Here,  $E_i$  is the “intact” Young's modulus measured from the elastic portion of the  
 652 stress–strain curve of a rock deforming in compression (i.e., the data of Figure 7; Table  
 653 2). We take  $D = 0$  (see above) and assume an average GSI of 55 (Heap et al., 2020b).  
 654 We also determined the intact (Table 4) and rock-mass Young's modulus for Volvic  
 655 basalt. The data are presented in Figure 16 and show that the Young's modulus of a  
 656 rock-mass is much lower than intact Young's modulus. For example, the range of  
 657 Young's modulus for SUE suevite was reduced from 3.5–8.1 to 1.4–3.3 GPa (Figure  
 658 16). These data also show that the intact and rock-mass Young's modulus of suevite is  
 659 much lower than that for basalt (Figure 16). Of course, the Young's modulus of Volvic  
 660 basalt represents that value of a basalt with a porosity of 0.2 (Table 4), but the porosity,  
 661 and therefore Young's modulus, of basalt on Mars likely varies considerably. As for  
 662 our rock mass strength estimations, we can also provide rock mass Young's modulus  
 663 estimations for  $GSI = 55 \pm 10$ . The average rock mass Young's moduli in this GSI

664 range are 1.5-4.1, 1.3-3.6, 0.7-1.9, and 3.6-10.2 GPa for SUE, ALT, AU, and Volvic  
665 basalt, respectively.

666 The permeability of a rock-mass can be estimated if the macrofracture density  
667 and width, the permeability of the host-rock, and the permeability of the fractures is  
668 known (e.g., Heap and Kennedy, 2016). Once rock-mass permeabilities are known,  
669 these values can be used to estimate permeability on larger scales (e.g., Farquharson  
670 and Wadsworth, 2018). The permeability of a rock-mass,  $k_e$ , can be estimated using  
671 the following two-dimensional model that considers flow in parallel layers:

672

$$673 \quad k_e = \frac{(w_i \cdot k_0) + (w_f \cdot k_f)}{W} \quad (11),$$

674

675 where  $w_i$  and  $k_0$  and  $w_f$  and  $k_f$  and are the width and permeability of the intact host-  
676 rock and fractures, respectively, and  $W$  is the width of the rock-mass. If we take  $k_0$  to  
677 be  $5.0 \times 10^{-15} \text{ m}^2$  (Table 2), and assume a constant fracture width of 0.25 mm and a  $k_f$   
678 of  $1.0 \times 10^{-9} \text{ m}^2$  (see Heap and Kennedy, 2016), we can plot rock-mass permeability as  
679 a function of fracture density (Figure 17). To compare these data with those for basalt,  
680 we measured the permeability of one sample of Volvic basalt using the approach  
681 described in the methods section (Table 4). For the basalt calculations, we used  $k_0 =$   
682  $2.4 \times 10^{-14} \text{ m}^2$  (Table 4), and again assumed a constant fracture width of 0.25 mm and  
683 a  $k_f$  of  $1.0 \times 10^{-9} \text{ m}^2$  (Figure 17). We assume a constant fracture width of 0.25 mm  
684 because this fracture width corresponds to the laboratory-measured value of  $k_f$  (see  
685 Heap and Kennedy, 2016). The permeability of large-aperture fractures is likely higher  
686 than the values of  $k_f$  provided herein, the permeability of which could be estimated  
687 using  $k_f = h^2/12$  (Zimmerman and Bodvarsson, 1996), where  $h$  is the fracture width.

688           We find that, at very low fracture densities (between 0 and 2 fractures per  
689 metre), the basaltic rock-mass is more permeable than the suevite rock-mass (Figure  
690 17). Above two fractures per metre, the equivalent permeabilities of the basalt and  
691 suevite rock masses are very similar (Figure 17). Figure 17 also shows that only a few  
692 macroscopic fractures per metre can increase the permeability of the rock-mass by two  
693 orders of magnitude. Above a macroscopic fracture density of about 10 fractures per  
694 metre, the equivalent permeability of the rock-mass does not increase significantly as a  
695 function of increasing fracture density (Figure 17). We emphasise that the permeability  
696 of basalt on the sample scale can be very low (e.g.,  $10^{-19}$  m<sup>2</sup>; Nara et al., 2011) and will  
697 depend on both porosity and void-space connectivity. Therefore, the equivalent  
698 permeability of low porosity basalt on Mars may be considerably lower than the  
699 estimates for Volvic basalt provided herein, especially at low macrofracture densities.

700           Although there are a number of assumptions built into our upscaling  
701 calculations (for strength, Young's modulus, and permeability), these calculations are  
702 designed to emphasise that care should be taken when deciding on physical property  
703 values for large-scale models.

704

#### 705 4.7 Implications for impacted planetary bodies

706           Our findings show that suevites can have high porosities and low strength values  
707 (e.g., Figure 15), which is not surprising given the destructive nature of large impact  
708 events and the damage to rock masses that ensues (e.g., Morgan et al., 2016). For  
709 example, the compressive strength of suevite can be 2–4 times lower than that for  
710 basalt, a rock representative of much of the surface materials of the inner Solar System  
711 worlds (e.g., Byrne, 2019), at a depth of 0.5 km (for Mars; Figure 15). Our experiments

712 also show that suevite will be ductile at shallow depths (Figure 8), which will likely  
713 affect, for example, the propagation of large fractures and dykes.

714 Most of inner Solar System worlds show evidence for protracted histories of  
715 impact bombardment. The “intercrater plains” on Mercury, which occupy ~75% of the  
716 planet surface (Whitten et al., 2014; Byrne et al., 2018), are heavily cratered and record  
717 more than 4 Gyr of impacts (Marchi et al., 2013). The southern uplands on Mars are  
718 similarly cratered and comparably old (Tanaka et al., 2014), as is much of the Moon’s  
719 surface, especially the lunar highlands (e.g., Fassett et al., 2012b). Our study suggests  
720 that the rock masses that comprise these older, bombarded crusts are therefore  
721 substantially weaker and more porous and permeable than the younger plains material  
722 on these worlds, e.g., Mercury’s smooth plains, the northern lowland plains on Mars,  
723 and the lunar maria (cf. Byrne, 2019). Indeed, the difference in tectonic style of crustal  
724 shortening between the intercrater and smooth plains on Mercury has been attributed to  
725 the destruction, via impact bombardment, of any internal fabric in the former terrain  
726 type that the latter may still possess (Byrne et al., 2014). And although further  
727 weakening of crustal rocks by alteration likely did not have affect the airless bodies,  
728 e.g., Mercury or the Moon, such systems were probably present on Mars (Rathbun and  
729 Squyres, 2002; Osinski et al., 2013). (We exclude Venus from this discussion because  
730 the second planet has anomalously few impact craters (Phillips et al., 1991), and it not  
731 clear what effect(s) its high mean surface temperature (~470 °C) might have had on  
732 those rocks that were impacted.)

733 In any case, our results here add to the few available data for the strength of  
734 planetary materials damaged by impact bombardment. It is therefore prudent to use  
735 such values in calculations of, for instance, the strength of the upper lithosphere of  
736 planetary bodies, when employing values for intact rock can lead to substantially

737 different outcomes than for—more realistically—weak rock (e.g., Klimczak, 2015;  
738 Klimczak et al., 2015). Indeed, results from the NASA Gravity Recovery and Interior  
739 Laboratory (GRAIL) mission showed the Moon to have an average crustal porosity of  
740 12% to a depth of at least several kilometres (Wieczorek et al., 2013), varying  
741 regionally to as much as 22%—comparable to those values we find for the Seelbronn  
742 (SUE) and Altenbürg (ALT) samples.

743

## 744 **5 Conclusions**

745 We provide here petrophysical property data (porosity, permeability, Young's  
746 modulus, and uniaxial compressive strength) for suevite (a type of melt-bearing impact  
747 breccia) from the Ries crater in Germany. We further performed triaxial experiments  
748 to determine the pressure (i.e., depth) required for the brittle–ductile transition. Until  
749 now, petrophysical and mechanical data for melt-bearing impact breccia were rare or  
750 absent. It is our hope that these data, and our upscaling approaches, can improve fluid  
751 flow modelling, slope and crater stability assessments, crustal strength estimates, and  
752 physical weathering rate estimates for and the interpretation of large-scale structural  
753 features on planetary bodies on which protracted records of impact bombardment are  
754 preserved.

755

## 756 **Acknowledgements**

757 The first author acknowledges a Prix d'Espoirs grant, awarded by the University  
758 of Strasbourg. Olivier Lengliné is warmly thanked for driving the author to and from  
759 Nördlingen (Germany). Alexandra Kushnir is thanked for field assistance, Patrick Baud  
760 for discussions, Bertrand Renaudié for his help with the preparation of samples, and  
761 Marlène Villeneuve for determining the values of  $m_i$  using RocData. We thank Lea

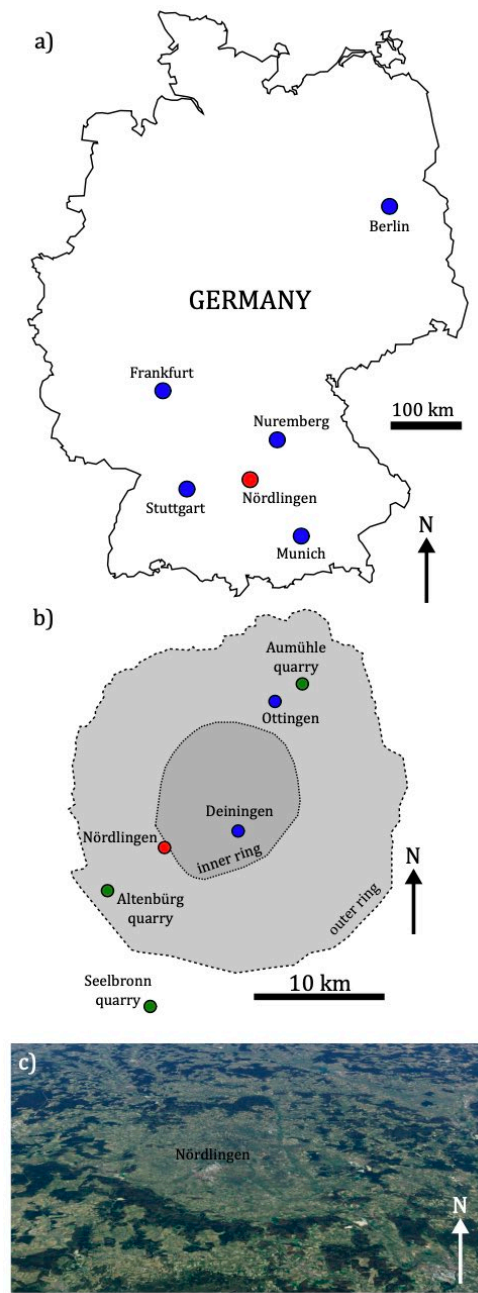


762 Mertens and Matthias Wittner for providing the suevite block from the Seelbronn  
763 quarry (Wittner - Steinbearbeitung und Denkmalpflege, Nördlingen, Germany). PKB  
764 acknowledges support from North Carolina State University. We thank Chris Okubo  
765 and one anonymous reviewer for comments that helped improve this manuscript.

766

767 **Data availability**

768 The data collected for this study can be found in Tables 1–4.



770

771 **Figure 1.** (a) Map of Germany showing the location of Nördlingen (in red). (b) Map  
 772 showing the outer- (dashed line) and inner-ring (dotted line) of the Ries impact crater.  
 773 The red circle shows the location of the town of Nördlingen. Green circles indicate the  
 774 locations of the quarries from where the experimental materials were sourced  
 775 (Aumühle, Altenbürg, and Seelbronn). (c) GoogleEarth™ image of the crater showing  
 776 the location of Nördlingen. The crater is approximately 20 km in diameter.

25 mm



Seelbronn (SUE)



Altenbürg (ALT)



Aumühle (AU)

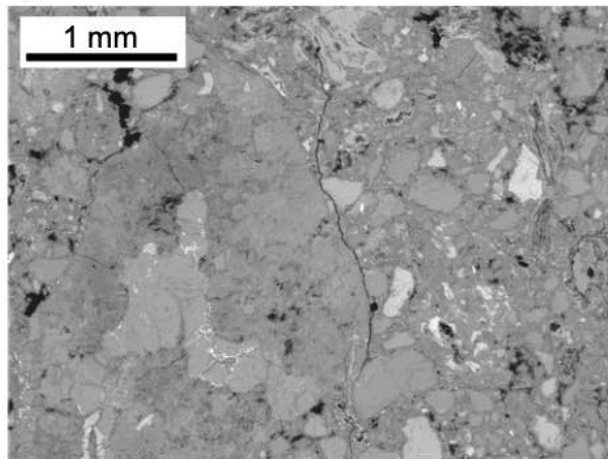
777

778 **Figure 2.** Photographs of representative 25 mm-diameter cylindrical core samples

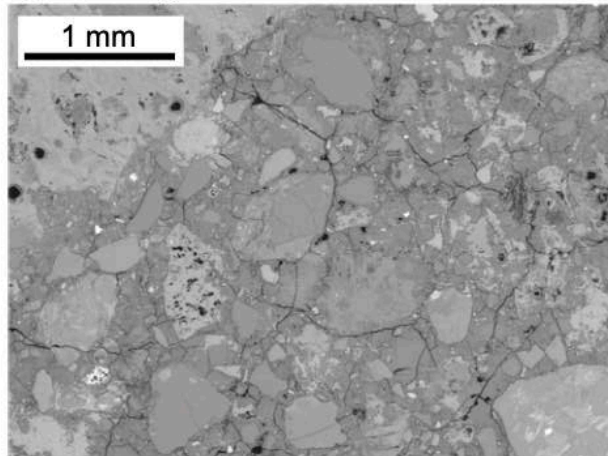
779 prepared from the blocks collected. The symbol below each photograph corresponds to

780 the symbols used in Figures 5, 6, and 7.

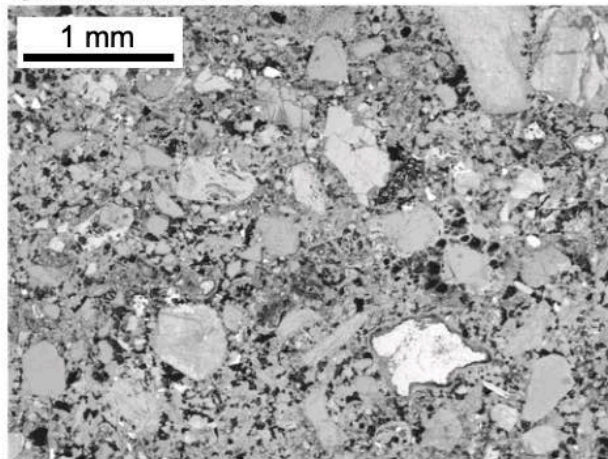
a) Seelbronn suevite



b) Altenbürg suevite



c) Aumühle suevite



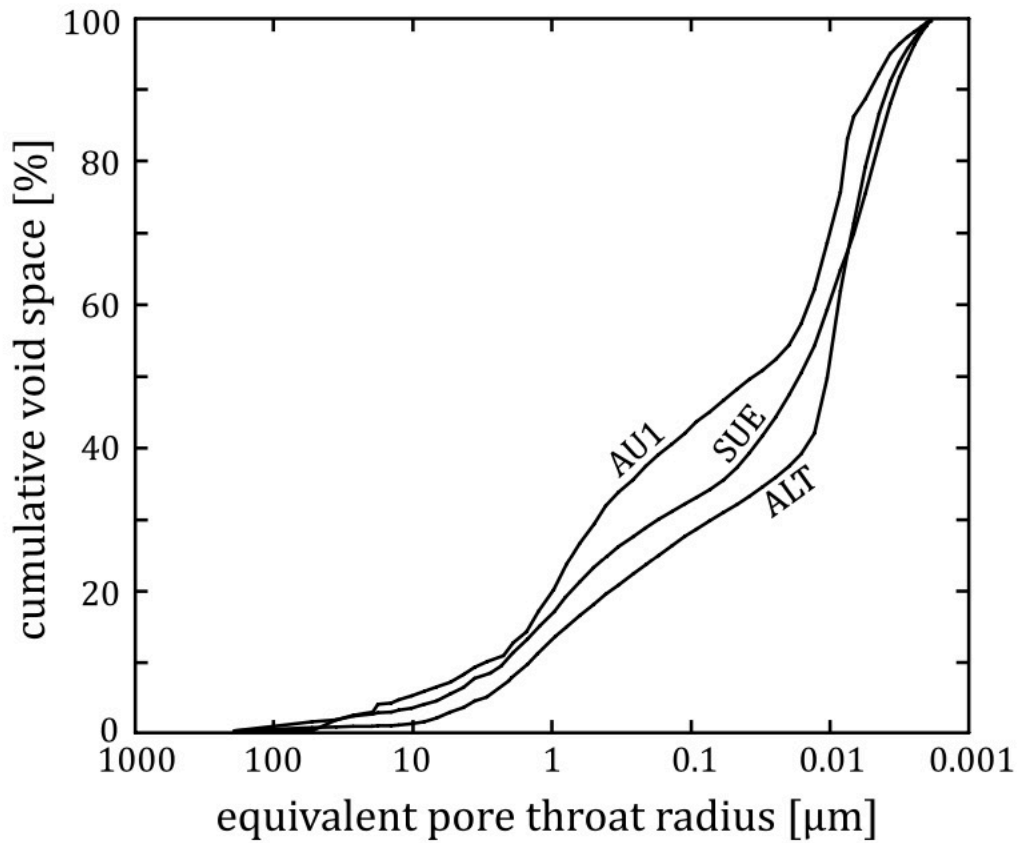
781

782 **Figure 3.** Backscattered electron microscope images of impact melt-bearing breccias

783 (“suevites”) from each of the three quarry sites: (a) Seelbronn; (b) Altenbürg; and (c)

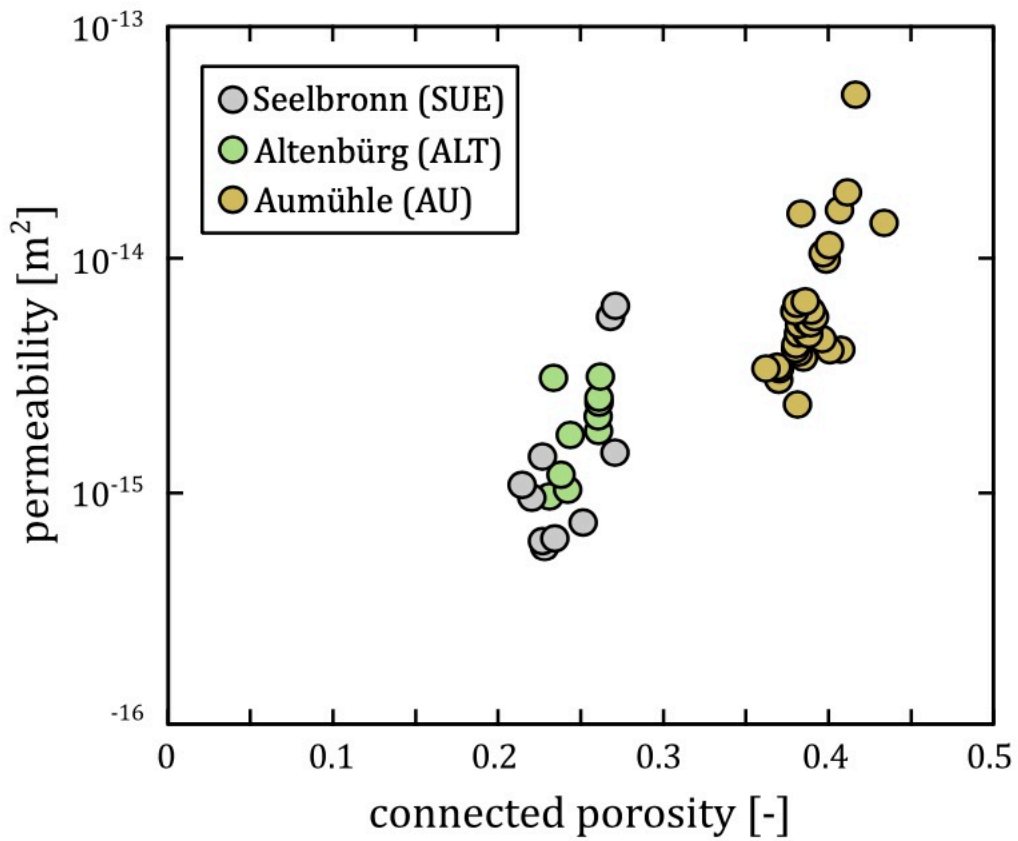
784 Aumühle (see Figure 1 for locations). Panels (a) and (b) are taken from Heap et al.

785 (2020a).



786

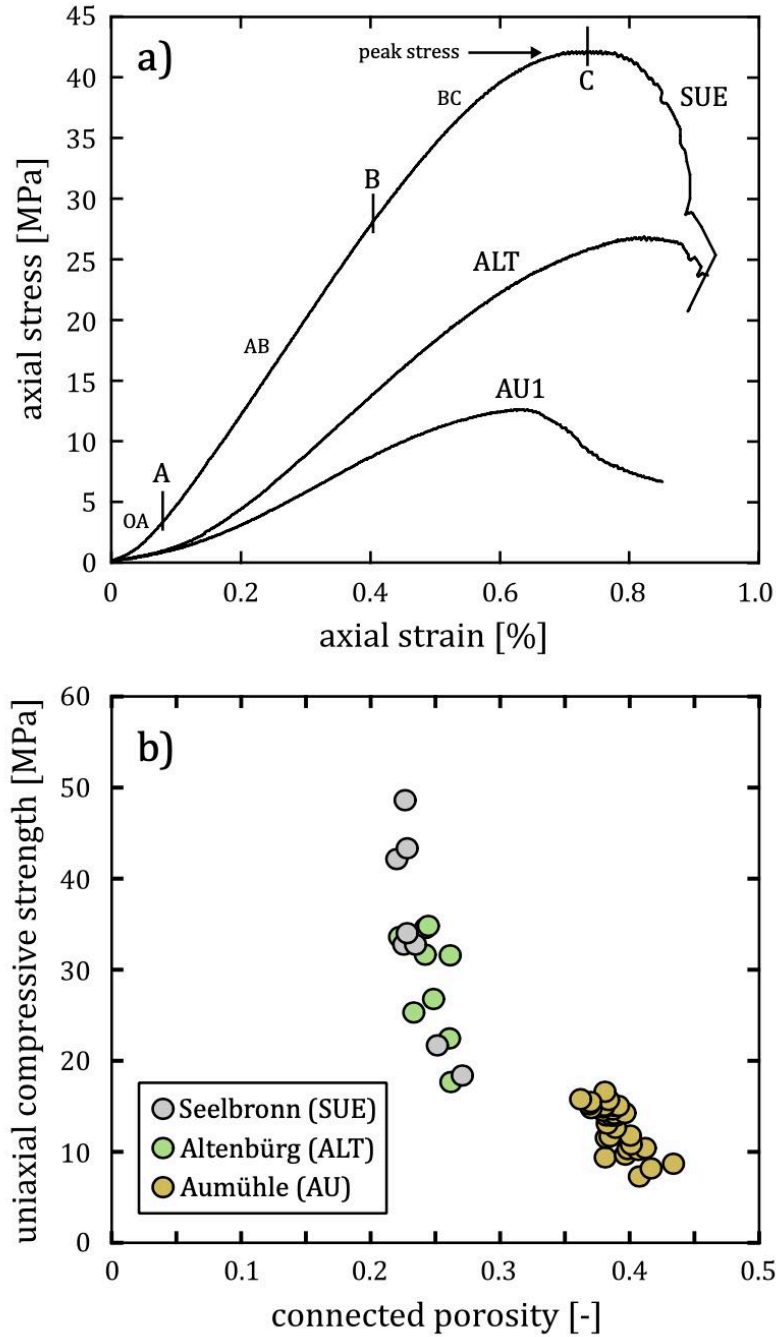
787 **Figure 4.** Cumulative void space as a function of the equivalent pore throat radius for  
 788 impact melt-bearing breccias (“suevites”) from each of the three quarry sites: Seelbronn  
 789 (SUE); Altenbürg (ALT); and Aumühle (AU1). (See Figure 1 for sample locations.)



790

791 **Figure 5.** Permeability as a function of connected porosity for impact melt-bearing  
 792 breccias (“suevites”) from each of the three quarry sites (see Figure 1 for locations).

793 “AU” includes samples from blocks AU1, AU2, and AU3 (see Table 2).



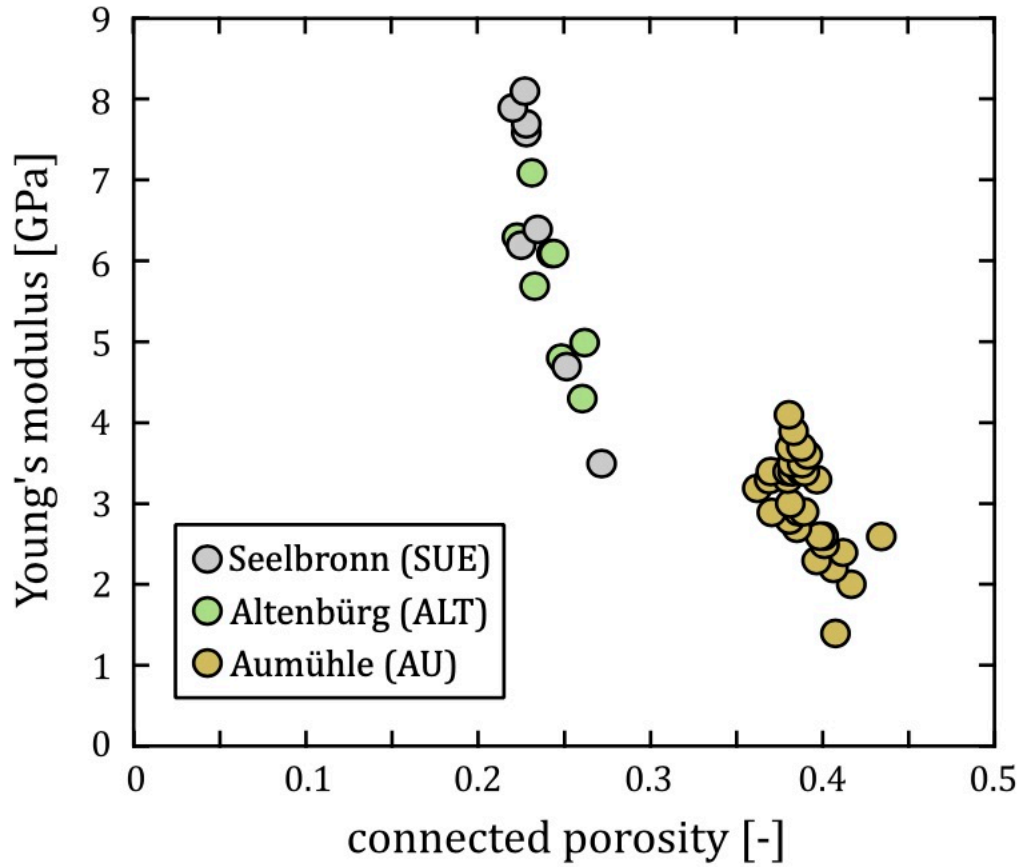
794

795 **Figure 6.** (a) Representative uniaxial stress–strain curves for impact melt-bearing  
 796 breccias (“suevites”) from each of the three quarry sites (see Figure 1 for locations).

797 The peak stress is labelled for the SUE curve. Sections OA, AB, and BC on the SUE  
 798 curve indicate the different parts of a typical stress–strain curve (see text for details).

799 (b) Uniaxial compressive strength as a function of connected porosity for these samples.

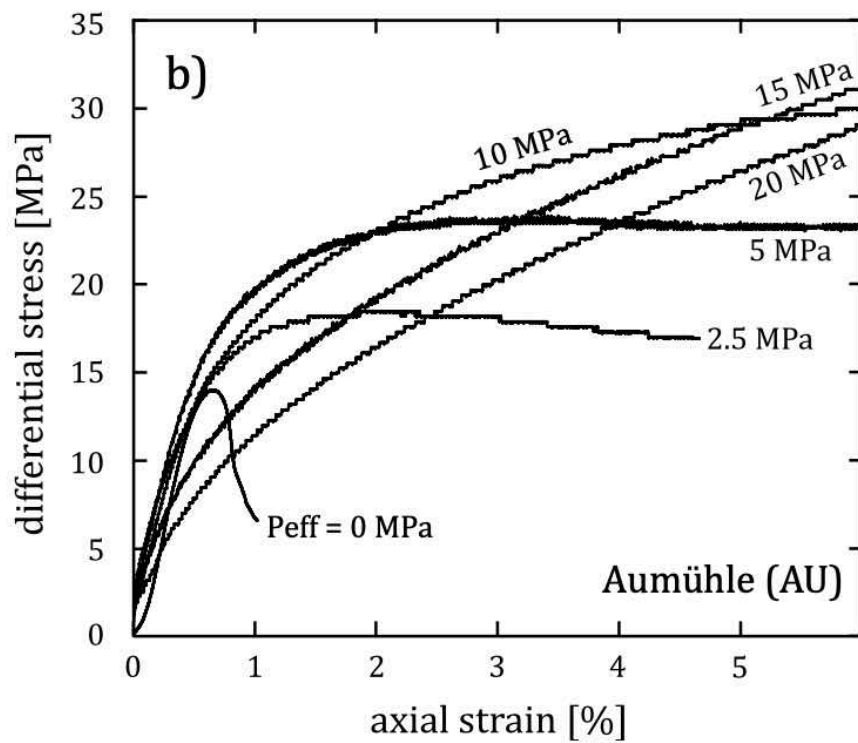
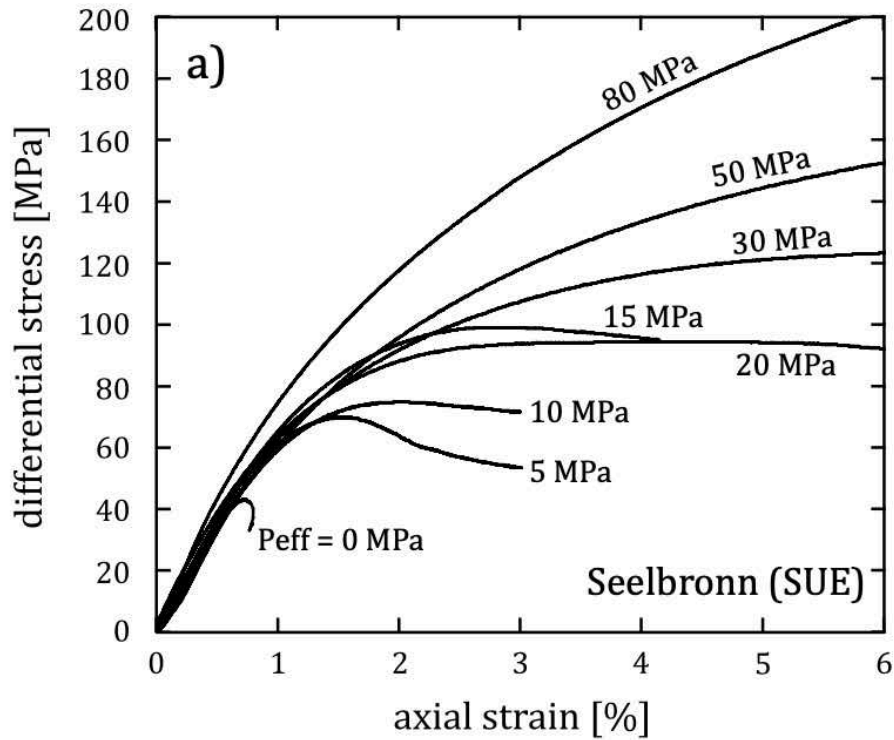
800 “AU” includes samples from blocks AU1, AU2, and AU3 (see Table 2).



801

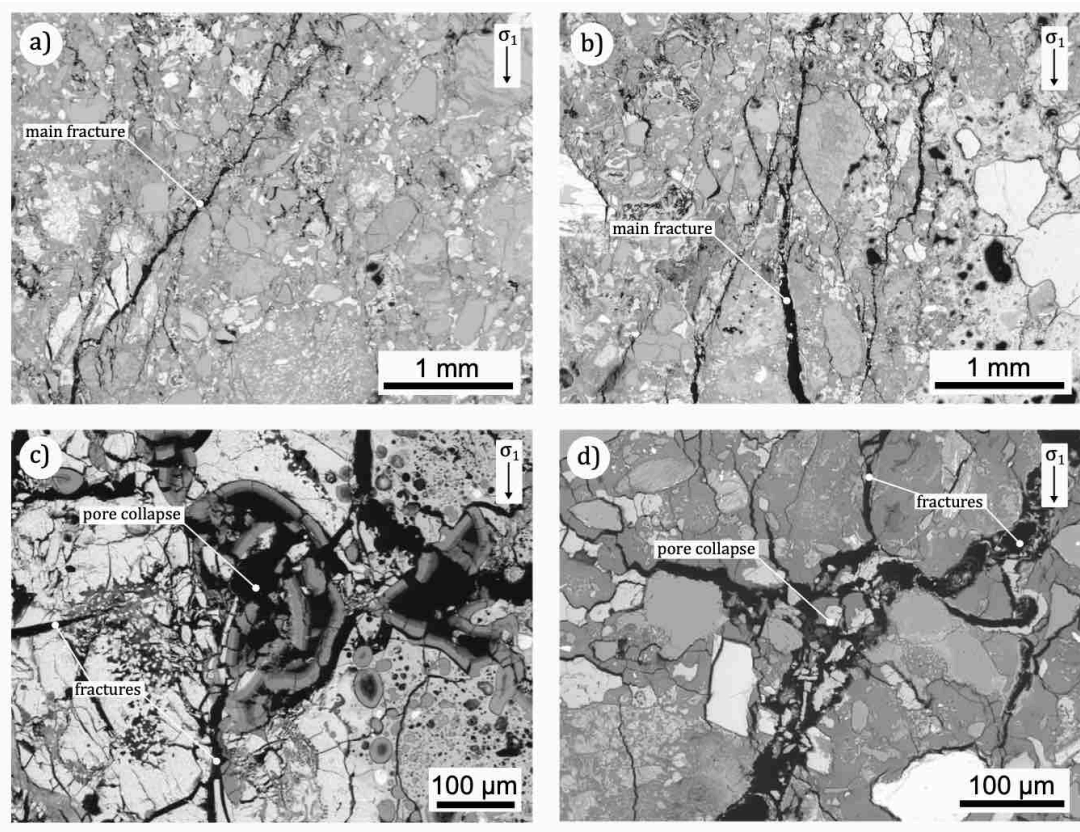
802 **Figure 7.** Static Young's modulus as a function of connected porosity for impact melt-  
 803 bearing breccias ("suevites") from each of the three quarry sites (see Figure 1 for  
 804 locations). "AU" includes samples from blocks AU1, AU2, and AU3 (see Table 2).





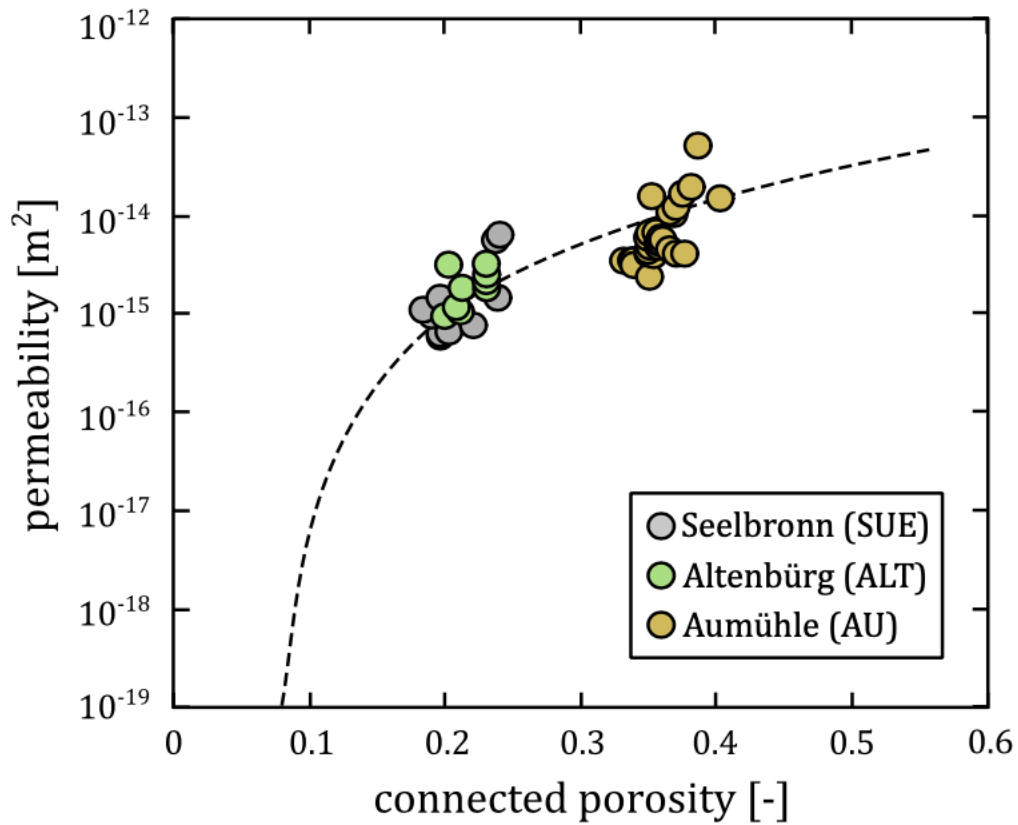
805

806 **Figure 8.** (a) Stress–strain curves for the impact melt-bearing breccias (“suevites”)   
 807 from Seelbronn (SUE) (see Figure 1 for location). (b) Stress–strain curves for the   
 808 samples from the Aumühle (AU1) quarry (see Figure 1 for location). In both plots, the   
 809 effective pressure ( $P_{\text{eff}}$ ) for each experiment is labelled next to the appropriate curve.



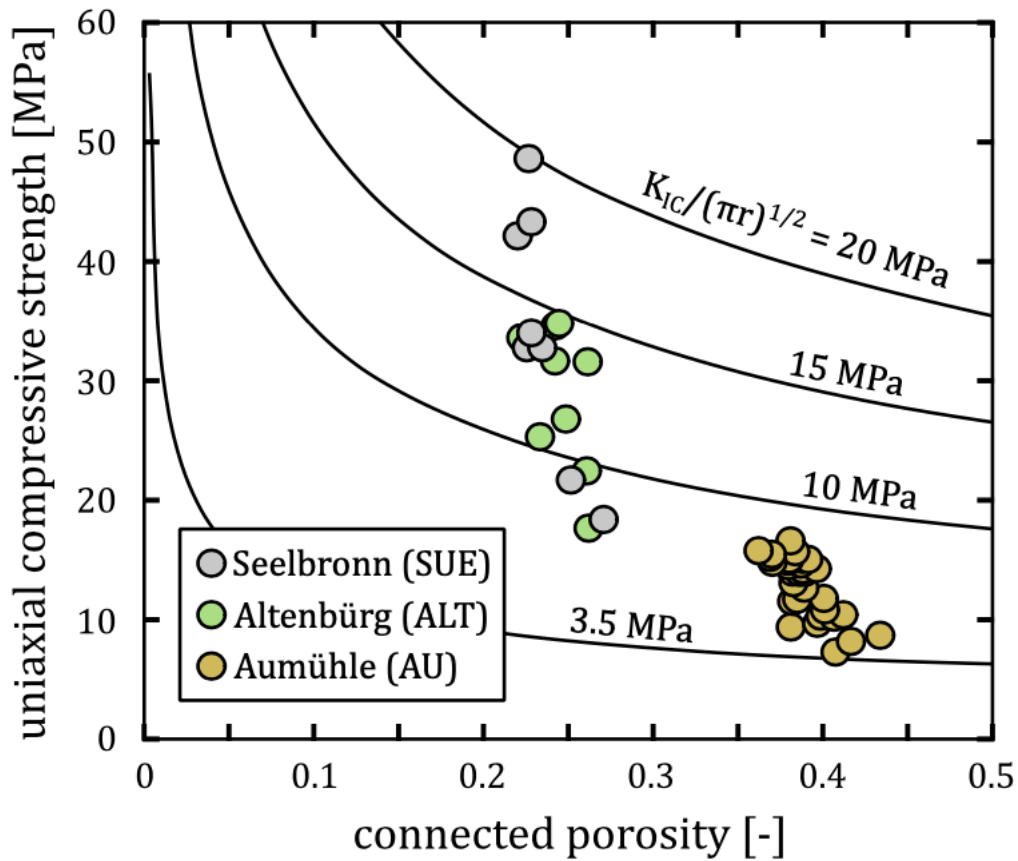
810

811 **Figure 9.** Backscattered electron microscope images of deformed samples of impact  
812 melt-bearing breccias (“suevites”) from the Seelbronn quarry (SUE) (see Figure 1 for  
813 location). Panels (a) and (b) show shear fractures (and accompanying microcracks)  
814 resulting from brittle failure. Panels (c) and (d) show the micromechanism driving  
815 porosity loss during the ductile experiments: cataclastic pore collapse (and  
816 accompanying microcracks).



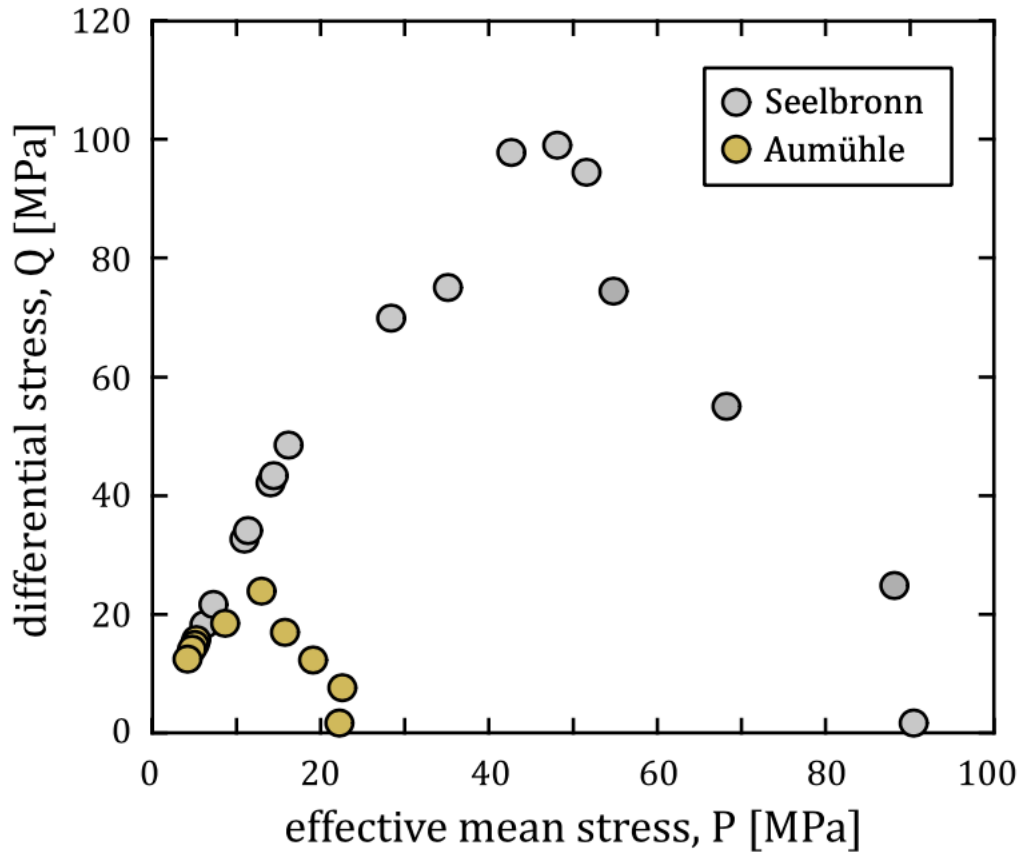
817

818 **Figure 10.** Permeability as a function of connected porosity for impact melt-bearing  
 819 breccias (“suevites”) from each of the three quarry sites (see Figure 1 for locations).  
 820 “AU” includes samples from blocks AU1, AU2, and AU3 (see Table 2). The dashed  
 821 line is the modelled curve (Equation (2)).



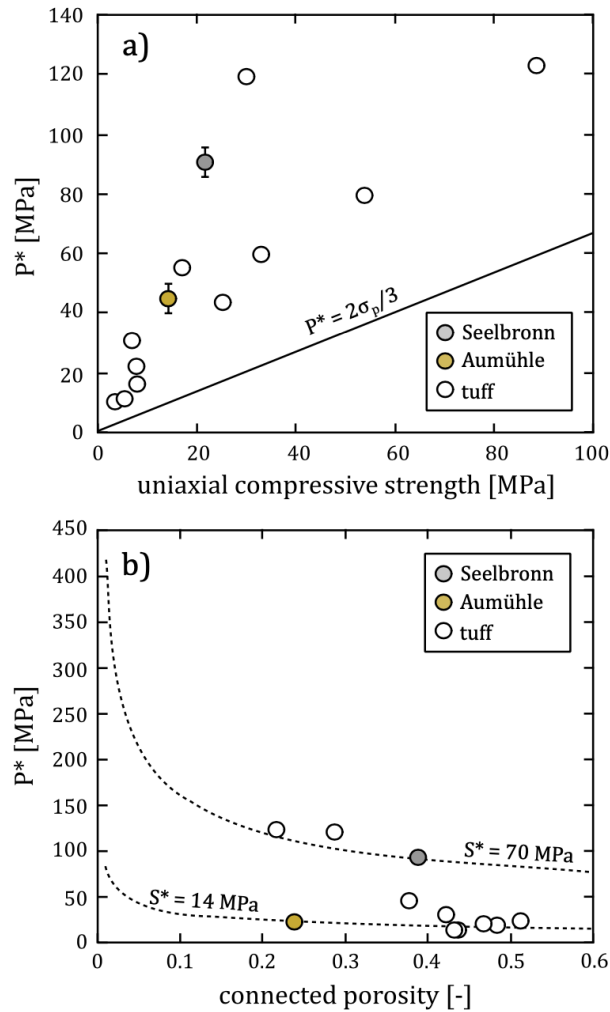
822

823 **Figure 11.** Uniaxial compressive strength as a function of connected porosity for  
 824 impact melt-bearing breccias (“suevites”) from each of the three quarry sites. “AU”  
 825 includes samples from blocks AU1, AU2, and AU3 (see Table 2). Modelled curves are  
 826 provided using the analytical approximation of Sammis and Ashby’s (1986) pore-  
 827 emanating crack model (Equation (3)).



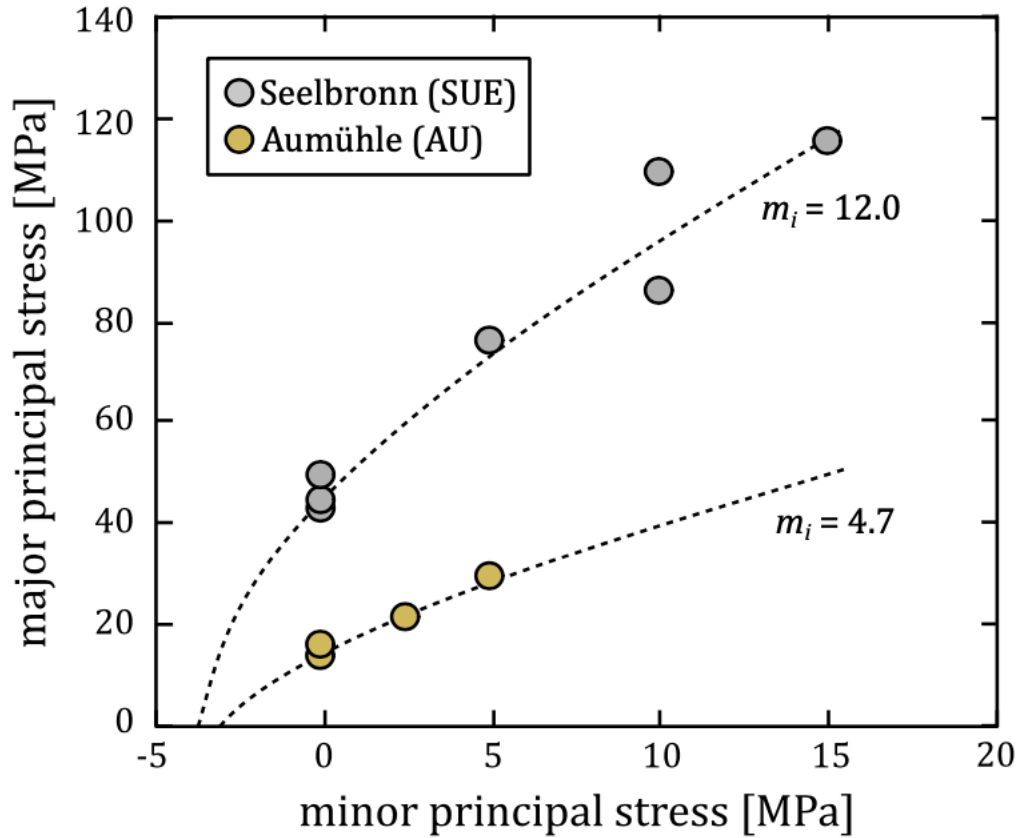
828

829 **Figure 12.** Differential stress,  $Q$ , (peak stress in the brittle regime and  $C^*$  in the ductile  
 830 regime) as a function of the effective mean stress,  $P$ , for impact melt-bearing breccias  
 831 (“suevites”) from Seelbronn (SUE) and Aumühle (AU1) (see Figure 1 for location)  
 832 (data given in Table 3).



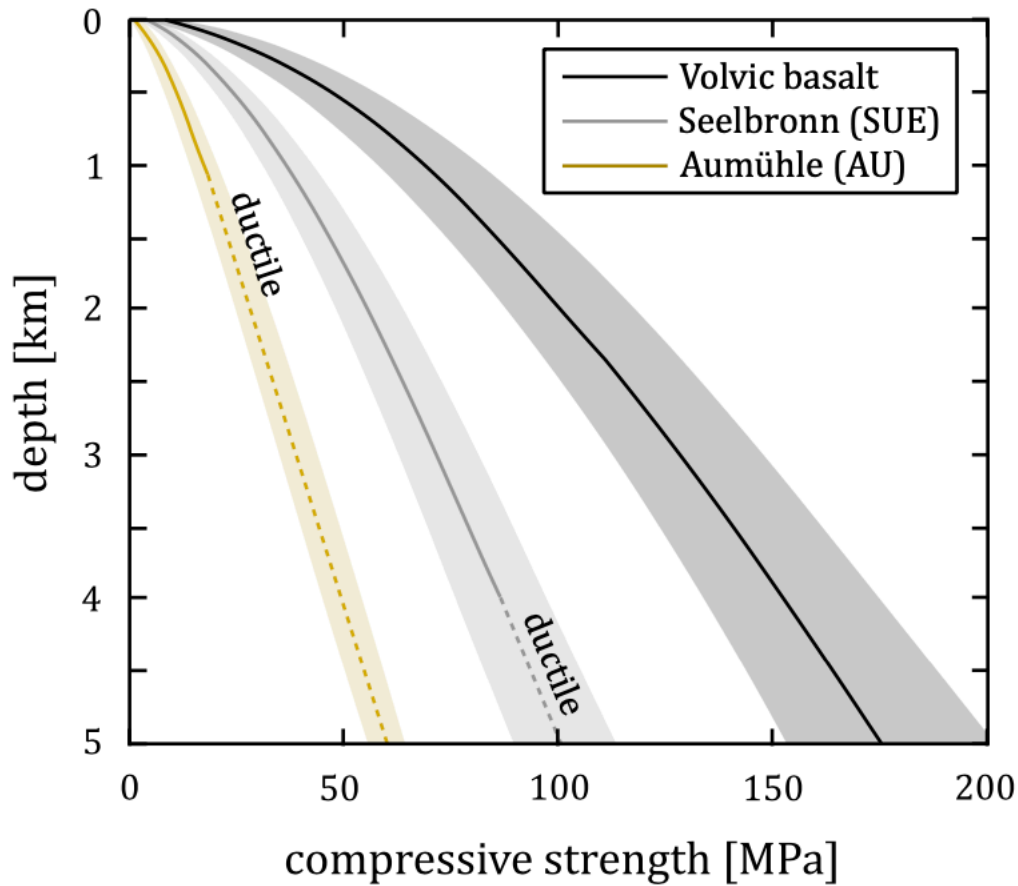
833

834 **Figure 13.** (a) The onset of inelastic hydrostatic compaction,  $P^*$ , as a function of the  
835 uniaxial compressive strength for the impact melt-bearing breccias (“suevites”) from  
836 Seelbronn (SUE) and Aumühle (AU1) (see Figure 1 for location). The position of  $P^*$   
837 for SUE and AU1 is estimated to be  $90 \pm 5$  and  $22 \pm 5$  MPa. The modelled curve is  
838 provided using Equation (4). (b) The onset of inelastic hydrostatic compaction,  $P^*$ , as  
839 a function of porosity for SUE and AU1 samples. The position of  $P^*$  for SUE and AU1  
840 is estimated to be  $90 \pm 5$  and  $22 \pm 5$  MPa (symbol size is larger than the error bar).  
841 Modelled curves are provided using Equation (5). On both panels we also provide data  
842 for tuff, a volcanic rock with a similar texture to the studied suevites (data from Heard  
843 et al., 1973; Aversa and Evangelista, 1998; Lockner and Morrow, 2008; Zhu et al.,  
844 2011; Heap et al., 2014b; Marmoni et al., 2017; Heap et al., 2018).



845

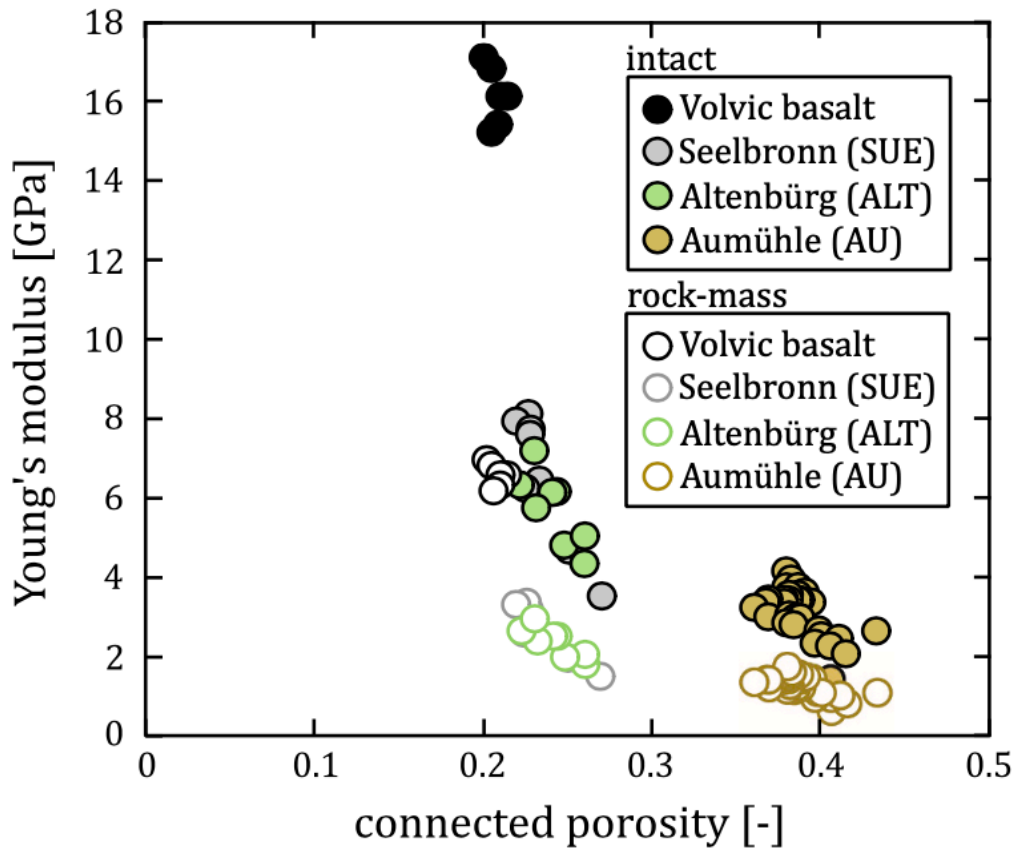
846 **Figure 14.** Data from triaxial experiments performed on samples of impact melt-  
 847 bearing breccias (“suevites”) from Seelbronn (SUE) and Aumühle (AU1) (see Figure  
 848 1 for location) plotted in principal stress space (major principal stress as a function of  
 849 minor principal stress). Best-fit modelled curves are provided using the data fitting  
 850 function (modified cuckoo fit algorithm, basic error summation, and absolute error  
 851 type) in RocData (Rocscience, 2017). The best-fit  $m_i$  values are given on the figure.



852

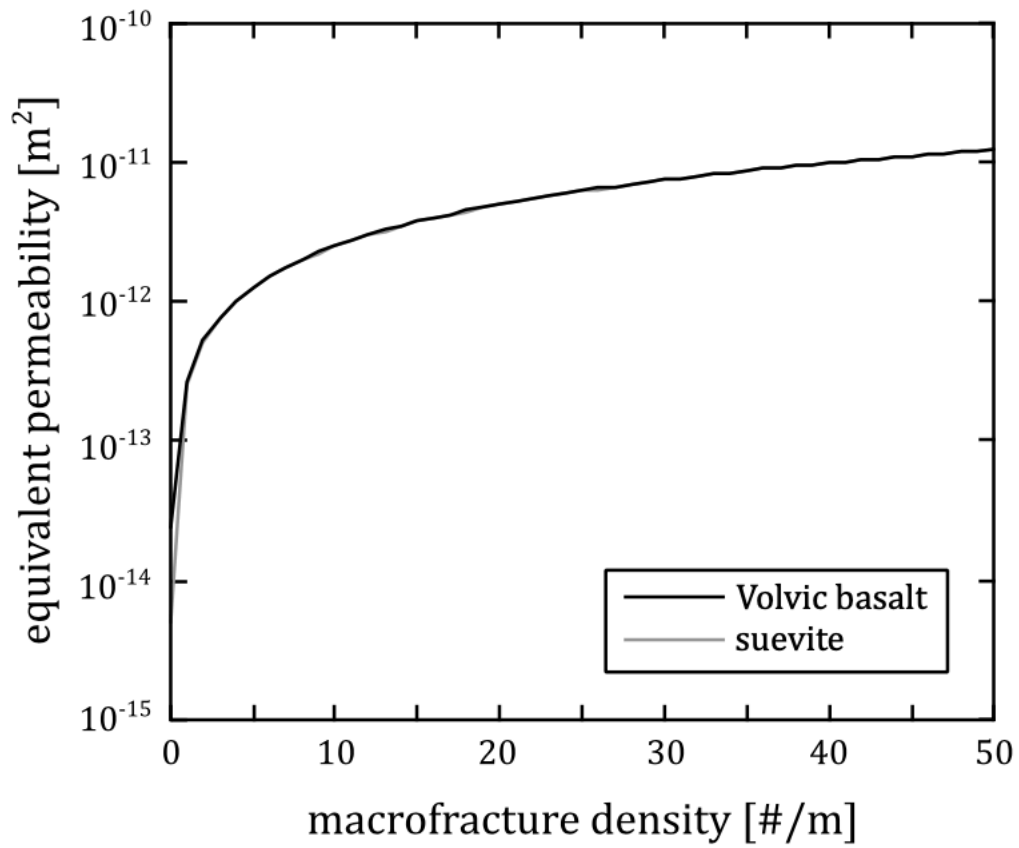
853 **Figure 15.** Crustal strength profiles (for Mars) for impact melt-bearing breccias  
 854 (“suevites”) from Seelbronn (SUE) and Aumühle (AU1), (see Figure 1 for location)  
 855 and for Volvic basalt (Chaîne des Puys, France), calculated with Equation (6). Dashed  
 856 lines indicate the depth range for ductile deformation, as informed by our experimental  
 857 data (Table 3); strength values provided by Equation (6) are not valid for this depth  
 858 range. Solid/dashed lines are for a Geological Strength Index (GSI) of 55 and the  
 859 shaded zones show strength profiles for  $GSI = 55 \pm 10$  (see text for details).





860

861 **Figure 16.** “Intact” (coloured symbols) and rock-mass (white symbols) Young’s  
 862 modulus as a function of connected porosity for impact melt-bearing breccias  
 863 (“suevites”) from Seelbronn (SUE), Altenbürg (ALT), and Aumühle (AU1) (see Figure  
 864 1 for locations) and for Volvic basalt (Chaîne des Puys, France). Rock-mass Young’s  
 865 moduli are calculated using Equation (10) and using a GSI of 55.



866

867 **Figure 17.** Equivalent permeability as a function of macrofracture density for impact

868 melt-bearing breccia (“suevite”), calculated using Equation (11).

869 **Table 1.** Quantitative bulk mineralogical composition, determined using X-ray powder  
870 diffraction (XRPD), for the five impact melt-bearing breccias (“suevite”) blocks used  
871 in this study. The error associated with the estimation of the content of amorphous  
872 phases is a conservative estimate. “n.d.”: not detected.

873

<b>Mineral</b>	<b>Altenbürg (ALT) (values in wt.%)</b>	<b>Seelbronn (SUE) (values in wt.%)</b>	<b>Aumühle 1 (AU1) (values in wt.%)</b>	<b>Aumühle 2 (AU2) (values in wt.%)</b>	<b>Aumühle 3 (AU3) (values in wt.%)</b>
Quartz	9 ± 1	10 ± 1	11 ± 1	11 ± 1	11 ± 1
Coesite	1.7 ± 0.3	1.4 ± 0.3	0.7 ± 0.2	0.8 ± 0.2	0.9 ± 0.2
K-feldspar	3.1 ± 0.5	1.7 ± 0.5	4.5 ± 0.8	3.5 ± 0.5	4.3 ± 0.4
Plagioclase	18 ± 2	27 ± 3	17 ± 2	16 ± 2	15 ± 2
Calcite	13 ± 1	2.1 ± 0.2	2.6 ± 1.2	1.1 ± 1.2	0.6 ± 0.3
Smectite	25 ± 3	23 ± 3	31 ± 4	34 ± 4	37 ± 4
Amphibole	1.7 ± 0.3	n.d.	n.d.	n.d.	n.d.
Biotite	2.0 ± 0.5	1.9 ± 0.5	1.8 ± 0.5	1.8 ± 0.5	1.9 ± 0.5
Hematite/Goethite	0.2 ± 0.1	0.2 ± 0.1	0.4 ± 0.2	0.4 ± 0.2	0.2 ± 0.2
Amorphous phases	28 ± 5	33 ± 5	32 ± 8	32 ± 8	30 ± 8

874

875

876 **Table 2.** The physical and mechanical properties (porosity, permeability, Young’s  
877 modulus, and uniaxial compressive strength) of dry impact melt-bearing breccia  
878 (“suevite”) collected from the Ries impact crater, Germany. Asterisk denotes data from  
879 Heap et al. (2020a).  
880

Sample	Connected porosity	Permeability (m <sup>2</sup> )	Young’s modulus (GPa)	Uniaxial compressive strength (MPa)
SUE1*	0.241	$6.33 \times 10^{-15}$	3.5	18.4
SUE2*	0.197	$6.22 \times 10^{-16}$	8.1	48.6
SUE3*	0.238	$5.68 \times 10^{-15}$	-	-
SUE4*	0.221	$7.40 \times 10^{-16}$	4.7	21.7
SUE5*	0.190	$9.53 \times 10^{-16}$	7.9	42.2
SUE6*	0.184	$1.08 \times 10^{-15}$	-	-
SUE7*	0.197	$1.42 \times 10^{-15}$	-	-
SUE8*	0.204	$6.32 \times 10^{-16}$	6.4	32.8
SUE9*	0.240	$1.47 \times 10^{-15}$	-	-
SUE10*	0.198	$5.84 \times 10^{-16}$	7.7	43.3
SUE12*	0.195	-	6.2	32.8
SUE20*	0.228	-	7.6	34.1
ALT1*	0.242	$1.02 \times 10^{-15}$	6.1	31.7
ALT2*	0.261	$2.54 \times 10^{-15}$	-	31.6
ALT3*	0.261	$2.44 \times 10^{-15}$	-	-
ALT4*	0.233	$3.13 \times 10^{-15}$	5.7	25.3
ALT5*	0.243	$1.76 \times 10^{-15}$	-	-
ALT6*	0.262	$3.16 \times 10^{-15}$	5.0	17.7
ALT7*	0.238	$1.19 \times 10^{-15}$	-	-
ALT8*	0.261	$1.84 \times 10^{-15}$	4.3	22.5
ALT9*	0.261	$2.13 \times 10^{-15}$	-	-
ALT10*	0.231	$9.54 \times 10^{-16}$	7.1	33.0
ALT12*	0.245	-	6.1	34.8
ALT13*	0.223	-	6.3	33.6
ALT14*	0.242	-	6.1	34.6
ALT15*	0.248	-	4.8	26.9
AU1_1	0.387	$4.78 \times 10^{-15}$	3.4	14.7
AU1_2	0.392	$5.66 \times 10^{-15}$	3.6	15.0
AU1_3	0.388	$5.38 \times 10^{-15}$	3.7	15.0
AU1_4	0.384	$1.57 \times 10^{-14}$	3.9	15.7
AU1_5	0.386	$6.59 \times 10^{-15}$	2.9	14.0
AU1_6	0.380	$4.29 \times 10^{-15}$	4.1	15.0
AU1_7	0.396	$4.53 \times 10^{-15}$	3.3	14.3

AU1_8	0.389	$5.30 \times 10^{-15}$	2.9	12.6
AU1_9	0.390	$6.02 \times 10^{-15}$	3.4	14.0
AU1_10	0.380	$5.97 \times 10^{-15}$	3.3	14.8
AU1_11	0.388	$4.77 \times 10^{-15}$	3.5	14.2
AU2_1	0.383	$5.24 \times 10^{-15}$	3.4	13.1
AU2_2	0.381	$2.37 \times 10^{-15}$	3.7	16.6
AU2_3	0.370	$3.05 \times 10^{-15}$	3.4	15.5
AU2_4	0.408	$4.10 \times 10^{-15}$	1.4	7.3
AU2_5	0.401	$4.08 \times 10^{-15}$	2.5	10.8
AU2_6	0.382	$4.00 \times 10^{-15}$	3.5	14.0
AU2_7	0.380	$4.11 \times 10^{-15}$	3.4	14.5
AU2_8	0.381	$6.43 \times 10^{-15}$	2.8	9.4
AU2_9	0.369	$3.46 \times 10^{-15}$	3.3	15.2
AU2_10	0.385	$3.79 \times 10^{-15}$	2.7	11.7
AU2_11	0.370	$3.35 \times 10^{-15}$	2.9	14.8
AU2_12	0.362	$3.40 \times 10^{-15}$	3.2	15.8
AU3_1	0.397	$1.06 \times 10^{-14}$	2.3	9.7
AU3_2	0.417	$5.10 \times 10^{-14}$	2.0	8.2
AU3_3	0.399	$9.96 \times 10^{-15}$	2.6	10.3
AU3_4	0.434	$1.44 \times 10^{-14}$	2.6	8.7
AU3_5	0.382	$4.78 \times 10^{-15}$	3.0	11.5
AU3_6	0.412	$1.94 \times 10^{-14}$	2.4	10.4
AU3_7	0.407	$1.63 \times 10^{-14}$	2.2	10.2
AU3_8	0.401	$1.15 \times 10^{-14}$	2.6	11.8

881

882

883 **Table 3.** Mechanical data from the triaxial deformation experiments on dry impact  
 884 melt-bearing breccia (“suevite”) performed for this study. The asterisk denotes samples  
 885 that were held at a differential stress of 2 MPa and the confining pressure was increased  
 886 until C\* was reached (see methods section for more details).

887

Sample	Connected porosity	Confining pressure (MPa)	Pore fluid pressure (MPa)	Effective pressure (MPa)	Peak differential stress, Q (MPa)	C* (MPa)	Effective mean stress, P (MPa)
SS2	0.252	5	0	5	69.9	-	28.3
SS3	0.242	10	0	10	75.1	-	35.0
SS4	0.255	15	0	15	99.2	-	48.1
SS6	0.246	20	0	20	94.6	-	51.5
SS7	0.251	30	0	30	-	74.5	54.8
SS8	0.252	50	0	50	-	54.9	68.3
SS10*	0.247	90	0	90	-	2.0	90.7
SS11	0.234	80	0	80	-	24.7	88.2
SS12	0.237	10	0	10	97.9	-	41.6
AU1_1	0.376	5	0	5	23.9	-	13.0
AU1_3	0.383	10	0	10	-	17.1	15.7
AU1_7	0.385	15	0	15	-	12.2	19.1
AU1_11	0.385	20	0	20	-	7.6	22.5
AU1_12	0.377	2.5	0	2.5	18.5	-	8.7
AU1_16*	0.384	22	0	22	-	2.0	22.7

888

889

890 **Table 4.** The physical and mechanical properties (porosity, permeability, Young's  
891 modulus, and uniaxial compressive strength) of dry Volvic basalt (Chaîne des Puys,  
892 France). Asterisk denotes data from Heap et al. (2020b).

893

<b>Sample</b>	<b>Connected porosity</b>	<b>Permeability (m<sup>2</sup>)</b>	<b>Young's modulus (GPa)</b>	<b>Uniaxial compressive strength (MPa)</b>
VB2	0.20	$2.35 \times 10^{-14}$	17.1*	94.6
VB6	0.21	-	15.4*	96.5
VB8	0.20	-	16.8*	84.2
VB9	0.21	-	15.2*	91.1
VB10	0.21	-	16.1*	94.8
VB12	0.21	-	16.1*	94.7

894

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