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► **To cite this version:**

Marion Baques, Louis de Barros, Maxime Godano, Clara Duverger, Hervé Jomard. Swarms and mainshock-aftershocks sequences are both triggered by fluids in the Ubaye Region (Western Alps). *Geophysical Journal International*, 2023, 235 (1), pp.920-941. 10.1093/gji/ggad280 . hal-04182990

HAL Id: hal-04182990

<https://hal.science/hal-04182990v1>

Submitted on 18 Aug 2023

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1 **Swarms and mainshock-aftershocks sequences are both triggered**
2 **by fluids in the Ubaye Region (Western Alps).**

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9 **Abbreviated title:**

10 Fluid-triggered mainshock-aftershocks and swarms

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21 **Abstract**

22 The Ubaye Region (French Western Alps) is one of the most seismically active regions in France. It is
23 regularly struck by mainshock-aftershocks sequences like in 1959 (ML5.3), seismic swarms (2003-
24 2004), and complex sequences (2012-2015) characterized by successive mainshocks clustered in time
25 and space. This diversity of seismic behaviour highlights the complex processes at play in this area.
26 To improve our understanding of these processes, in this study, we compile a regional catalogue of
27 existing focal mechanisms, completed by 100 new calculated focal mechanisms of aftershocks
28 following the 07/04/2014 mainshock (ML5.1). The oriented stress-state we reconstruct for different
29 periods and sub-areas are similar to each other and to previous published values focusing on swarm
30 periods. We then calculate fluid-pressure required to trigger the earthquakes. Most of the events
31 (65%) need fluid-overpressure between 15 and 40 MPa (17-to-40% of the hydrostatic pressure) with
32 a median value of 24%. Moreover, even the largest events, like the mainshocks in the 2012-2015
33 sequence, appear to be triggered by fluid-pressure, similarly as events within swarm sequences.
34 However, while fluid-overpressure decreases with time in an aftershock sequence, it varies randomly
35 at high levels during a swarm sequence. Therefore, based on a fault-valve model, we propose that: 1)
36 the fluids trapped in the fault plane tend toward lithostatic pressure and trigger the mainshock
37 rupture and 2) part of the aftershocks are induced by the diffusing fluid-pressure. On the contrary,
38 swarms need external, likely deep, fluid-pressure feedings. Fluid-pressure is likely to be a common
39 triggering mechanism of the seismicity in the Ubaye Region, even if the involved processes should
40 differ to explain the different types of seismic sequences.

41 **Keywords**

42 Europe; Earthquake dynamics; Seismicity and tectonics; Dynamics and mechanisms of faulting; Fluid-
43 driven seismicity

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47 **1. Introduction**

48 Two end-member behaviours may characterize earthquake sequences: mainshock-aftershocks
49 sequences and swarm sequences. On one hand, mainshock-aftershocks sequence is the most
50 common behaviour and starts with the largest magnitude earthquake, possibly damaging, followed
51 by aftershocks of lower magnitude. Aftershocks are mainly driven by coseismic stress transfer (Stein,
52 1999) and occur around the fault zone where the mainshock occurred. However, some of the
53 aftershocks sequences may also have been triggered by fluid-pressure, reactivating fault by high
54 fluid-pressure (e.g. Gunatilake, 2023). The number of aftershocks decays with time following an
55 Omori-Utsu's law (Utsu and Ogata, 1995). Such mainshock-aftershocks sequences occur in all
56 tectonically active contexts, as the tectonic stress loading directly triggers them. On the other hand,
57 swarm sequences are bursts of earthquakes, usually of small-to-moderate magnitudes, which are
58 clustered in time and space. They are characterized by a slow increase and decrease of the seismicity
59 rate without a clear pattern, and by the strongest event occurring at any time during the sequence.
60 They also occur in different tectonic settings: subduction zones (e.g. Japan, Kato *et al.* 2014;
61 Nishikawa and Ide, 2017), volcanic regions (e.g. Mayotte, Bertil *et al.* 2021) or stable intraplate areas
62 (e.g. Western Bohemia, Fischer *et al.* 2014). They can last for days (e.g. Mammoth Mountain, Shelly
63 and Hill, 2011; Mount Rainier, Shelly *et al.* 2013; Corinth Gulf, De Barros *et al.* 2020; South Korea, Son
64 *et al.* 2021) to years (e.g. Ubaye, Jenatton *et al.* 2007; Cahuilla swarm, Ross and Cochran, 2021;
65 Japan, Amezawa *et al.* 2021). To maintain this activity, a driving mechanism that lasts in time is
66 required. As swarms can also be induced by geothermal activities (e.g. Baisch *et al.*, 2015; Schoenball
67 *et al.*, 2014) or hydrocarbon resources exploitations (e.g. Goebel *et al.* 2016; Schultz *et al.* 2020) or
68 even triggered by rainfalls during monsoon period (e.g. Gunatilake *et al.*, 2021), they are commonly

69 associated with fluid-pressure diffusion at depth (e.g. Miller *et al.* 2004; Parotidis and Shapiro, 2004;
70 Terakawa *et al.* 2010). Classically, seismic failures are seen to be induced directly by the decrease of
71 normal effective stress associated with the increase of fluid-pressure. Seismic migration is then
72 explained by the fluid-pressure diffusion. Alternatively, swarm sequences were also observed to be
73 driven by slow-slips or aseismic deformation (Lohman and McGuire, 2007; Vallée *et al.* 2013). Fluid
74 and aseismic processes may also be imbricated to drive swarms as the first one drives the latter, as
75 observed in different contexts (Corinth Gulf, De Barros *et al.* 2020; Canada, Eyre *et al.* 2020) or
76 numerical simulations (e.g. Wynants-Morel *et al.* 2020). As no direct observations are possible in the
77 source area, swarm processes are still an open question.

78 In between these two end-member categories, more complex sequences exist. For example, some
79 sequences involve several cascading mainshocks, with a global swarm pattern (Colfiorito sequence,
80 Chiaraluce *et al.* 2003; Ubaye region, Thouvenot *et al.* 2016; Western Bohemia, Hainzl *et al.* 2016,
81 Jakoubková *et al.* 2018; Corinth Gulf, Kaviris *et al.* 2021). In the Ubaye Region (De Barros *et al.* 2019)
82 as well as in the Puerto Rico area (Ventura-Valentin and Brudzinski, 2021) or the Corinth Gulf (Kaviris
83 *et al.* 2021), authors observe that a mainshock started a seismic crisis, but with an abnormal number
84 of aftershocks that did not decay according to the expected Omori-Utsu's law and contained small
85 swarm sequences. In such cases, complex processes involving tectonic stress release as well as fluids
86 and/or aseismic-slip are required. The Amatrice-Visso-Norcia sequence (Appenines, Italy) was also a
87 complex seismic sequence during which the generation of internal fluids by decarbonisation seems
88 to be responsible for non-Omori type aftershock behaviour (Gunatilake and Miller, 2022).

89

90 The Ubaye Region, located in the Southwestern Alps, is one of the most seismically active area in
91 France. Mainshock-aftershocks, seismic swarms, and complex sequences have been observed since
92 the 18th century (Baques *et al.* 2021). In 2003-2004, more than 16,000 events were recorded in a
93 swarm sequence (Jenatton *et al.* 2007; Daniel *et al.* 2011; Leclère *et al.* 2013). The last seismic crisis
94 happened in 2012-2015 with two moderate earthquakes of magnitude ML4.8 and ML5.1, which

95 occurred at a similar location on 26/02/2012 and 07/04/2014, respectively. They were followed by
96 an abnormally high rate of seismicity (Thouvenot *et al.* 2016; De Barros *et al.* 2019), including several
97 small swarms at close distances from the mainshocks. Globally, this seismic crisis can be described as
98 a swarm, because the size of the seismic cluster is much larger than the spatial extent of the main
99 ruptures. As 1) the tectonic deformation is very limited in the Southern Alps (e.g. Mathey *et al.*
100 2021b; Walpersdorf *et al.* 2018), and 2) the areas just South or North of the Ubaye Valley are less
101 active seismically, complex processes are required to explain both the high rate of seismicity and its
102 complex patterns. For both the 2003-2004 swarm and the 2012-2015 sequence, fluid processes have
103 been raised (e.g. Leclère *et al.* 2013; Thouvenot *et al.* 2016; De Barros *et al.* 2019). For example,
104 seismic migration with a diffusivity of 0.002 to 0.06 m²/s were observed for the 2003-2004 and
105 within the 2012-2015 sequences (Thouvenot *et al.* 2016; De Barros *et al.* 2019). These values are
106 similar to the measured ones for fluid-induced seismicity in reservoir exploitations (Shapiro *et al.*
107 2002) and for other swarms worldwide (Talwani *et al.* 2007). Leclère *et al.* (2013) proposed that
108 creep compaction pressurized the fluid within the fault, which then drove the 2003-2004 swarm
109 migration. Daniel *et al.* (2011) also proposed that fluid process is required, because stress-transfer
110 through earthquake-earthquake interaction explains only 59% of the 2003-2004 swarm. De Barros *et*
111 *al.* (2019) suggested an interplay between stress-transfer and fluid-driven process since only some
112 clusters in the 2014 aftershocks followed a typical aftershocks pattern, while others clearly behaved
113 as seismic swarms.

114

115 In this study, we aim at better understanding the processes at play, focusing both on the 2014
116 aftershocks period that was recorded by a dense temporary network (e.g. Baques *et al.*, 2021) and on
117 specific seismic sequences, such as the 2003-2004 seismic swarm and the 2012-2015 seismic crisis, to
118 explain the peculiar seismic behaviour in the Ubaye Region.

119

120 Direct observations of fluid are not possible at crustal depth. However, as fluid-pressure induces
121 failures through reduction of the effective stress, it may trigger seismicity on fault planes that are
122 misoriented with respect to the tectonic stress-state (Sibson, 1990). We therefore aim to compare
123 the orientation of the observed failures, obtained through focal mechanisms, with the average
124 stress-state. Using Cauchy's equations, we can reconstruct the fluid-pressure that is required within
125 the faults to induce earthquakes. A similar approach was used by Leclère et al. (2013), but focusing
126 only on the 2003-2004 swarm period. They showed that the seismicity required from 20 to 50 MPa of
127 overpressure with ~35 MPa at the beginning of the swarm and increasing up to ~55 MPa during the
128 peak of activity.

129 After presenting the general seismo-tectonic context of the area, we first describe the different focal
130 mechanisms catalogues we use for the Ubaye Region, including the well-instrumented two-month
131 period after the 2014 mainshock. We then present the methods used to reconstruct the stress-state
132 and the fluid-pressure needed to trigger the seismicity. We apply these methods on the Ubaye
133 Region and refine the results looking at sub-areas, for different time periods, and for the 2014
134 aftershocks. Finally, we propose and discuss a conceptual model that explains the role of fluid-driven
135 process in the Ubaye Valley area.

136

137

138

139 **2. Seismo-tectonic context**

140 **2.1. Tectonic context**

141 The Alpine chain is divided into two zones, internal and external, delimited by the Frontal Penninic
142 Thrust (FPT, Fig. 1). East of the FPT, in the intern zone, the Ubaye Region is mainly composed of the

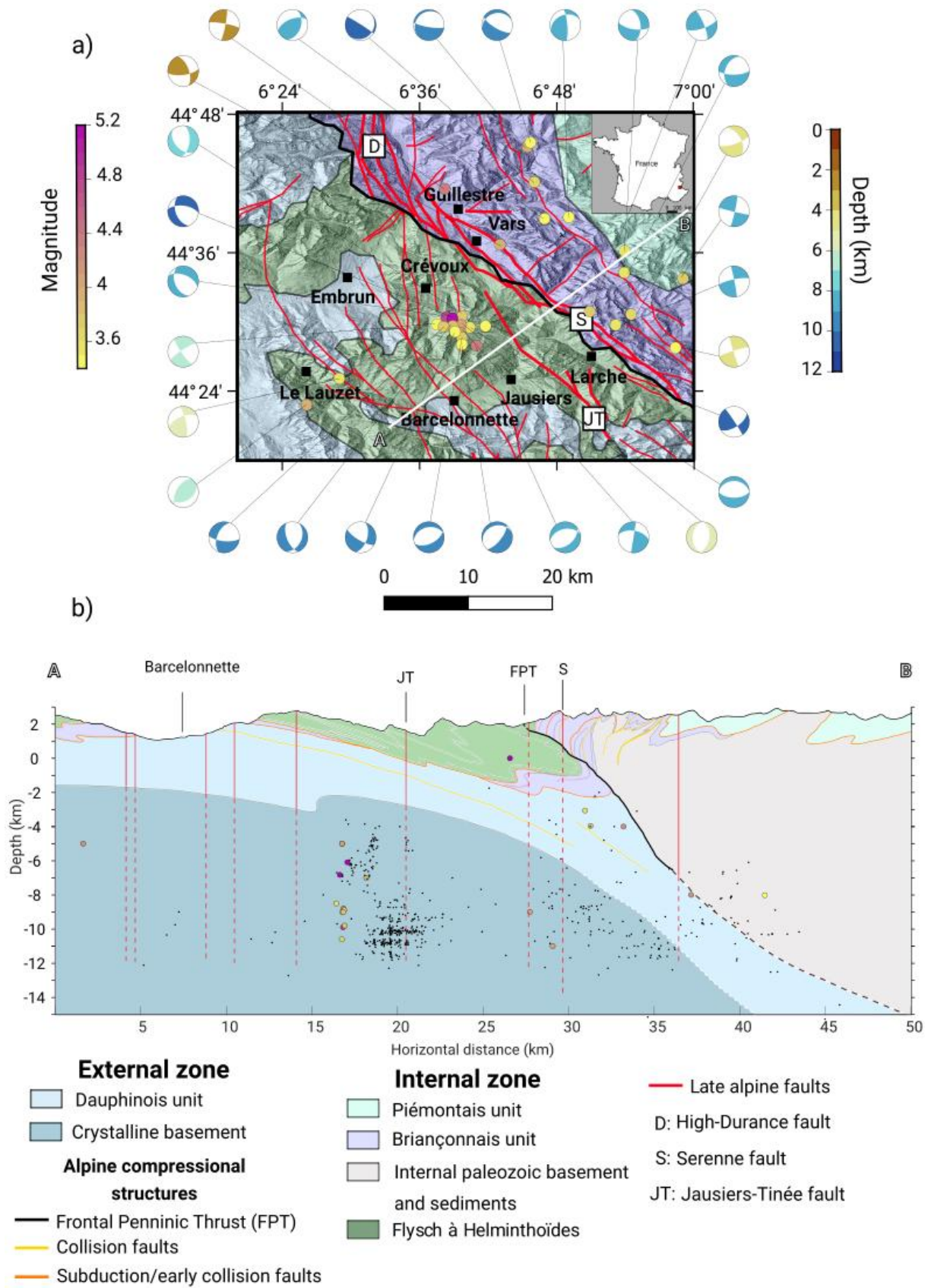
143 Briançonnais unit, which consists of a pile of thick low grade metamorphic nappes (e.g. Sue *et al.*
144 2007). West of the FPT, the Dauphinois units is composed of a 1-to-2-km thick Mesozoic to Cenozoic
145 sedimentary rock cover (sandstone, marls, and limestone, Kerckhove *et al.* 1978) laying on the
146 crystalline basement (granitoid; Baietto *et al.*, 2009). Overlaying some parts of the Dauphinois units is
147 the 1-to-2 km thick “Flysch à Helminthoïdes”. This cover is also called the Embrunais-Ubaye nappes
148 (Kerckhove, 1969) and is composed of late Cretaceous turbiditic series that were thrust westward
149 during the Oligocene period (e.g. Gratier *et al.* 1989).

150 Crossing perpendicularly the Alpine structures, the Ubaye Valley gives us a view of the long
151 geological history of compression and extension resulting from the setting of the Western Alps. The
152 late Alpine faults as defined by Kerckhove (1969) and Tricart (2004) are the more recently recorded
153 deformations. It consists in a series of normal and right lateral strike-slip faults formed or reactivated
154 during the last Mio-Pliocene setting of the chain, and showing field evidence of neotectonic activity
155 (Sue, 1998; Sue *et al.* 2007). Current seismicity seems to develop along them (Jenatton *et al.* 2007;
156 Sue *et al.* 2007; Le Goff *et al.* 2009; Sanchez *et al.* 2010; Mathey *et al.* 2020; Fig. 1). A majority of
157 these faults are oriented N-S to NW-SE with some minor conjugate structures oriented E-W to NE-SW
158 (Kerckhove, 1969; Sue and Tricart, 2003; Tricart and Schwartz, 2006). The main identified active
159 faults in the Ubaye Region are the High-Durance, the Serenne, and the Jausiers-Tinée faults (Fig. 1;
160 Baques *et al.* 2021; Mathey *et al.* 2020). The first is oriented N-S and is prolonged to the Southeast by
161 the Serenne fault. These two faults are located east of the Frontal Penninic Thrust while the Jausiers-
162 Tinée fault (external part) is parallel to and at about 10 km south of the Serenne fault.

163 Using satellite geodesy, there is no measurable horizontal strain (e.g. Walpersdorf *et al.* 2018). For
164 the vertical deformation, Mathey *et al.* (2021b) used INSAR data to obtain the vertical strain-rate in
165 the Western Alps and found a range of $\sim -0.7\text{mm/yr}$ to $\sim 0.5\text{mm/yr}$ for the Ubaye Region (see Fig. 3 in
166 Mathey *et al.* 2021b). Therefore, vertical deformation is very limited and not homogenous in the
167 Ubaye Region.

168 Several studies computed earthquake focal mechanisms and calculated stress-state in the Ubaye
169 Region and its surrounding areas to determine the tectonic stress regime. The mechanisms showed a
170 dominant extensional regime with a strike-slip component for the Ubaye Region (Fig. 1). Mathey et
171 *al.* (2021b) found that the extensional component is oblique to the strike of the Alpine belt, while it
172 was assumed normal until now (Sue *et al.* 2007). Similar stress-states have been reconstructed for
173 the last swarms (2003-2004 and 2012-2015) occurring in the Ubaye Region (Leclère *et al.* 2013;
174 Fojtíková and Vavryčuk, 2018).

175



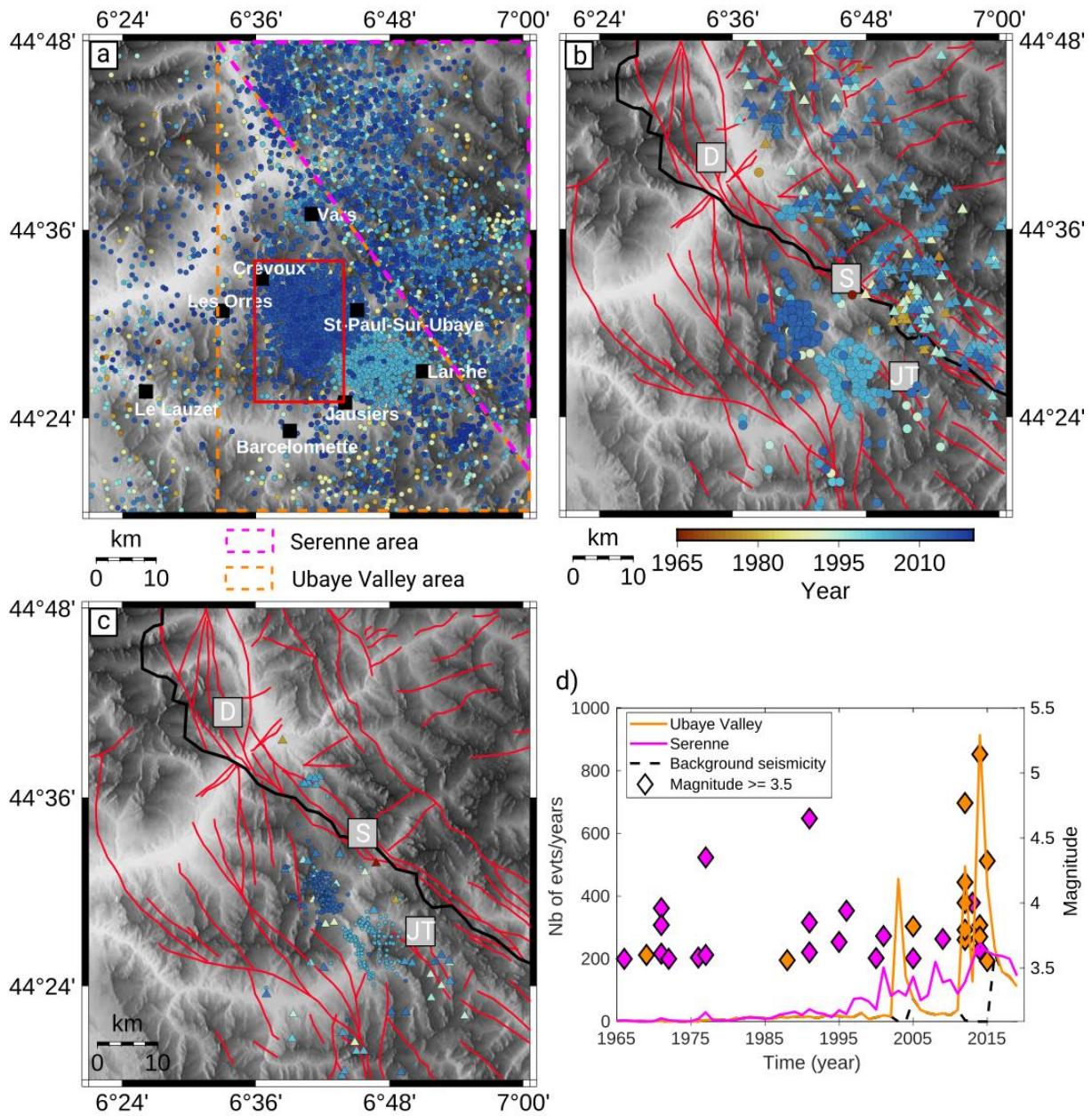
176

177 Figure 1: a) Simplified geological map of the Ubaye Region with focus on the late Alpine faults. (red

178 lines, from Tricart, 2004). Dots correspond to the location of earthquakes with their local magnitude

179 ML (from the LDG, Duverger *et al.* (2021)) greater than 3.2, the colour being proportional to ML.
180 Focal mechanisms of the largest events are from several catalogues: Fréchet and Pavoni (1979),
181 Leclère *et al.* (2013), Thouvenot *et al.* (2016), FMHex20 (Mazzotti *et al.* 2021) and Mathey *et al.*
182 (2021a). Their colours show the depth of the events. The clusters of events between Crévoux,
183 Jausiers, and Barcelonnette (2003-2004 and 2012-2015 swarms) have been artificially dispersed for
184 better visualization. b), Simplified geological cross-section oriented N050°, modified from Dumont *et*
185 *al.* (2022). The dots show the events with a computed focal mechanism (Fig. 2b). Black dots are
186 events with magnitude smaller than 3.2, while larger magnitude are coloured according to the scale
187 in a).. Only the events contained in a buffer of 10km apart from the cross-sections are represented.
188 The late Alpine faults are extended at depth by dotted lines for illustration but no information on the

189 dip of the faults at depth is available



190

191 Figure 2: a) Map of the instrumental seismicity since 1965 (LDG catalogue, <http://www-dase.cea.fr>).

192 The orange and pink dashed-line zones delimit the Ubaye Valley and the Serenne area respectively,

193 and the red square shows the 2014 relocated aftershock area (Fig. 3). b) Map of the earthquakes

194 with a computed focal mechanism in the Ubaye Valley (circle) and the Serenne area (triangle)

195 coloured by the year of occurrence. c) Map of the earthquakes with computed focal mechanisms for

196 the 2003-2004 swarms (light blue circle), the 2012-2015 sequence (dark blue circle), and of the

197 background seismicity (triangle) coloured by the year of occurrence. In b) and c), the red lines

198 correspond to the late Alpine faults mapped by Tricart (2004): D = High-Durance fault; S = Serenne
199 fault; JT = Jausiers-Tinée fault. d) Distribution of earthquakes in the LDG catalogue as function of the
200 time (years) for two different areas: the Ubaye Valley (orange) and the Serenne area (purple). The
201 dashed black line (Background seismicity) corresponds to events that occurred in the Ubaye Valley
202 during the periods: 1965-2002, 2005-2011, and 2016-2020. The diamonds correspond to the events
203 with magnitude greater than or equal to 3.5.

204

205 **2.2. Seismic activity**

206 Even if background seismicity is scattered in the full area, most events are located in two main areas:
207 in a large band along the Serenne and High-Durance faults in the internal units later referred as
208 “Serenne area”, and in large clusters between Crévoux, St-Paul-sur-Ubaye, and Barcelonnette in the
209 external units later referred as “Ubaye Valley area” (Fig. 2). The seismicity in these two areas behaves
210 differently (Baques *et al.* 2021; Larroque *et al.* 2021). In the Ubaye Valley, the seismicity mainly
211 occurs in clusters (Fig. 2a) and shows both mainshock-aftershocks and swarm sequences. On the
212 contrary, in the Serenne area, the seismicity is more scattered without clear clustered zones. The
213 seismicity mainly develops as mainshock-aftershocks sequences. Yet, the presence of small seismic
214 swarms cannot be discarded as no specific studies have been made on this area. In the following, we
215 briefly describe the two clusters that occurred in 2003-2004 and 2012-2015 in the Ubaye Valley area.

216 The 2003-2004 swarm (Fig. 2, light blue circles) started in early 2003 with seismic activity slowly
217 increasing until July 2003. The seismicity rate remained at a high level for 3 months with the
218 strongest earthquakes (maximum magnitude of $M_{LDG}3.4$) occurring during the end of the summer
219 2003. Then the seismicity rate slowly decreased to reach the background seismicity rate in the
220 summer 2004 (Jenatton *et al.* 2007; Baques *et al.* 2021). Finally, a small burst of activity occurred at
221 the end of 2004. More than 16,000 earthquakes were detected between 3km and 8km depth

222 (crystalline basement) and aligned along a 9km long structure oriented N150°E. Leclère *et al.* (2013)
223 computed 74 focal mechanisms that showed combined right lateral strike-slip motions along N150°-
224 to-N175°E structures with extensional components. Among the focal mechanisms, one of their nodal
225 planes shows similar orientation as the main direction of the cluster (~N150°E, Jenatton *et al.* 2007)
226 and to the orientation of the Jausiers-Tinée fault system. Leclère *et al.* (2013) proposed that the
227 Embrunais-Ubaye nappes, as well as the thick Mesozoic marly layers play the role of cap rock,
228 allowing the fluids to pressurize the upper crust and generate earthquake swarms.

229

230 Ten years later, the 2012-2015 sequence started 7 km north-west of the 2003-2004 swarm, (Fig. 2,
231 dark blue circles) on 26 February 2012 with a $M_{LDG}4.8$ (Mw4.2) earthquake at 8.8 km depth. It was
232 followed two years later, on 7 April 2014, by a larger shock of $M_{LDG}5.2$ (Mw4.8), which occurred
233 exactly at the same location but slightly deeper (9.9 km, Thouvenot *et al.* 2016). These mainshocks
234 were felt in Nice and Grenoble, 100 km from the epicentre area (Sira *et al.* 2012, 2014; Courboulex *et*
235 *al.* 2013). Following the two mainshocks, intense seismic crises started, and the 2014 crisis
236 instantaneously reactivated the 2012 crisis area. The seismicity then globally decreased as aftershock
237 sequences (Thouvenot *et al.* 2016; De Barros *et al.* 2019). During the two months (10 April 2014 - 10
238 June 2014) following the 7 April 2014 mainshock, a temporary network of 7 seismic stations was
239 deployed at 5 to 16 km from the mainshock epicentre. More than 9,000 events were detected by
240 template-matching and ~6,000 of them were relocated (Fig. 3) using a double-difference relocation
241 by De Barros *et al.* (2019). A modified Omori-Utsu's law (Utsu, 1961) fitted on these events showed a
242 $p=0.77$ time exponent, which is slower than usual and reflects a more complex behaviour (see Fig. 5
243 in De Barros *et al.*, 2019). This aftershock sequence is interpreted as composed of intertwined
244 aftershock-like clusters and swarm sequences (De Barros *et al.* 2019). After the 2014 mainshock-
245 aftershocks sequence, an event of $M_{LDG}3.6$ (Mw=3.4) occurred in mid-April 2015, followed by
246 another of $M_{LDG}4.3$ (Mw=4.2) in November 2015. Despite some classical mainshock-aftershocks

247 sequence characteristics, the full crisis (2012-2015) follows a swarm-like behaviour with the largest
248 event in the middle of the temporal sequence and a spatial extension much larger than the rupture
249 dimension of the largest event. The majority of the seismicity occurred in the crystalline basement
250 between 4 km and 10 km depth and aligned along a N160°E major structure and some N50°E minor
251 structures (Fig. 3, Fig. S1). These orientations are retrieved in the 13 focal mechanisms computed by
252 Thouvenot *et al.* (2016). In addition, no seismicity occurred in the area between the 2003-2004
253 swarm and the 2012-2015 sequence (Thouvenot *et al.* 2016). The activated structures, well
254 illuminated at depth (Thouvenot *et al.* 2016; De Barros *et al.* 2019) are not identified in the surface
255 geology, possibly due to the compliant sedimentary materials, which make it difficult to locate and
256 recognize active deformation. On the contrary, active structures are identified at the surface in the
257 internal zone (Sue *et al.* 2007), but the diffuse nature of seismicity does not allow to clearly associate
258 the seismicity to well identified single fault systems.

259

260

261 **3. Data: focal mechanisms**

262 In this study, we based our analysis on existing focal mechanisms computed previously in the Ubye
263 Region, and on focal mechanisms specially computed for the two months aftershocks following the 7
264 April 2014 mainshock.

265

266 **3.1 Ubye Region**

267 Thereafter, we gathered all the available focal mechanisms in a catalogue (full Ubye Region
268 catalogue). This catalogue is composed of the FMHex20 catalogue from Mazzotti *et al.* (2021),
269 mechanisms from Fréchet and Pavoni (1979), Leclère *et al.* (2013), Thouvenot *et al.* (2016), and

270 Mathey *et al.* (2021a). In total, 799 focal mechanisms composed the full Ubaye Region catalogue,
271 those with a magnitude larger than $M_{LDG}3.4$ (4% of the events) are displayed in Fig. 1. About 39% of
272 the focal mechanisms occurred with a pure strike-slip to strike-slip motion with a normal component,
273 while ~40% of the events show a pure normal to normal with a strike-slip component mechanism
274 (Fig. S2, Tab. S1). The remaining part shows strike-slip with a reverse component (12%) or reverse
275 mechanism (~9%). Two orientations stand out: N145°-to-N250°, N330°-to-N65° (Fig. S3). These
276 orientations are compatible with the active faults in the area. Indeed, the N145°-to-N250° is similar
277 to the orientation of the Serenne fault (N150°), while the N330°-to-N65° direction is consistent with
278 the High-Durance fault and with conjugate faults of the Serenne fault (Fig. 1).

279 We then refine the full catalogue by separating mechanisms from the Ubaye Valley (453 events, Fig.
280 2b) and from the Serenne area (314 events, Fig. 2b). We also create two subsets of focal mechanisms
281 for the Ubaye Valley: the 2003-2004 swarm area (138 events, Fig. 2c) and the 2012-2015 sequence
282 area (260 events, Fig. 2c). The remaining focal mechanisms in the Ubaye Valley are gathered to form
283 a background seismicity catalogue (55 events, Fig. 2c) that covers the periods of 1965-2002, 2005-
284 2011, and 2016-2020. No significant differences are observed either in the distributions of the fault
285 mechanisms or in the nodal plane orientations among the different sub-catalogues (see Tab.S1 and
286 Fig. S2).

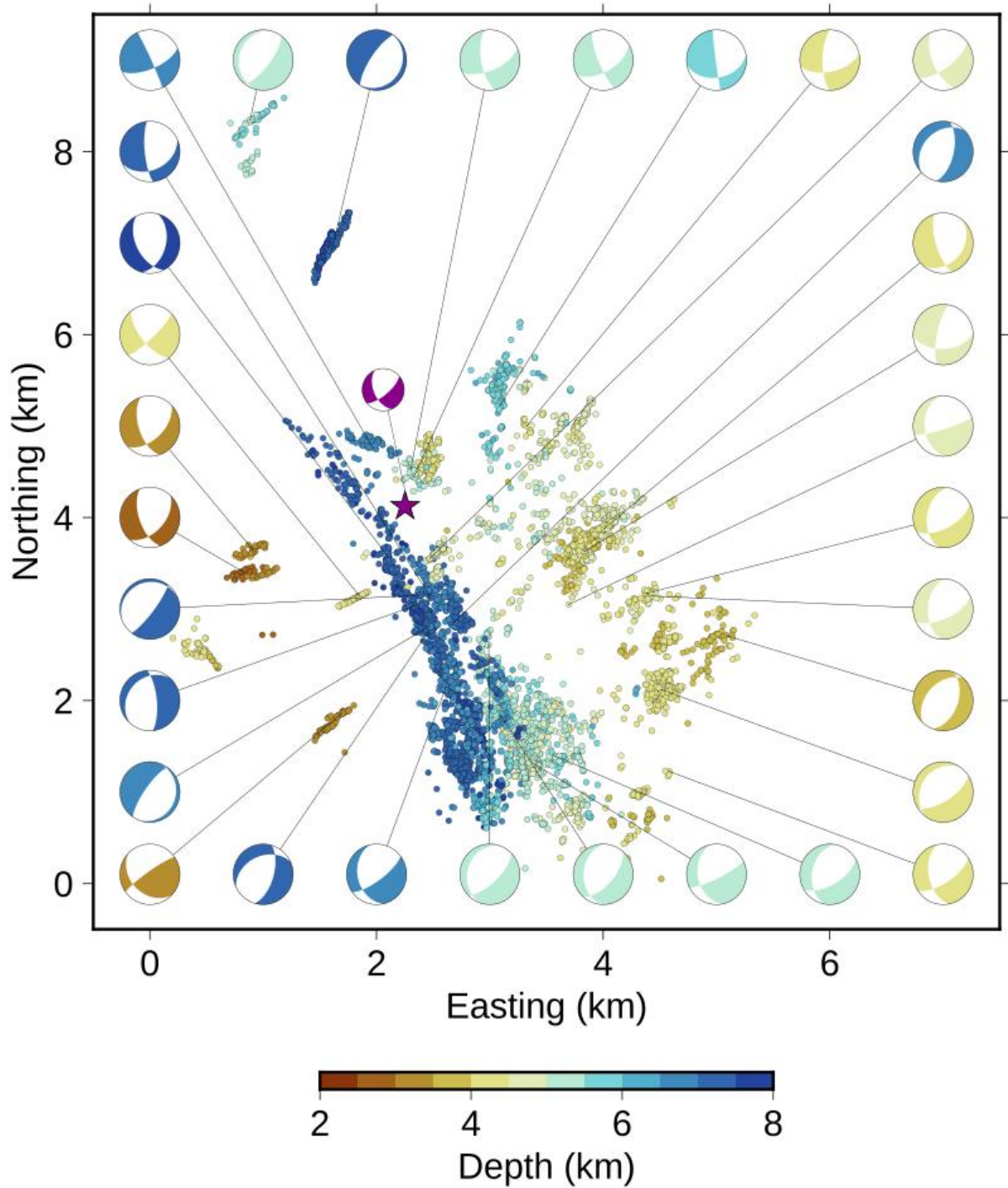
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288 **3.2 2014 Aftershocks**

289 For this sequence, only 3 focal mechanisms were already computed (Thouvenot *et al.* 2016; Mathey
290 *et al.* 2021a). We choose to refine the analysis by computing the focal mechanisms of the largest
291 events in the relocated catalogue of De Barros *et al.*, 2019. In total, 15 seismic stations were used: 7
292 from the temporary network and 8 permanent ones (3 from the INGV network (University of Genoa.,
293 1967; INGV Seismological Data Centre, 2006); 5 from the RESIF network (RESIF, 1995)). For each

294 event, we invert simultaneously the P-wave polarity and the S-to-P-waves amplitude ratio to
295 determine the focal mechanisms (Godano (2009), Godano et al. (2014)). For the amplitude ratio, we
296 first picked the amplitude of the first arrival for the P- and S-waves on the three components and
297 calculated the norm for both wave types. We carefully check the consistency of our results with the
298 focal mechanisms obtained by Thouvenot et al. (2016). Examples of focal mechanisms and their
299 uncertainties, together with the P-wave polarities and the S/P amplitude ratio, are shown in
300 Supplementary Fig. S4. We computed the mechanisms for which polarity and amplitude ratio are
301 available from at least 7 stations. At the end of the inversion, we kept solutions for which inversion
302 leads to score (inversed misfit function) greater than 0.4 (see Godano et al. 2014). The 100 focal
303 mechanisms then form the 2014 relocated aftershocks catalogue.

304 The 2014 relocated aftershocks catalogue is mainly composed of 70% of pure-normal to normal
305 mechanisms with a strike-slip component and 20% of strike-slip mechanisms with a normal
306 component (Tab.S1 and Fig. S2). Even if the ratio of normal mechanisms is higher than in the Ubaye
307 Valley catalogue (Fig. S2), we retrieve similar orientations for the nodal planes (N170°-to-N230°,
308 N355°-to-N70°; Fig. S3). However, we do not retrieve the N155° orientation proposed for the 2014
309 mainshock. Globally, the 2014 catalogue shows a wide variety of focal mechanisms and orientations,
310 which suggests that complex processes may be in play in this aftershock sequence (Fig. 3). Finally, as
311 the Ubaye catalogue and the 2014 aftershocks come from different origins, they will not be used
312 together, but in separate inversion process, for the stress-state reconstruction and the fluid-pressure
313 computation.



314

315 Figure 3: Map of the aftershocks relocated by De Barros et al. (2019) in the two months following the
 316 7 April 2014 mainshock, with a subset of 32 (out of 100) computed focal mechanisms. The colours
 317 show the event depths. The purple star corresponds to the location of the 2014 mainshock with its
 318 focal mechanisms from Thouvenot et al. (2016).

319

320

321 **4. Methods**

322 **4.1 Shape and orientation of the stress tensor**

323 To estimate the fluid-overpressure required to trigger earthquakes, we use the focal mechanisms to
324 reconstruct the stress-state orientation. To do that, we use the MSATSI program (Martínez-Garzón *et*
325 *al.* 2014), which is a MATLAB version of the SATSI program developed by Hardebeck and Michael
326 (2006). It is based on the Wallace (1951) and Bott (1959) assumption: the slip vector is assumed
327 parallel to the maximal shear-stress. The stress-state is then determined through a grid-search
328 inversion in order to maximize the shear-stress acting on the slip direction for all earthquakes. The
329 MSATSI software requires as input the dip direction (strike+90°), the dip angle, and the rake, and
330 computes the orientation of the principal stresses and the shape-ratio φ (Michael, 1984):

$$\varphi = \frac{\sigma_2 - \sigma_3}{\sigma_1 - \sigma_3} \quad (1)$$

331 The shape-ratio informs on the stress regime in the region associated with the orientation of the
332 principal stresses (e.g. Gephart and Forsyth, 1984). Equivalently, we will use the R-ratio, defined as
333 $R = 1 - \varphi$:

$$R = \frac{\sigma_1 - \sigma_2}{\sigma_1 - \sigma_3} \quad (2)$$

334 A R-ratio of 0 indicate a purely extensive regime, while values of 0.5 and 1 indicate strike-slip and
335 compressive regime, respectively. We also calculate the uncertainties associated with the orientation
336 of each stress component and shape-ratio, following the approach of Martínez-Garzón *et al.* (2016).
337 For the 2014 relocated aftershocks catalogue, we used the mean uncertainty obtained when
338 calculating the focal mechanisms to perturb the strike (+/- 25°), dip (+/- 12°), and rake (+/- 36°).
339 These uncertainties are obtained by calculating 100 times a solution for each focal mechanism. As we

340 did not have information about the uncertainty associated with the published focal mechanism, we
341 chose to perturb the parameters with a maximum of +/- 30°, which is consistent with the
342 uncertainties we obtained for the 2014 relocated aftershocks. Then, randomly perturbing 100 times
343 the strike, dip, and rake of each mechanism assuming a uniform distribution, we create 100 new
344 catalogues, on which we apply 100 bootstraps operations. The inversions then lead to 10,000 values
345 of shape-ratio, trend and plunge for each stress. From then, we calculate the mean and the standard
346 deviation (Tab. S2, Fig. S5, Fig. S6).

347 We then aim at exploring how the stress-state is variable in space and time, calculating it for the
348 different focal mechanism catalogues defined in previous section: full Ubaye Region, Ubaye Valley,
349 Serenne area, Background seismicity, 2003-2004, 2012-2015, and 2014 relocated aftershocks. The
350 results of these inversions are presented in the Results section.

351

352 **4.2 Fluid-pressure**

353 As mentioned in the introduction, fluid-pressure was highlighted as a driving process of the Ubaye
354 Region seismicity. Two previous studies (Daniel *et al.* 2011; Leclère *et al.* 2013) calculated the fluid-
355 overpressure needed to explain the 2003-2004 swarm and both assumed a stress-state with two
356 horizontal components (σ_1 and σ_3) and one vertical component (σ_2). However, in our case, the
357 stress-state inversed for the 2012-2015 sequence (Fojtíková and Vavryčuk, 2018) has only one
358 horizontal component (σ_3) and two components (σ_1 and σ_2) neither vertical nor horizontal.
359 Therefore, we generalize the approach used by Collettini and Trippetta (2007), Terakawa *et al.*
360 (2010), Leclère *et al.* (2013), and Jansen *et al.* (2019), to calculate the fluid-overpressure.

361 As neither σ_1 nor σ_2 are vertical, we first express the vertical stress (σ_v) as a function of the principal
362 stress components using Cauchy's equation. Assuming a lithostatic stress, we get:

$$\sigma_v = \cos^2(\theta_1)\sigma_1 + \cos^2(\theta_2)\sigma_2 = \rho gZ \quad (3)$$

363 With θ_1 and θ_2 corresponding to the angle between the vertical stress and the plunge of σ_1 and σ_2 . ρ
 364 is the rock density taken at 2700 kg/m^3 ; g is the gravitational constant (rounded at 10 m/s^2) and Z is
 365 the depth in meters.

366 Using the R-ratio (eq. 2) determined in the stress inversion, we can then reorganize eq. 3 to express
 367 σ_1 and σ_2 as function of σ_3 :

$$\sigma_1 = \frac{\rho g Z - R \cos^2(\theta_2) \sigma_3}{(1 - R)(\cos^2(\theta_2) + \cos^2(\theta_1))} \quad (4)$$

$$\sigma_2 = \frac{\rho g Z - R(\rho g Z + \cos^2(\theta_1) \sigma_3)}{\cos^2(\theta_1) + \cos^2(\theta_2) - R \cos^2(\theta_2)} \quad (5)$$

368 We then assume that σ_3 depends linearly on the depth (Z), such that:

$$\sigma_3 = AZ \quad (6)$$

369 Therefore, the only unknown left to determine σ_1 and σ_2 in eq. 4 and 5 is the coefficient A . To
 370 determine it, we first compute the shear and normal stress that apply on the faults. As we do not
 371 know yet which of the nodal planes inferred from the mechanism the fault plane is, both nodal
 372 planes are considered as possibly activated. The shear stress τ and normal stress σ_n are related to
 373 the principal stress components using Cauchy's equations:

$$\tau^2 = (\sigma_1 - \sigma_2)^2 l^2 m^2 + (\sigma_2 - \sigma_3)^2 m^2 n^2 + (\sigma_3 - \sigma_1)^2 n^2 l^2 \quad (7)$$

$$\sigma_n = l^2 \sigma_1 + m^2 \sigma_2 + n^2 \sigma_3 \quad (8)$$

374 With l^2 , m^2 , and n^2 , the cosine direction of σ_1 , σ_2 , and σ_3 respectively.

375 We can then define the slip-tendency α for all nodal planes following Mattila (2015) and Jansen et al.
 376 (2019). This parameter describes how close the fault stress-state is from failure. A fault is at a critical
 377 stress state for α equal to 1, while it gets further from failure when α decreases. If we first assume
 378 that all faults are at a hydrostatic pressure, α is defined as:

$$\alpha = \frac{\tau}{\mu(\sigma n - \rho_f g Z)} \quad (9)$$

379 With $\rho_f=1000 \text{ kg/m}^3$ the water density and μ the friction coefficient. We here assume a standard
 380 value of $\mu=0.6$ (Byerlee, 1970).

381 When combining eq. 4, eq. 5 and eq. 6 with eq. 7 and eq. 8, the slip-tendency (eq. 9) only depends on
 382 A. We then perform a grid search with A varying from 0 to ρg , with a step of 1 N/m^3 . For each value
 383 of A, we calculate the slip-tendency for all fault planes. We then select the best A-value, 1) given that
 384 the slip tendency cannot be higher than 1, and 2) assuming that at least one event occurs with a fluid
 385 pressure equal to the hydrostatic pressure, meaning that the most favourably oriented fault plane is
 386 reactivated with hydrostatic fluid pressure (Terakawa *et al.* 2010; Terakawa *et al.* 2012). This
 387 assumption tends to minimize the fluid pressure required to reach the reactivation threshold.
 388 Therefore, we are looking for an A-value for which the maximal slip tendency is equal to 1. Once A is
 389 determined, σ_1 , σ_2 , and σ_3 (eq. 4, 5, and 6), as well as the stress state τ and σn acting on the faults
 390 (eq. 7 and eq. 8) can be computed. Assuming cohesionless fault, we can then compute the fluid-
 391 pressure that is required to induce failure using the Amontons' law:

$$PF = \sigma n - \frac{\tau}{\mu} \quad (10)$$

392 For each event, a fluid-pressure value is computed for both nodal planes. To discriminate the fault
 393 plane from the auxiliary plane, we could use either the misfit angle, which is the difference in the
 394 direction between the slip and the maximal shear stress on the fault plane if considering the Wallace
 395 and Bott assumption (Wallace, 1951; Bott, 1959), or the slip tendency (eq. 9), hence the lowest fluid
 396 pressure, in the Mohr-Coulomb theory. For $\sim 75\%$ of the events, the plane with the highest slip
 397 tendency is also the plane with the lower misfit angle or one of two planes with similar misfit angle
 398 (see supplementary Fig. S7). For $\sim 25\%$ of the events, there is a discrepancy between the plane with
 399 the lower misfit angle and the one with the highest slip tendency. As the (Wallace, 1951) and (Bott,
 400 1959) assumption may show some limits (Kakurina *et al.*, 2019) and because we compute the full

401 stress tensor, we choose to keep as the fault plane the nodal plane the closest to failure, which is
 402 also the plane with the lowest fluid-pressure. The average misfit angles of the chosen fault planes is
 403 $\sim 19^\circ$ for the 2014 aftershocks (see supplementary Fig. S7), with only a few events (7%) with a misfit
 404 angle larger than 40° , but lower than 75° . To further validate that our fault plane determination is
 405 correct, we compare the chosen fault plane with the one proposed by the fault instability criterion
 406 (Vavryčuk *et al.* 2013). The fault instability criterion gives the nodal plane that is the closest to the
 407 rupture without resorting of stress-state computation. It only depends on the friction coefficient μ ,
 408 the R-ratio, and the direction cosines from the principal stresses (Vavryčuk *et al.* 2013):

$$I = \frac{\tau_r + \mu (\sigma_r + 1)}{\mu + \sqrt{1 - \mu^2}} \quad (11)$$

409 where τ_r and σ_r are equal to:

$$\tau_r = \sqrt{l^2 + (2R - 1)^2 m^2 + n^2 - (-l^2 + (2R - 1)m^2 + n^2)^2} \quad (12)$$

$$\sigma_r = -l^2 + (2R - 1)m^2 + n^2 \quad (13)$$

410 We find that 99% of the identified fault planes are similar by both approaches, confirming their
 411 determination.

412

413 The fluid-pressure depends on the depth as it is assumed to be higher than the hydrostatic pressure:
 414 $Ph = \rho_f g Z$ (14). To work with comparable values, we subsequently use the fluid-overpressure,
 415 defined as the part of fluid-pressure above the hydrostatic pressure (MPa):

$$\Delta PF = PF - Ph \quad (15)$$

416 We can then define the overpressure ratio (%) by normalizing the overpressure by the hydrostatic
 417 pressure:

$$\frac{\Delta PF}{Ph} = \frac{(PF - Ph)}{Ph} * 100 \quad (16)$$

418 Equivalently, another way to observe overpressure is through the pore-fluid factor (Sibson, 1990;
419 Cox, 2016), which is the ratio of the pore-fluid pressure to the lithostatic pressure. We choose in this
420 paper to use the overpressure ratio to compare with other studies using similar methods (e.g. Jansen
421 *et al.* 2019), but results are shown in supplementary figures (Fig. S8, Fig. S9, Fig. S10, Fig. S11).

422 At the end, we obtain the fluid-overpressure needed to trigger slip on all fault planes of the focal
423 mechanisms. We also estimate fluid-overpressure uncertainties by randomly perturbing 1000 times
424 the strike, dip, and rake of the focal mechanisms on a range of 30° in a uniform distribution law. We
425 then run the full inversion process on these 1000 perturbed catalogues, to reconstruct the stress-
426 state and overpressures, leading to 1000 perturbed values of fluid-pressure (Fig. S12). Since the
427 distribution is not symmetrical, we quantify the uncertainties for each event by measuring the
428 overpressure range for which 34% (half of one standard deviation if the distribution was Gaussian) of
429 the calculated values are above or below the initial fluid-pressure value. We then compute the
430 median of these uncertainties for the different catalogues. We obtain an overpressure ratio
431 uncertainty of -15% and +21% for Ubaye Valley, of -18% and +24% for Serenne, and of -17% and
432 +15% for 2014 relocated aftershocks.

433

434 **5. Results**

435 **5.1 Orientation and amplitude of the stress tensor**

436 We determine the stress-state for the two months following the 2014 mainshock as well as for the
437 full Ubaye Region, its two sub-areas (Serenne and Ubaye Valley) and the particular sequences (2003-
438 2004 and 2012-2015 sequences). Fig. 4 summarizes the stress-state orientations calculated here and
439 compares them to the published ones (Leclère *et al.* 2013; Fojtíková and Vavryčuk, 2018).

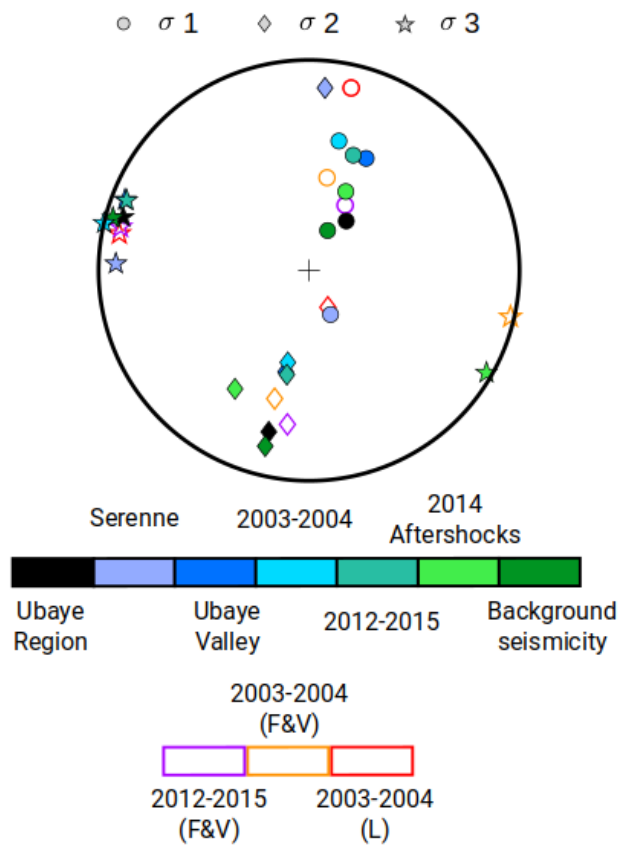
440 For the full Ubaye Region, we obtain a σ_1 oriented $N37^\circ \pm 8^\circ$ and plunging $66^\circ \pm 6^\circ$, a σ_2 oriented
441 $N192^\circ \pm 2^\circ$ plunging $22^\circ \pm 6^\circ$, and a sub-horizontal σ_3 oriented $N285^\circ \pm 1^\circ$. The σ_3 is the only
442 horizontal stress component and is tilted $\sim 65^\circ$ from the normal of the Frontal Penninic Thrust. The
443 other two stress components, σ_1 and σ_2 , are less well-resolved (Tab.S2, Fig.S6). Indeed, they can
444 easily vary in a $N190^\circ$ vertical plane, which suggests that both components have similar amplitudes.
445 Furthermore, the Ubaye Region shows strong variations in the topography, with more than 2000 m
446 between valleys and mountain summits. As σ_1 and σ_2 depend on the depth, topographical effects
447 may impact the plunge, as well as the amplitude, of these stress components.

448 We then refine this analysis by looking for spatial variations for the two sub-areas: Ubaye Valley and
449 Serenne. In both cases, σ_3 orientation is found similar to the Ubaye Region one while there are
450 differences in the plunge for the other two components. Both σ_1 and σ_2 are within the $N190^\circ$ vertical
451 plane, but σ_1 is found subvertical (71°) in the Serenne area, while it shows a 41° plunge in the Ubaye
452 Valley. Accordingly, σ_2 is sub-horizontal (14°) and more vertical (49°) in the Serenne and Ubaye
453 Valley, respectively. As mentioned above, this difference might be (partly) due to topographical
454 effects.

455 The stress-state orientations for the two main sequences of 2003-2004 and 2012-2015 and for the
456 remaining focal mechanisms (background seismicity) in the Ubaye Valley are found very similar to
457 the Ubaye Valley one (Fig. 4, Tab.S2). Some very limited differences ($<5^\circ$) in the plunge of σ_1 and σ_2
458 might be explained by the variation of topography or uncertainties in the computation. Same
459 observations can be noticed when looking specifically at the aftershocks of the 2014 event: the
460 stress-state does not show significant differences with the other ones.

461 Moreover, the stress-state orientation we determined are similar to the ones computed by Fojtíková
462 and Vavryčuk (2018), the differences being within the uncertainty range. The stress-state computed
463 by Leclère et al. (2013) for the 2003-2004 sequence shows a subvertical σ_2 , while they use the same
464 catalogue of mechanisms as Fojtíková and Vavryčuk (2018) who found plunging of 49° (Tab.S2).

465 We also retrieve similar R-ratio ≈ 0.31 , for the different catalogues (Tab.S2). Their values are also
 466 compatible with the ones found by Leclère et al. (2013) for the 2003-2004 swarm and by Fojtíková
 467 and Vavryčuk (2018) for the 2003-2004 and 2012-2015 swarms with a mean R-ratio ≈ 0.29 . This value
 468 indicates extension and strike-slip regime in this region, which is in agreement with geodetic
 469 observations.



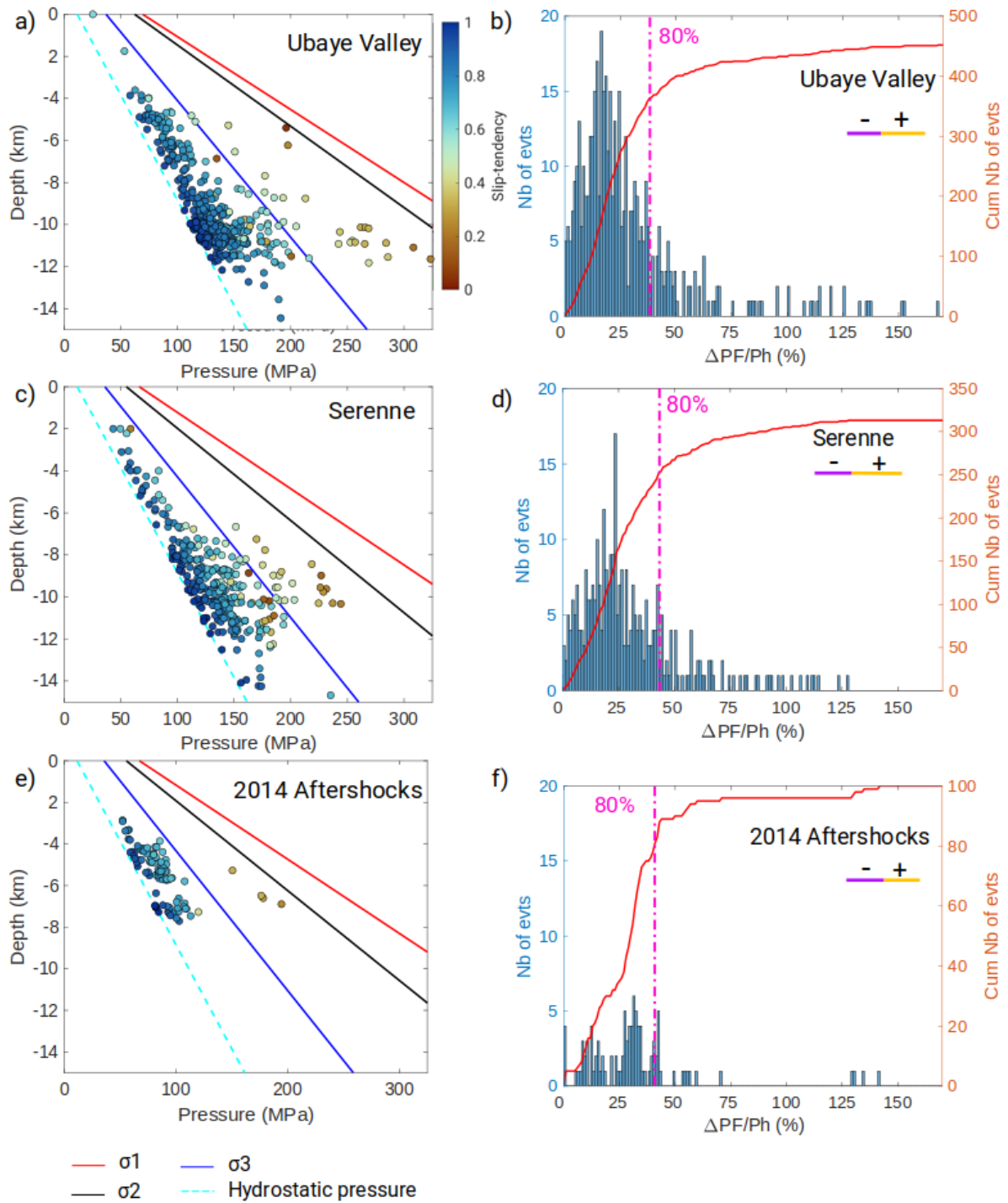
470
 471 Figure 4: Polar view of the stress-state orientations (circle for σ_1 , diamond for σ_2 , and stars for σ_3).
 472 Computed stress-states are shown as filled symbols coloured by regions (full Ubaye Region, and its
 473 two sub-areas: Serenne and Ubaye Valley, Fig. 2b); or periods (2003-2004 swarm, 2012-2015
 474 sequence, background seismicity, Fig. 2c; and 2014 aftershocks, Fig. 3). The red, pink, and purple
 475 unfilled symbols correspond to published stress-states: “F&V” for Fojtíková and Vavryčuk (2018); “L”
 476 for Leclère et al. (2013).

477

478

479 The magnitude of the principal stresses is then determined (Figs 5a, c, and e, for Ubaye Valley,
480 Serenne, and 2014 relocated aftershocks catalogues, respectively). In all cases, they are found to be
481 very similar, as $\sigma_3 \approx 15 * Z$ (MPa), with the depth Z in km. Equivalently, the deviatoric stress is found to
482 increase with depth as $\sigma_1 - \sigma_3 \approx 13 * Z$ (MPa). As expected, σ_1 is close to σ_2 , even if their magnitudes
483 vary slightly among cases, as the stress-state inversion reveals differences, albeit small, in their
484 plunge. Once again, these differences may be explained by topographic effect, as the 2014
485 aftershocks occurred below the Parpaillon massif, which culminates at 3000 m.

486 In summary, the stress-states and the R-ratio are found similar for the different catalogues, and
487 consistent with previously published results. Therefore, anomalies in the stress-state orientation or
488 magnitudes cannot explain the swarm behaviours or the complex aftershock sequences of the 2014
489 earthquake. As the stress-state seems spatially homogeneous constant, we then mainly focus on the
490 Ubaye Valley, Serenne area, and 2014 aftershocks to explore their fluid-overpressure behaviour.



491

492 Figure 5: Fluid-pressure versus depth (left panels: a, c, e) and histograms of overpressure ratios (right
 493 panels: b, d, f), for the Ubaye Valley (a, b), the Serenne area (c, d), and the 2014 aftershocks (e, f). In
 494 a), c), and e), the dots show the fluid-pressure required to reactivate the selected fault planes
 495 coloured by the slip-tendency at hydrostatic pressure. The depth is below sea level. The light blue,
 496 blue, black, and red lines correspond to the hydrostatic pressure, σ_3 , σ_2 , and σ_1 , respectively. In b),

497 d), f), the histograms show the number of events per overpressure ratio (%), while the red lines show
498 the cumulative number of events. The purple and yellow bars correspond to the common negative
499 and positive uncertainty that are found to all the fluid-overpressure values. The pink dashed line
500 shows 80% of the cumulative number of events. The number of earthquakes as function of pore-fluid
501 factor is also available in Supplementary Materials (Fig. S8).

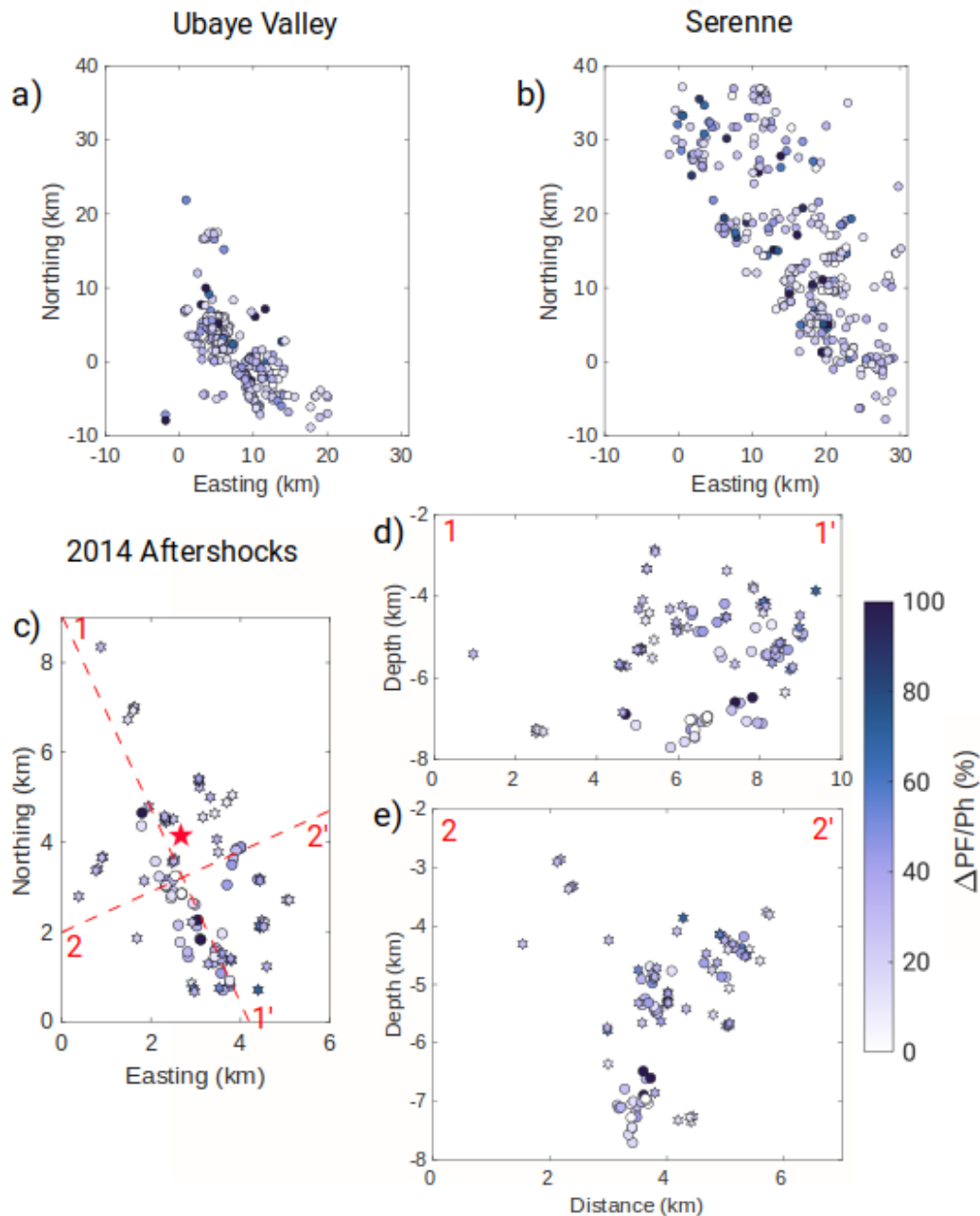
502

503 **5.2 Fluid-overpressure**

504 Figs 5a, c, and e shows the fluid-pressure computed for the nodal planes the closest to rupture.
505 These planes are assumed to be the fault planes of the seismic event. This assumption is supported
506 by the correlation of our results with the values of the fault instability criterion described in section
507 4.2. All computed fluid-pressures are equal or above the hydrostatic pressure, as we assumed a
508 minimal hydrostatic pressure. Only a few events require an overpressure higher than σ_3 . In this case,
509 failure might imply some tensile component. Some of the published focal mechanisms used in our
510 catalogues have quality indexes (Mathey *et al.* 2021a). We find that events with a fluid-pressure
511 value higher than σ_3 have the lowest quality index (Figs S11b and d). We then discard events with a
512 fluid-pressure value higher than σ_3 from the analysis as their mechanisms might be fraudulent with
513 large errors.

514 To get rid of the dependency on depth, we also look at the overpressure ratios. We consider that the
515 overpressure required to trigger the events is significant when it is higher than the computed
516 minimum uncertainties: 15% for Ubaye Valley (Fig. 5b), 18% for Serenne (Fig. 5d), and 17% for the
517 2014 aftershocks (Fig. 5f). Considering these uncertainties, 68% (Ubaye Valley), 61% (Serenne), and
518 69% (2014 aftershocks) of the events required overpressure to be triggered, leading to median
519 overpressure ratio of 27%, 30%, and 32% for the Ubaye Valley, Serenne, and 2014 aftershocks,
520 respectively. In absolute values, at least 46% of the events require fluid-overpressure less than 45

521 MPa for the Serenne area. For the 2014 aftershocks catalogue, at least 60% of the events need less
 522 than 27 MPa of overpressure. 48% of events require less than 38 MPa of overpressure for the case of
 523 the Ubaye Valley area (Fig. S14). The three catalogues lead to similar values of median overpressure:
 524 24 MPa, Ubaye Valley catalogue; 30 MPa, Serenne catalogue; 20 MPa, 2014 relocated aftershocks
 525 catalogue.



526
 527 Figure 6: Map view of the overpressure ratio (%) for the Ubaye Valley (a), the Serenne area (b), and
 528 the 2014 relocated aftershocks (c). In c), the red star shows the location of the 2014 mainshock

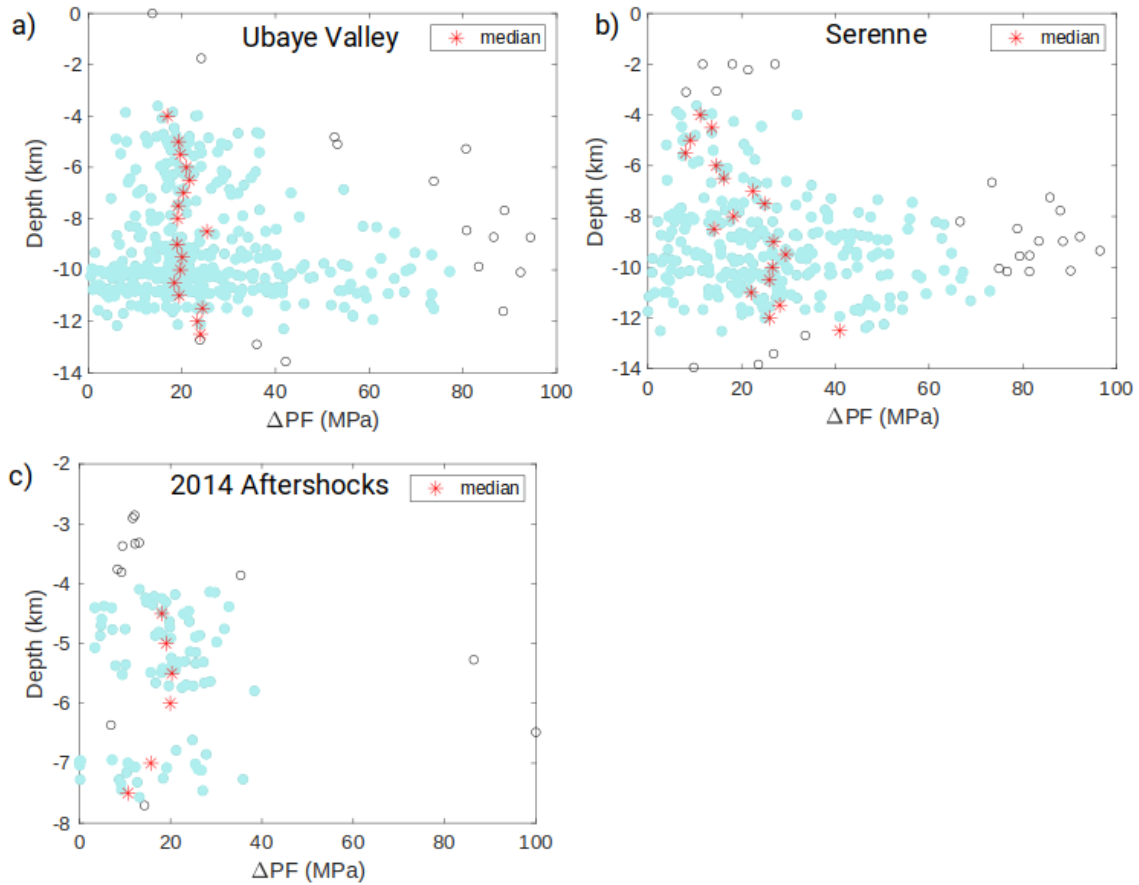
529 (Thouvenot *et al.* 2016). The dashed red-lines show the cross-sections that are used in d) and e). In
530 c), d), and e), the hexagrams and circles correspond to events belonging to swarm-like clusters and
531 aftershock-like clusters, respectively. The depth is below sea level. The same figure, with pore-fluid
532 factor as function of location is available in Supplementary Materials (Fig. S9).

533

534

535 We then analyse the spatial distribution of the reconstructed fluid-pressure (Fig. 6), but we do not
536 see any clear tendency. Indeed, despite an important scattering in overpressure ratios, no consistent
537 anomalies are observed on map view, neither for the 2014 relocated aftershocks (Fig. 6c) nor for the
538 Ubaye Valley and Serenne areas (Figs 6a and b). The families that composed the 2014 aftershocks
539 were classified by De Barros *et al.* (2019) into aftershock-like or swarm-like behavior based on their
540 space, time and magnitude distribution (Figs 6c, d, and e). The median value of overpressure ratios
541 for aftershock-like clusters is slightly lower (27%) than the one for swarm-like clusters (31%). This is
542 to be expected, as swarms are commonly driven by fluid-driven processes. However, no clear
543 difference is observed spatially between events belonging to clusters with a different behavior (Figs
544 6c, d, and e).

545 There is also no clear dependency between the depth and fluid-overpressure for the Ubaye Valley
546 (Fig. 7a), as the median values are found nearly constant between 4-12km depth. For the 2014
547 aftershocks, we also observe a constant fluid-overpressure as a function of depth (Fig. 7c). Therefore,
548 for both catalogues, the overpressure does not seem to depend on the hydrostatic pressure. On the
549 contrary, for the Serenne area, the reconstructed overpressure seems to increase with depth, and
550 therefore with the value of the hydrostatic level. This difference in behaviour between Serenne and
551 Ubaye Valley might suggest a difference either in the geological context or in the driving processes.



552

553 Figure 7: Fluid-overpressure (MPa) versus depth for the Ubaye Valley (a), the Serenne area (b), and
 554 the 2014 relocated aftershocks (c). Each dot corresponds to a focal mechanism and depth is directly
 555 associated to each solution. Each dot corresponds to an event with a computed focal mechanism.
 556 Red stars show the median value of the fluid-overpressure on a 500-m-depth sliding window with a
 557 250 m overlap. Only median values calculated with at least 5 events are represented. Empty circles
 558 correspond to events having an overpressure above σ_3 , and are thus discarded for the median
 559 calculation. The depth is below sea level.

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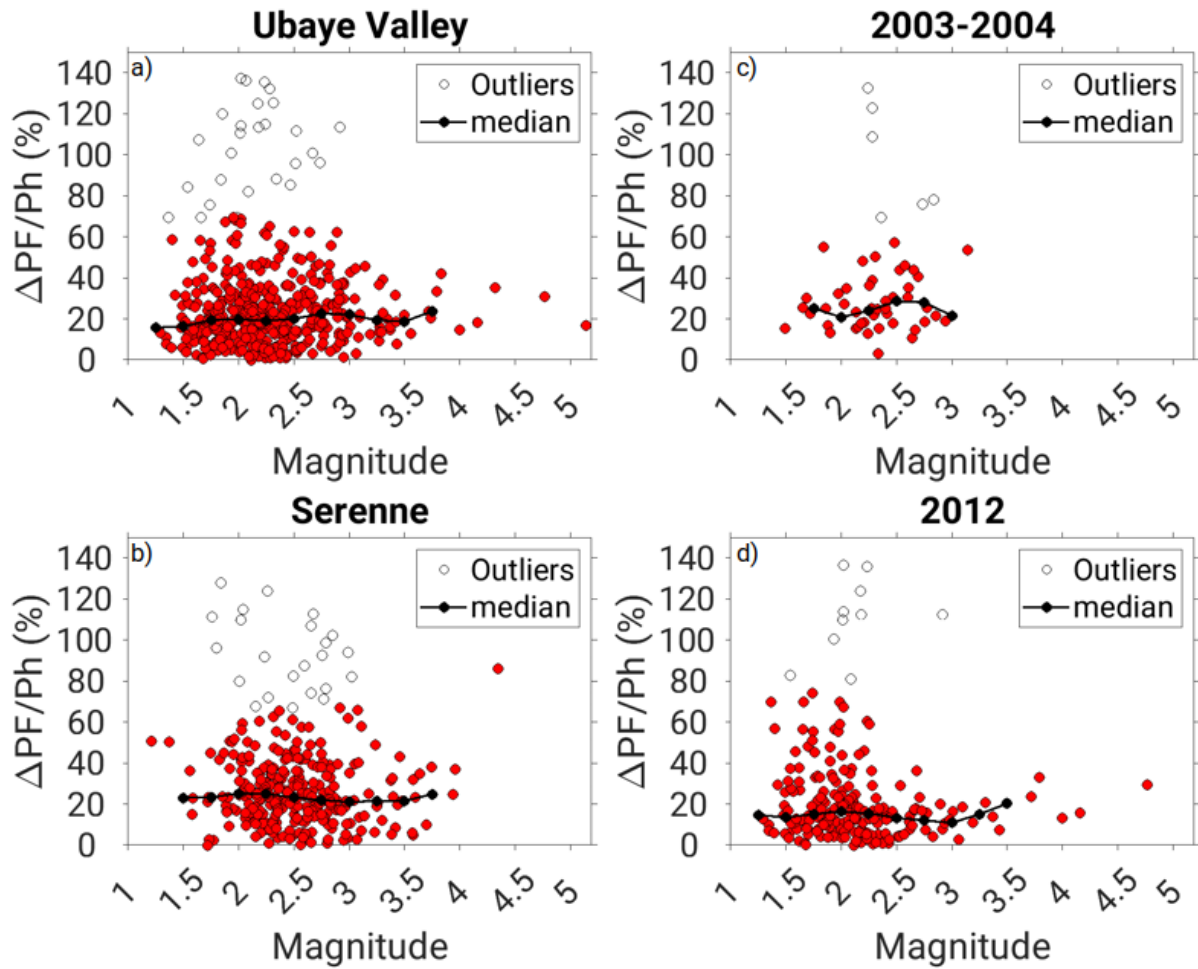
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562 Fig. 8 represents the overpressure ratio as a function of the event magnitude (ML) for the different
 563 areas and seismic sequences. For the sake of homogeneity, we use the local magnitude (ML)
 564 computed by the LDG (Duverger *et al.* 2021) for all events in our focal mechanism catalogues. In both

565 the Ubaye Valley and the Serenne areas, the median of the reconstructed fluid-pressure values does
566 not depend on the event magnitude. The same is observed for the overpressure (Fig. S15).
567 Particularly, the largest events need similar level of fluid-pressure as the average for the smallest
568 ones. The same observation is made when focusing on particular sequences, such as the seismicity in
569 2012 (Fig. 8d). The reconstructed overpressure ratio appears more scattered for low magnitude
570 events ($ML < 2.5$) as the number of events decreases with magnitude. This might also indicate a
571 variability in the processes or a larger uncertainty on the focal mechanisms for lower magnitude
572 events. However, the mainshock that started this sequence ($ML 4.8$, 26/02/2012) needs $\sim 30\%$ of
573 overpressure ratio (overpressure of ~ 35 MPa, Fig. S15d), which is above the median value of the
574 reconstructed overpressure ratio and one of the largest for events of magnitude larger than 2.5 (Fig
575 8d). This suggests that fluid-pressure is required to trigger both large and small events, whatever
576 their magnitude.

577 To confirm this statement, we look in detail at fluid-pressure uncertainties for particular
578 earthquakes. Figure 9 represents the distribution of fluid-overpressures obtained by perturbing 1000
579 times the source fault properties (strike, dip, and rake) of all events (see section 4.2). For each of
580 these perturbed catalogues, we run again the full analysis, including the inversion for the stress-state
581 orientation, amplitude and the fluid-overpressure. We then focus on 9 events with magnitude ML
582 from 2.9 to 5.2. Two of these earthquakes occurred in the Serenne area (Figs 9b and c) in Feb. 1996,
583 and 2000, two were from the background seismicity of the Ubaye Valley (Figs 9a and f) in Jan. 1996
584 and 2005, and two happened during the 2003-2004 swarm (Figs 9d and e). The last three events are
585 the largest earthquakes of the 2012-2015 sequence, which occurred on 26/02/2012 (Fig. 9g),
586 07/04/2014 (Fig. 9h), and 06/11/2015 (Fig. 9i). Some of these events unambiguously need fluid-
587 overpressure to reach the reactivation criterion (Figs 9b, c, e) as their distributions show a clear peak
588 around a high fluid-overpressure value. The overpressure distributions of the earthquakes occurring
589 in 2003 and 2012, (Figs 9d, g and h) are wider but still suggest an activation by fluid-pressure.
590 However, this is not as clear for the Jan. 1996, 2009 and 2015 events (Figs 9a, f and i) for which their

591 distribution is squeezed toward low values of overpressure. Overall, by considering uncertainties, the
592 data presented in Fig. 9 confirm the need for overpressures triggering earthquake, even if the
593 reconstructed values of overpressures might be variable. Therefore, this confirms that even the
594 largest earthquakes are likely to have been caused by fluid pressure, whether they occurred as part
595 of a swarm or as an isolated earthquake. To refine this observation, we look at how the overpressure
596 ratio evolves with time within a swarm and during an aftershock sequence (Fig. 10). During the 2003-
597 2004 swarm (Fig. 10a), the overpressure ratio varies in time at high level without following a clear
598 pattern. On the contrary, during a mainshock-aftershocks sequence, like in 2012 (Fig. 10b), high
599 values of overpressure ratio (up to 68%) are observed at the beginning of the sequence, right after
600 the mainshock. It then slowly decays in the next 6 months, to stabilize below 30% (Fig. 10b, black
601 line). Therefore, it suggests that the difference between a swarm and a mainshock-aftershock
602 sequence may be more related to the evolution of the fluid-overpressure in time rather than to its
603 absolute values. Both required overpressure in their driving process, but the overpressure ratio
604 decays with time in the mainshock-aftershocks sequence and not in a swarm episode.



605

606

Figure 8: Overpressure ratio as function of local magnitude for the Ubaye Valley (a), the Serenne area

607

(b), the 2003-2004 swarm (c), and after the 2012 mainshock (d). The local magnitude comes from the

608

LDG catalogue (<http://www-dase.cea.fr>). The empty circles show events with overpressure ratio

609

higher than σ_3 , which are therefore discarded from the analysis. The black line shows the floating

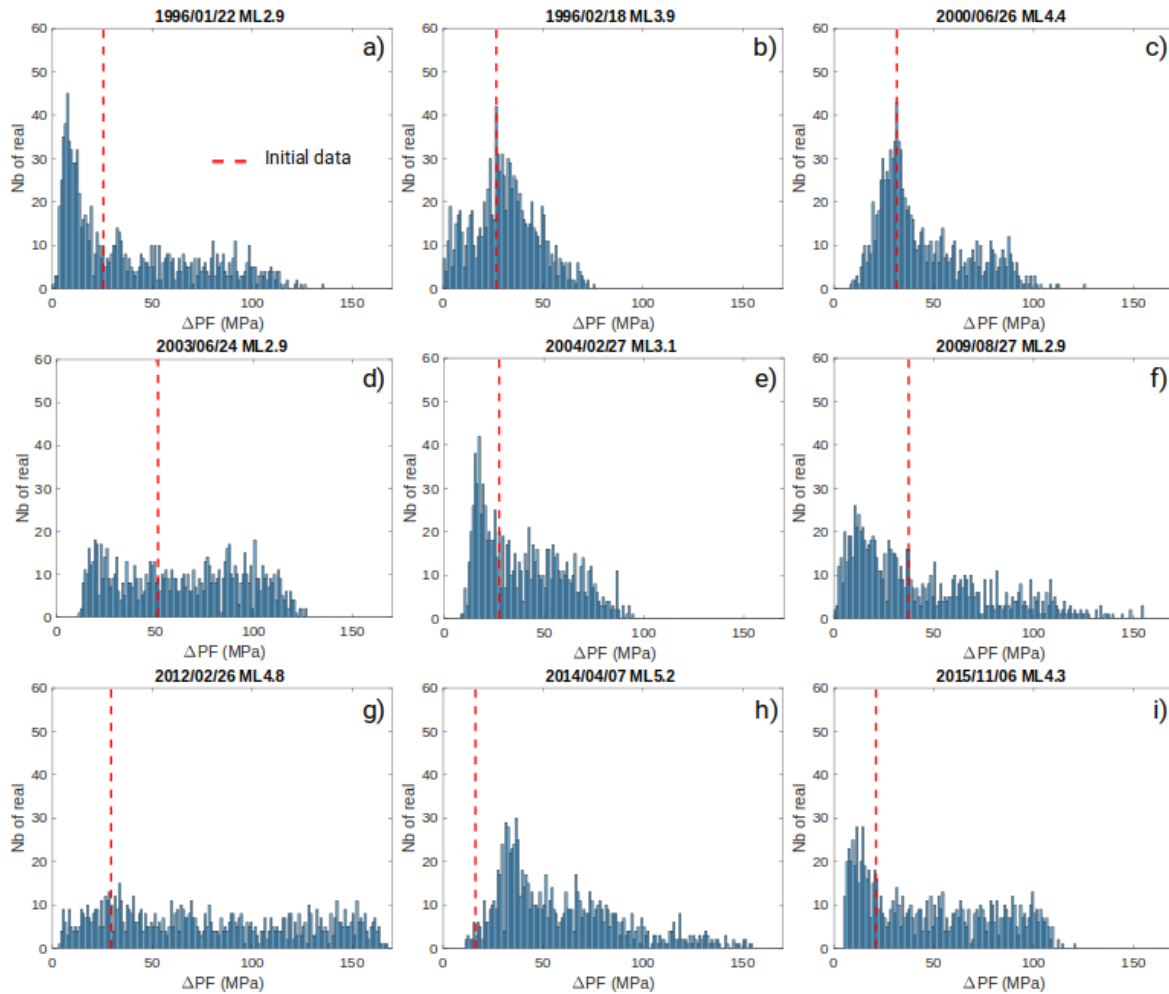
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median calculated for every 0.25 bin of magnitude with at least 5 events. The same figure, with pore-

611

fluid factor as function of magnitude is available in Supplementary Materials (Fig. S10).

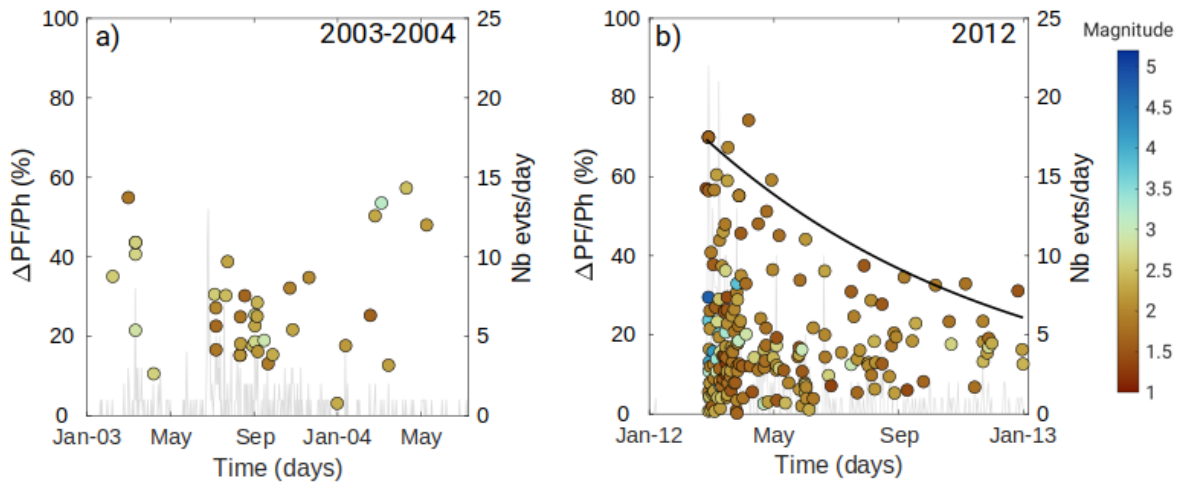
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613

614 Figure 9: Distribution of overpressure obtained by the inversion of 1000 perturbed catalogues of
 615 focal mechanisms. For each perturbed catalogue, the stress-state is first inverted before being used
 616 in the pressure calculation (see section 4). The number of realizations is the number of perturbed
 617 catalogues leading to fluid-overpressure value in a 1 MPa bin. Results are shown for 9 selected
 618 earthquakes occurring in the background seismicity (a, f), along the Serenne fault (b, c), during the
 619 2003-2004 swarm (d, e) during the 2012-2015 sequence (g to i). The local magnitude of the events is
 620 taken from the LDG catalogue (<http://www-dase.cea.fr>). The dashed-red line corresponds to the
 621 initial value of fluid-overpressure calculated using the unperturbed catalogue.

622



623

624 Figure 10: Overpressure ratio as function of time for the 2003-2004 swarm (a) and after the 2012
 625 mainshock (b). The colour corresponds to the local magnitude from the LDG (Duverger *et al.* 2021).
 626 The grey bars (right axis) show the temporal distribution of the earthquakes, from the LDG catalogue
 627 (<http://www-dase.cea.fr>). Only the events with an overpressure ratio lower than σ_3 and in the LDG
 628 catalogue are represented. In b), the black line fits the 90th percentile maximum overpressure ratio
 629 value every month following an exponential law. The same figure, with pore-fluid factor as function
 630 of time is available in Supplementary Materials (Fig. S11).

631

632

633 6. Discussion

634 From the focal mechanisms, we reconstructed the stress-state and the fluid-overpressure that trigger
 635 earthquakes for different areas and time periods in the Ubaye Region. We showed that the majority
 636 of the events need overpressure at median value of 20-30 MPa, or equivalently at 27-32% of the
 637 hydrostatic pressure. We also showed that there are no significant spatial variations of either the
 638 stress-state nor the fluid-overpressure. Even the largest events are triggered by fluid-pressure, but
 639 the time evolution of fluid-overpressure is different for a mainshock-aftershocks sequence or for a
 640 swarm sequence.

641 In this discussion, we first analyse the different sources of uncertainties that may impact our results.
642 Then, we discuss the stress state and overpressure results, in order to propose an interpretative
643 mechanism for fluid-driven aftershock seismicity and swarms.

644

645 **6.1 Source of uncertainties**

646 The main uncertainties we have to deal with come from the accuracy of the focal mechanisms, as
647 they are used to compute both the stress state and the overpressures. However, we do not have
648 precise information on the uncertainties of the focal mechanisms we used, as they come from
649 various origins. We therefore have to assume a range of uncertainties. Then, the statistical approach
650 based on the focal mechanism's perturbation (section 3.2) allows the propagation of those
651 uncertainties through the full workflow to estimate the stress-state and fluid-pressure uncertainties.
652 The error bars in Fig. 5 or the distribution of overpressures for the 1000 perturbed catalogues (Fig. 9)
653 show that the reconstructed overpressures may vary, but that the requirement of fluid pressure is
654 reliable.

655 We show that the reconstructed fluid-pressures, on average, do not show horizontal spatial
656 variations (Fig. 6). On average, they are also found constant with depth (Fig. 7) in the Ubaye Valley,
657 while they seem to increase with depth in the Serenne area. These observations are based on our
658 confidence in the location of events. However, we do not have information on location uncertainties
659 the focal mechanisms for the regional catalogues. On the contrary, the 2014 aftershocks were
660 relocated through a double-difference algorithm, with limited relative location uncertainties (De
661 Barros *et al.* 2019). Using this specific catalogue, we do not observe any horizontal spatial pattern
662 (Fig. 6c) but a constant trend with depth (Fig. 7c), similarly to the Ubaye Valley patterns. By
663 extrapolation, we may conclude that, on average, the location uncertainties do not likely bias the
664 spatial results.

665 In our analysis, we focused only on the variation of fluid-overpressure (ΔPF). However, to reach
666 failure, we could also have an increase of shear stress (overstress, $\Delta\tau$). Particularly, the aftershocks
667 that are close from the 2014 mainshock can be reactivated by Coulomb stress transfer from the
668 mainshock (Stein, 1999). In this case, the overstress should be smaller than the stress drop of the
669 mainshock, which is assumed to be close to 10 MPa (Kanamori and Anderson, 1975). From eq. 10,
670 the required overstress $\Delta\tau$ can be expressed as a function of the computed ΔPF :

$$\Delta\tau = \mu\Delta PF \quad (17)$$

671 For most of the events (~68%), the overstress that is needed is above 10 MPa. Therefore, even if part
672 of the required perturbation might be due to shear-stress increase, the reconstructed values suggest
673 that fluid-overpressure is anyway required for a large majority of events. As we could not
674 discriminate between shear-stress increase and fluid-overpressure, and for sake of simplicity, we
675 therefore focus our interpretations on fluid-driven processes.

676

677 **6.2 Stress-state**

678 The reconstructed stress-state shows a horizontal σ_3 , striking $\sim N280^\circ$, and σ_1 and σ_2 in a vertical
679 plane oriented $\sim N190^\circ$. Their plunge in this plane varies, as σ_2 is found between sub-horizontal and
680 45° dipping. Their reconstructed stress magnitude is however very close (Fig. 5), meaning that the
681 plunge variations can be easily explained, by, for example, topographical effect. In NE Japan, beneath
682 low-altitude areas, the reverse regime (σ_3 vertical) is dominant while high altitude areas show a
683 strike-slip regime (σ_2 vertical) (Yoshida et al. 2015). Similarly, the Ubaye Region shows large altitude
684 differences ($\sim 2000\text{m}$) between summits and valleys. Therefore, additional weight from high summit
685 mountains may increase the vertical stress, leading to a σ_1 plunge closer to the vertical beneath
686 summits, while it will get more horizontal beneath valleys.

687 The inversion of focal mechanisms for the different sub-areas lead to very similar stress-state, in both
688 its orientation and its amplitude. The stress-state orientation is consistent with results from other
689 studies (Fojtíková and Vavryčuk, 2018). Particularly, similar stress-states are found for the Serenne
690 area, characterized by well-identified faults and diffuse seismicity, and for the Ubaye Valley where
691 recent seismicity occurred as swarms or mixed sequence along hidden faults. Therefore, large spatial
692 heterogeneity in the stress-state cannot be used to explain the variability in the seismic behaviours,
693 even if we cannot discard very localized variations (Heap *et al.* 2010) that are below the resolution
694 we have with focal mechanisms inversion.

695 We do not observe differences in the stress-state either when using the full time period or focusing
696 on perturbed periods of dense seismic activity, such as during the 2014 aftershocks sequence. We
697 could have expected a perturbation of the stress-state because of aseismic afterslip that drives the
698 aftershocks. For example, Duboeuf *et al.* (2021) observed perturbed stress-state associated with fault
699 reactivation by fluid-injection in decametric experiments. As more than 90% of the deformation was
700 observed to be aseismic, they attributed the stress perturbation to the aseismic slip. Same conclusion
701 was reached by Schoenball *et al.* (2014) during a geothermal exploitation. Unlike aseismic slip, fluid-
702 pressure may not influence the determination of stress-state orientation. Indeed, as fluid-pressure
703 could trigger slips on faults that are badly oriented in whatever direction toward the stress-state, the
704 focal mechanisms inversion leads to an average stress-state, similar to an unperturbed one.
705 Therefore, the constant stress-state we reconstructed suggests that large aseismic slip is not a likely
706 mechanism to explain the high rate of seismicity. Even if the spatial resolution brought by the focal
707 mechanisms cannot allow us to conclude on the presence of aseismic slip at small-scale, fluid-
708 pressure seems to be the dominant process.

709

710

6.3 Fluid-pressure values

711 Several studies (Jenatton *et al.* 2007; Daniel *et al.* 2011; Leclère *et al.* 2013; Thouvenot *et al.* 2016; De
712 Barros *et al.* 2019) highlighted the role of fluid-pressure as the main driving process required to
713 explain the high rate of seismicity for the 2003-2004 swarm and the 2012-2015 aftershocks. We here
714 find that significant overpressure (above uncertainties) is required for most of the events (~77%),
715 with a median overpressure of 22 %. This observation is independent of the areas, as both Serenne
716 area, and Ubaye Valley area, show similar pressure levels, despite their different seismic behaviour.
717 This need for fluid-pressure is also seen for all types of seismic sequences, including mainshocks,
718 their following aftershocks, or swarm sequences (Figs 6d and e). On average, we find that the
719 overpressure ratio should be up to 40% and, equivalently, that the fluid-overpressure is up to 70
720 MPa, even if 80% of the events need less than 40 MPa. Other fluid-overpressure analysis was carried
721 out in the Ubaye Valley for the 2003-2004 swarm by Leclère *et al.* (2013). Using a Mohr-Coulomb
722 analysis, they obtained a required overpressure of up to 55 MPa, consistently with the ones we
723 found for the 2003-2004 period and for the other periods and areas.

724 Determination of fluid-overpressure was also done for other swarms in different contexts. The Mogul
725 swarm (2008, e.g. Ruhl *et al.* 2016) occurred in the Nevada state in a hydrothermal context. Using a
726 similar methodology as the one used here, Jansen *et al.* (2019) computed the fluid-overpressure
727 required to trigger the Mogul seismicity and confirmed their results with numerical modelling. They
728 obtained a median overpressure ratio of 42% that is more than for the full Ubaye Region (22%). This
729 difference might be linked to the presence of hot springs in the Mogul swarm area, which revealed
730 an elevated fluid-pressure level in the medium. They also found fluid-overpressure up to 30 MPa,
731 which is consistent with our values. For some events, they also found fluid-pressure higher than the
732 minimum magnitude stress component. They proposed that other processes (static stress transfer,
733 transient poroelastic stress changes, aseismic slip) should explain the high fluid-pressure values
734 obtained.

735 Fluid-driven process was also proposed for the Vogtland/Western Bohemia area that was struck by
736 several swarms (e.g. Heinze *et al.*, 2017). Kurz *et al.* (2004), using finite element modelling, calculated
737 the fluid-overpressure required to trigger the different earthquake swarms in the 1990-1997 period.
738 They estimated a need of fluid-overpressure up to 50 MPa, with some events needing overpressure
739 as high as 150 MPa, which they deemed unrealistic. Hainzl *et al.* (2012) also computed fluid-
740 overpressure for the beginning of the 2008 Western Bohemia swarm. Using stress changes and
741 hydrofracture model, they estimated a need of 20-30 MPa of fluid-overpressure. Terakawa *et al.*
742 (2010) used a similar method as Leclère *et al.* (2013) to estimate the fluid-overpressure required for
743 the Mw6.3 L'Aquila earthquake and its foreshocks-aftershocks. They found a range of fluid-
744 overpressure between 20 to 50 MPa. In this case, even if a strong earthquake was likely induced by a
745 fluid-driven process, the range of fluid-overpressure is similar to the one inferred on natural swarms
746 where the magnitude rarely exceeds ML4 (Ma and Eaton, 2009). Therefore, it seems that, for natural
747 swarms induced by fluids, fluid-overpressure with values up to ~50 MPa is observed.

748 During injection-induced swarms, the behaviour of the fluid-pressure and therefore the fluid-
749 overpressure is monitored. Baisch *et al.* (2010) observed the evolution of the fluid-pressure in a case
750 of fluid injection experiments in Soultz-sous-Forêts (France). They observed a fluid-overpressure up
751 to 18 MPa at the reservoir depth (~4km). For another experimental zone, Basel Switzerland,
752 Terakawa *et al.* (2012) retrieved a required fluid-overpressure up to 30 MPa. These values of fluid-
753 overpressure in injection-induced swarms are also similar to the ones we found here and for other
754 natural swarms (Mogul, Vogtland/Western Bohemia).

755

756 **6.4 Origin of pressurised fluids**

757 One of the main questions that remains when involving fluids in the driving processes is where they
758 come from and how they get pressurized. A first possibility is the meteoric origin. Some swarm

759 sequences (e.g. Mt Hochstaufen, Hainzl *et al.* 2006; Mogul, Jansen *et al.* 2019) were explained by the
760 infiltration of rainwater in the medium that increased the fluid-pressure and triggered the seismicity.
761 For the 2003-2004 swarm, Leclère *et al.* (2012, 2013) also proposed that meteoric fluids infiltrated in
762 the Argentera massif and then flowed to the seismic area of the 2003-2004 swarm. The difference in
763 elevation between the Argentera summits (>2500 m) and the Ubaye Valley (~1000 m) allow a fluid-
764 pressure to develop in the swarm area, assuming that faults are highly permeable with the
765 Embrunais-Ubaye nappes acting as cap-rock. This is consistent with the chemical analysis of a
766 geothermal spring located 20 km south-east of the Ubaye Valley (Bagni di Vinadio, Baietto *et al.*
767 2009; Perello *et al.* 2001). Water is shown to have a meteoric origin, before flowing up to ~5km
768 depth, close to the seismicity depth. Such a model was also used to explain the seismicity of the
769 Mogul swarm, which also occurred below a valley surrounded by high summits (Ruhl *et al.* 2016;
770 Jansen *et al.* 2019). Following this idea, we could expect to see a spatial tendency in the overpressure
771 distribution, with a decrease of overpressure with the distance to the Argentera and Pelvoux
772 summits. This is not observed, as the overpressure does not show any clear spatial tendency.
773 Moreover, as such a model requires a cap-rock to allow pressure to develop, the seismicity away
774 from the Embrunais-Ubaye nappes, as the Serenne earthquakes, should not require fluid-pressure.
775 Finally, if the Embrunais-Ubaye nappes acted as a cap-rock, the seismicity would dominantly develop
776 just below it, at ~2 km depth below sea level. This is not observed, as the seismicity mainly equally
777 develops from ~4 to 12 km below sea level.

778 Another possibility is the presence of a deep reservoir. This was proposed to explain the seismicity of
779 several swarms (Duverger *et al.* 2015, 2018; Hainzl *et al.* 2016; Jansen *et al.* 2019; De Barros *et al.*
780 2020; Ross *et al.* 2020). The reservoir is at first impermeable, then due to some perturbations (stress-
781 change, tectonic loading ...), it becomes permeable by place, which releases the fluids that then flow
782 in the medium. The fluids could migrate either upward (Duverger *et al.* 2018; Jansen *et al.* 2019) or
783 horizontally (Farrell *et al.* 2010). This is for example the main interpretation for the complex seismic
784 behaviours in the Corinth rift. The low-permeability Phyllade nappe at 6-8 km depth leads to

785 pressurized fluid beneath it. Its failure then allows fluid-pressure to diffuse, triggering swarm
786 seismicity (Duverger *et al.* 2018). In the Ubaye Valley, the reservoir needs to be of the scale of the
787 studied region, as fluid-pressure is required for the majority of the events. This might not be likely,
788 as, in this case, overpressures are expected to be at similar values and not scattered as observed.
789 Moreover, a hydrological barrier at depth is required to maintain the reservoir at high pressure, but
790 there is no geological evidence for such a layer at a regional scale. However, a smaller reservoir
791 beneath the starting location of the 2003-2004 swarms could explain it, as seismic migrations might
792 then be related to fluid flow (Shapiro *et al.* 1997).

793 A third possibility to explain the fluid-pressure is that the fluids are locally pressurized within the
794 faults, through a model of fault-valve behaviour (Sibson, 1990) or creep-compaction (Leclère *et al.*
795 2015). If the fault is sealed by a hydraulic barrier with low-permeability, the fluid-pressure tends to
796 increase from a hydrostatic pressure to a lithostatic one. This increase of pressure then promotes the
797 fault slip. Doing so, the hydraulic barrier in the fault may break, causing the over-pressurized fluids to
798 migrate upward. This model was used to explain different swarms (e.g. Cahuilla swarm, Ross *et al.*
799 2020; Southern California, Ross and Cochran, 2021; Haenam swarm, Son *et al.* 2021; Mammoth
800 Mountain, Shelly *et al.* 2015). A remaining question is the nature of the hydraulic barrier within the
801 fault. There is so far no evidence of a layer in the basement at depth that could produce a low-
802 permeability cap-rock. Leclère *et al.* (2015) suggested that creep compaction of the fault can lead to
803 impermeable barriers in the fault zone. Alternatively, the presence of clay within the faults by direct
804 chemical and physical alteration may produce hydraulic barriers. Finally, the presence of Embrunais-
805 Ubaye nappes, which lie on clay-rich materials (Lias, Trias, and Callovo-Oxfordien) may prevent a fast
806 overpressure dissipation, which may explain why seismicity organizes as swarms in the Ubaye Valley
807 and not in the crystalline massifs around (Argentera and Pelvoux). We therefore detail how such a
808 model may be applied in the Ubaye Valley in the next section.

809

6.5 Mechanisms for fluid-driven aftershock seismicity and swarms

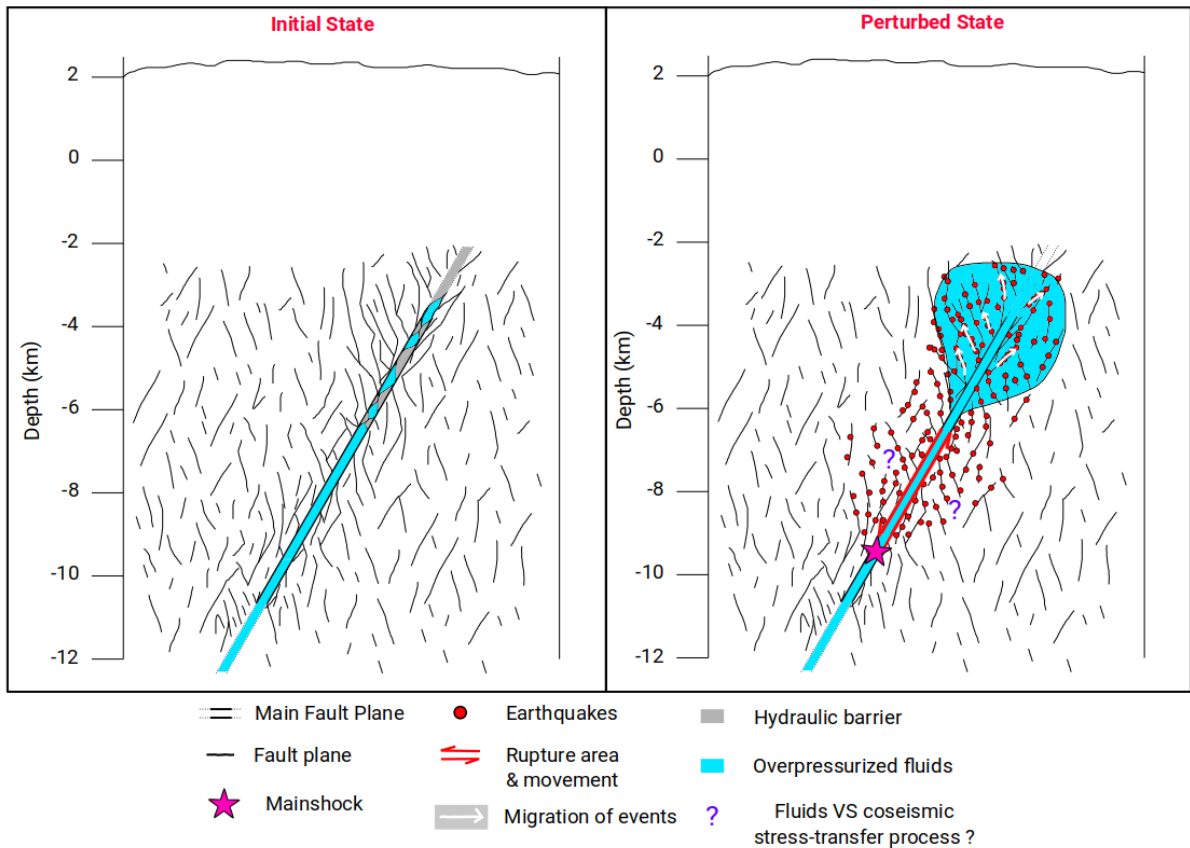
The fault valve behaviour is the model we use to explain the different types of seismicity, and particularly the high rate of aftershocks observed after the 2014 events. As the required fluid-pressure decreases with time, the mainshock is likely to initiate the fluid-pressure diffusion. Initially, the fluids are trapped by hydraulic barriers, made of clay lenses from gouge materials or mineral recrystallization, within the 2014 mainshock fault as it may tend toward lithostatic pressure (Fig. 11, Initial State). The fluid-pressure then increases as it equilibrates with the confining stress through time, and may tend toward the lithostatic pressure. This leads the faults reaching the reactivation criterion, generating the 2014 mainshock, which also breaks the hydraulic asperities. The pressure then readjusts along the rupture plane, and trigger events on smaller faults close to or above the main fault plane (Fig. 11, Perturbed State). This is consistent with a fluid-overpressure constant with depth, as observed in Fig. 7c. The fluid-pressure, together with Coulomb stress transfer, therefore participates in the triggering of the aftershocks close to the main fault plane, as fluid-pressure seems also required for some of the deep events around the mainshock (Fig. 7c). As this mechanism mainly involved fluid-pressure readjustment, with limited fluid flow, it does not imply a seismic migration with depth, which indeed was not observed in the 2014 aftershocks. Later on, in the shallow clusters, fluids flow along the reactivated planes leading to seismic migrations consistent with the observations made by De Barros et al. (2019) (Fig. 11, Perturbed state). By this mechanism, the pressure decreases and tends to equilibrate with hydrostatic pressure, leading to the decay with time of the reconstructed pressure. Globally, the very high rate of aftershocks observed during this sequence may be explained by the release of overpressure by the mainshock on top of the coseismic stress transfer.

During a swarm sequence, like in 2003-2004, the fluid-pressure does not decrease with time (Fig. 8a) and triggered seismicity for nearly two years. Moreover, seismic migration consistent with fluid

835 diffusion is observed for swarms. Therefore, a source of fluid that continuously feeds the swarm,
836 such as a reservoir, is likely beneath the swarm. By analogy with the mainshock-aftershock sequence,
837 the reservoir could be a set of fractures and faults in which fluids are initially trapped and
838 pressurized, even if our study does not bring any information on its nature. However, the initial
839 failures that then release the fluids are probably aseismic here, likely because they are at large
840 depth, below the seismogenic depth. Creep compaction associated with an aseismic slip at depth
841 (Leclère *et al.* 2015) might also be a mechanism to pressurize the fluids. In addition to pressure
842 perturbation, fluid is flowing, leading to seismic migration. In this case, the swarm sequence
843 corresponds to the last step of the model we proposed for the mainshock-aftershocks sequence,
844 similarly to the swarm behaviour sequences observed in the shallow clusters in 2014 (Fig. 11,
845 Perturbed State, top). In the 2003-2004 case, a large volume of water is supposed to flow in order to
846 maintain the swarm active for two years with a seismic migration. The volume that triggers this
847 swarm is estimated to be around $1e7 \text{ m}^3$ (Danré *et al.*, 2022b), similarly to e.g. the long duration
848 Cahuilla swarm, which was also interpreted to be driven by natural fluid-injection (Ross and Cochran,
849 2021). Assuming a reservoir made of a fracture network with a similar size as the seismic swarm, its
850 total volume will be about 32 km^3 (8 km x1 km x4 km; Daniel et al., 2011). Therefore, the fluid
851 volume ($1e7 \text{ m}^3$) required to induce this swarms may be reached by an equivalent porosity decrease
852 of only 0.03%. Such a value seems reasonable considering the porosity of a fractured fault zone that
853 may act as reservoir (Géraud, Diraison and Orellana, 2006; Jeppson, Bradbury and Evans, 2010;
854 Jeanne *et al.*, 2012). However, this process also needs to be combined with the shear-stress transfer.
855 Indeed, as mentioned by Daniel et al. (2011), the earthquake-earthquake interaction triggered 59%
856 of the seismicity for the 2003-2004 swarm, and the 41% remaining then might be associated with the
857 fluid-pressure processes. This model may also explain the 1977-1978 swarm (Fréchet and Pavoni,
858 1979) and the 1989 swarm (Guyoton et al. 1990) that occurred in the Serenne area, indicating similar
859 processes in this area.

860 In the Serenne area, fluid-overpressure is found increasing with depth, as does the hydrostatic
861 pressure. Therefore, it is likely that the pressurization occurred at different depths within the faults.
862 In our model, the seismic behaviours observed there can be similar to mainshock-aftershocks
863 sequence, where the mainshock is triggered by fluids and breaks barriers to allow pressure to diffuse
864 upward. The pressure might then trigger the aftershock seismicity, similarly to the 2014 mainshock
865 process. Moreover, in the Serenne area, the seismicity is more diffuse than in the Ubaye Valley. It
866 might mean that 1) the fluids might be locally pressurized within the faults at different depths, 2)
867 there is no fluid-pressure feeding system at depth that can trigger large swarms, and 3) fault areas
868 that can be pressurized are of smaller size or disconnected. Therefore, the main differences between
869 these two zones may be related to its fault architecture, and particularly to its hydraulic barriers.
870 Finally, the differences in behaviour between Ubaye Valley and Serenne area might be also impacted
871 by the differences in the lithology composition and by the presence of the Ubaye nappes. In
872 particular, the latter, by acting as a shallow cap-rock, may lead to a slow decay pressure
873 perturbation, hence to much longer seismic sequences in the Ubaye Valley.

874



875

876 Figure 11: Conceptual model to explain the different behaviours of the seismicity in the Ubaye Valley.

877 Left, initial state without medium perturbations other than tectonic loading. Right, perturbed state

878 showing mainshock-aftershocks behaviour (bottom) and swarm behaviour (top).

879

880

881

882 7. Conclusion

883 Based on the reconstructed focal mechanisms, we determined the stress-state and estimated the

884 overpressures required to trigger earthquakes in the Ubaye Region. We refined this analysis by

885 focusing on specific areas and periods of seismic activity. For the Ubaye Valley, the average stress

886 component orientations are N24°E, 41° for σ_1 , N201°E, 49° for σ_2 , and N293°E, sub-horizontal for σ_3 .

887 Similar values were obtained for the Serenne area, with some differences in the plunge of σ_1 and σ_2 ,
888 which may be due to topographic effects.

889 We found that most of the events (77%) are triggered by fluids, with a median overpressure of 22%.
890 Overall, the majority of the events (65%), for all sub-areas and periods, required between ~17% and
891 ~40% (~15 MPa and ~40 MPa) of overpressure, which is similar to other natural or induced swarms.
892 Even the events with the largest magnitude seem to be triggered by fluid-overpressure. Therefore,
893 mainshock-aftershocks sequences and swarms are both likely triggered by fluids. The difference in
894 behaviours do not seem to be linked to the overpressure values, but to its time evolution. While
895 overpressure decays with time in an aftershock sequence, it stays at a high level for swarms. Hence,
896 different origin of fluid-pressure may be at play. For the mainshock-aftershock sequences, we
897 proposed that the main faults get pressurized, which trigger the first, largest earthquake. Seismicity
898 is generated on shallower structures through fault-valve behaviour and stress balance. This model is
899 in accordance with the observations made by De Barros *et al.* (2019). On the contrary, swarms need
900 a constant feeding of pressure from a deeper reservoir, likely below the seismogenic depth.
901 Therefore, we here highlighted the role of fluid in both types of earthquake sequences.

902

903

904 **Acknowledgements**

905 Some of the figures were made using GMT (Wessel *et al.* 2019), and the colour scales used in the
906 different figures come from the Scientific Colour Map of Crameri (2018). The focal mechanism
907 catalogues are taken from the supplementary materials of Thouvenot *et al.* (2016), Mathey *et al.*
908 (2021a), and Mazzotti *et al.* (2021).

909 M. Baques would like to thank IRSN and CEA for her funding grants. We would also like to thank
910 Henri Leclère, which provides the focal mechanisms catalogue for the 2003-2004 swarm. We also

911 thank the CEA, which provides us the LDG catalogue for the 1965-2020 period in the studied area.
912 We thank the editors, two anonymous reviewers and F. Thouvenot for their thoughtful comments
913 that have helped to improve the manuscript.

914 **Data Availability**

915 Seismic data used for the focal mechanisms of the 2014 aftershocks are available online through the
916 RESIF portal (<http://seismology.resif.fr/>), with permanent stations on
917 <https://doi.org/10.15778/RESIF.FR> and temporary ones on <https://doi.org/10.15778/RESIF.XG2014>
918 and through the INGV portal (<http://terremoti.ingv.it/>), with permanent stations on
919 <https://doi.org/10.13127/SD/X0FXNH7QFY>. The resulting focal mechanisms catalogues, the stress-
920 states, the fluid-pressure values, and the codes developed during this paper are available in request.
921 The LDG catalogue (<http://www-dase.cea.fr>) used in this study is provided in the supplementary
922 materials.

923 **Supporting information**

924 Supplementary figures and data:

925 Supplementary_materials.docx

926 LDG_catalogue.txt

927

928

929

930 **References**

931 Amezawa, Y., Maeda, T., Kosuga, M., 2021. Migration diffusivity as a controlling factor in the duration
932 of earthquake swarms. *Earth, Planets and Space* 73, 1–11.

933 Baietto, A., Perello, P., Cadoppi, P., Martinotti, G., 2009. Alpine tectonic evolution and thermal water
934 circulations of the Argentera Massif (South-Western Alps). *Swiss Journal of Geosciences* 102, 223–
935 245.

936 Baisch, S., Rothert, E., Stang, H., Vörös, R., Koch, C., McMahon, A., 2015. Continued geothermal
937 reservoir stimulation experiments in the Cooper Basin (Australia). *Bulletin of the Seismological*
938 *Society of America* 105, 198–209.

939 Baisch, S., Vörös, R., Rothert, E., Stang, H., Jung, R., Schellschmidt, R., 2010. A numerical model for
940 fluid injection induced seismicity at Soultz-sous-Forêts. *International Journal of Rock Mechanics and*
941 *Mining Sciences* 47, 405–413.

942 Baques, M., De Barros, L., Duverger, C., Jomard, H., Godano, M., Courboux, F., Larroque, C., 2021.
943 Seismic activity in the Ubaye Region (French Alps): a specific behaviour highlighted by mainshocks
944 and swarm sequences. *Comptes Rendus. Géoscience* 353, 1–25. <https://doi.org/10.5802/crgeos.76>

945 Bertil, D., Mercury, N., Doubre, C., Lemoine, A., Van der Woerd, J., 2021. The unexpected Mayotte
946 2018–2020 seismic sequence: a reappraisal of the regional seismicity of the Comoros. *Comptes*
947 *Rendus. Géoscience* 353, 1–25.

948 Bott, M.H.P., 1959. The mechanics of oblique slip faulting. *Geological magazine* 96, 109–117.

949 Byerlee, J.D., 1970. The mechanics of stick-slip. *Tectonophysics* 9, 475–486.

950 Chiaraluce, L., Ellsworth, W.L., Chiarabba, C., Cocco, M., 2003. Imaging the complexity of an active
951 normal fault system: The 1997 Colfiorito (central Italy) case study. *Journal of Geophysical Research:*
952 *Solid Earth* 108.

953 Collettini, C., Trippetta, F., 2007. A slip tendency analysis to test mechanical and structural control on
954 aftershock rupture planes. *Earth and Planetary Science Letters* 255, 402–413.

955 Courboux, F., Dujardin, A., Vallee, M., Delouis, B., Sira, C., Deschamps, A., Honore, L., Thouvenot, F.,
956 2013. High-Frequency Directivity Effect for an Mw 4.1 Earthquake, Widely Felt by the Population in
957 Southeastern France. *Bulletin of the Seismological Society of America* 103, 3347–3353.
958 <https://doi.org/10.1785/0120130073>

959 Cox, S.F., 2016. Injection-driven swarm seismicity and permeability enhancement: Implications for
960 the dynamics of hydrothermal ore systems in high fluid-flux, overpressured faulting regimes—An
961 invited paper. *Economic Geology* 111, 559–587.

962 Cramer, F., 2018. Scientific colour maps. Zenodo. <https://doi.org/10.5281/zenodo.1243862>

963 Daniel, G., Prono, E., Renard, F., Thouvenot, F., Hainzl, S., Marsan, D., Helmstetter, A., Traversa, P.,
964 Got, J.-L., Jenatton, L., 2011. Changes in effective stress during the 2003–2004 Ubaye seismic swarm,
965 France. *Journal of Geophysical Research: Solid Earth* 116.

966 Danré, P., De Barros, L., Cappa, F., 2022. Inferring fluid volume during earthquake swarms using
967 seismic catalogs. *Geophysical Journal International*.

968 De Barros, L., Baques, M., Godano, M., Helmstetter, A., Deschamps, A., Larroque, C., Courboux, F.,
969 2019. Fluid-Induced Swarms and Coseismic Stress Transfer: A Dual Process Highlighted in the
970 Aftershock Sequence of the 7 April 2014 Earthquake (Ml 4.8, Ubaye, France). *Journal of Geophysical*
971 *Research: Solid Earth* 124, 3918–3932.

972 De Barros, L., Cappa, F., Deschamps, A., Dublanchet, P., 2020. Imbricated Aseismic Slip and Fluid
973 Diffusion Drive a Seismic Swarm in the Corinth Gulf, Greece. *Geophysical Research Letters* 47,
974 e2020GL087142.

- 975 Duboeuf, L., De Barros, L., Kakurina, M., Guglielmi, Y., Cappa, F., Valley, B., 2021. Aseismic
 976 deformations perturb the stress state and trigger induced seismicity during injection experiments.
 977 *Geophysical Journal International* 224, 1464–1475.
- 978 Dumont, T., Schwartz, S., Guillot, S., Malusa, M., Jouvent, M., Monié, P., Verly, A., 2022. Cross-
 979 propagation of the western Alpine orogen from early to late deformation stages: Evidence from the
 980 Internal Zones and implications for restoration. *Earth-Science Reviews* 104106.
- 981 Duverger, C., Godano, M., Bernard, P., Lyon-Caen, H., Lambotte, S., 2015. The 2003–2004 seismic
 982 swarm in the western Corinth rift: Evidence for a multiscale pore pressure diffusion process along a
 983 permeable fault system. *Geophysical Research Letters* 42, 7374–7382.
- 984 Duverger, C., Lambotte, S., Bernard, P., Lyon-Caen, H., Deschamps, A., Nercessian, A., 2018. Dynamics
 985 of microseismicity and its relationship with the active structures in the western Corinth Rift (Greece).
 986 *Geophysical Journal International* 215, 196–221.
- 987 Duverger, C., Mazet-Roux, G., Bollinger, L., Guilhem Trilla, A., Vallage, A., Hernandez, B., Cansi, Y.,
 988 2021. A decade of seismicity in metropolitan France (2010-2019): the CEA/LDG methodologies and
 989 observations. *BSGF - Earth Sciences Bulletin* 192.
- 990 Eyre, T.S., Zecevic, M., Salvage, R.O., Eaton, D.W., 2020. A long-lived swarm of hydraulic fracturing-
 991 induced seismicity provides evidence for aseismic slip. *Bulletin of the Seismological Society of*
 992 *America* 110, 2205–2215.
- 993 Farrell, J., Smith, R.B., Taira, T., Chang, W.-L., Puskas, C.M., 2010. Dynamics and rapid migration of
 994 the energetic 2008–2009 Yellowstone Lake earthquake swarm. *Geophysical Research Letters* 37.
- 995 Fischer, T., Horálek, J., Hrubcová, P., Vavryčuk, V., Bräuer, K., Kämpf, H., 2014. Intra-continental
 996 earthquake swarms in West-Bohemia and Vogtland: a review. *Tectonophysics* 611, 1–27.
- 997 Fojtíková, L., Vavryčuk, V., 2018. Tectonic stress regime in the 2003–2004 and 2012–2015 earthquake
 998 swarms in the Ubaye Valley, French Alps. *Pure and Applied Geophysics* 175, 1997–2008.
- 999 Fréchet, J., Pavoni, N., 1979. Etude de la sismicité de la zone Briançonnaise entre Pelvoux et
 1000 Argentera (Alpes Occidentales) à l'aide d'un réseau de stations portables. *Eclogae Geologicae*
 1001 *Helvetiae* 72, 763–779.
- 1002 Gephart, J.W., Forsyth, D.W., 1984. An improved method for determining the regional stress tensor
 1003 using earthquake focal mechanism data: application to the San Fernando earthquake sequence.
 1004 *Journal of Geophysical Research: Solid Earth* 89, 9305–9320.
- 1005 Géraud, Y., Diraison, M., Orellana, N., 2006. Fault zone geometry of a mature active normal fault: A
 1006 potential high permeability channel (Pirgaki fault, Corinth rift, Greece). *Tectonophysics* 426, 61–76.
- 1007 Godano, M., 2009. Etude théorique sur le calcul des mécanismes au foyer dans un réservoir et
 1008 application à la sismicité de la saline de Vauvert (Gard) (PhD Thesis). Université Nice Sophia Antipolis.
- 1009 Godano, M., Deschamps, A., Lambotte, S., Lyon-Caen, H., Bernard, P., Pacchiani, F., 2014. Focal
 1010 mechanisms of earthquake multiplets in the western part of the Corinth Rift (Greece): influence of
 1011 the velocity model and constraints on the geometry of the active faults. *Geophysical Journal*
 1012 *International* 197, 1660–1680.

- 1013 Goebel, T.H.W., Hosseini, S.M., Cappa, F., Hauksson, E., Ampuero, J.P., Aminzadeh, F., Saleeby, J.B.,
1014 2016. Wastewater disposal and earthquake swarm activity at the southern end of the Central Valley,
1015 California. *Geophysical Research Letters* 43, 1092–1099.
- 1016 Gratier, J.-P., Ménard, G., Arpin, R., 1989. Strain-displacement compatibility and restoration of the
1017 Chaînes Subalpines of the western Alps. *Geological Society, London, Special Publications* 45, 65–81.
- 1018 Gunatilake, T., 2023. Dynamics between earthquakes, volcanic eruptions, and geothermal energy
1019 exploitation in Japan. *Scientific Reports* 13, 4625.
- 1020 Gunatilake, T., Heinze, T., Miller, S.A., Kemna, A., 2021. Hydraulically conductive fault zone
1021 responsible for monsoon triggered earthquakes in Talala, India. *Tectonophysics* 820, 229117.
- 1022 Gunatilake, T., Miller, S.A., 2022. Spatio-temporal complexity of aftershocks in the Apennines
1023 controlled by permeability dynamics and decarbonization. *Journal of Geophysical Research: Solid*
1024 *Earth* 127, e2022JB024154.
- 1025 Guyoton, F., Fréchet, J., Thouvenot, F., 1990. La crise sismique de Janvier 1989 en Haute-Ubaye
1026 (Alpes-de-Haute-Provence, France): étude fine de la sismicité par le nouveau réseau SISMALP.
1027 *Comptes rendus de l'Académie des sciences. Série 2, Mécanique, Physique, Chimie, Sciences de*
1028 *l'univers, Sciences de la Terre* 311, 985–991.
- 1029 Hainzl, S., Fischer, T., Čermáková, H., Bachura, M., Vlček, J., 2016. Aftershocks triggered by fluid
1030 intrusion: Evidence for the aftershock sequence occurred 2014 in West Bohemia/Vogtland. *Journal of*
1031 *Geophysical Research: Solid Earth* 121, 2575–2590.
- 1032 Hainzl, S., Fischer, T., Dahm, T., 2012. Seismicity-based estimation of the driving fluid pressure in the
1033 case of swarm activity in Western Bohemia. *Geophysical Journal International* 191, 271–281.
- 1034 Hainzl, S., Kraft, T., Wassermann, J., Igel, H., Schmedes, E., 2006. Evidence for rainfall-triggered
1035 earthquake activity. *Geophysical Research Letters* 33.
- 1036 Hardebeck, J.L., Michael, A.J., 2006. Damped regional-scale stress inversions: Methodology and
1037 examples for southern California and the Coalinga aftershock sequence. *Journal of Geophysical*
1038 *Research: Solid Earth* 111.
- 1039 Heap, M.J., Faulkner, D.R., Meredith, P.G., Vinciguerra, S., 2010. Elastic moduli evolution and
1040 accompanying stress changes with increasing crack damage: implications for stress changes around
1041 fault zones and volcanoes during deformation. *Geophysical Journal International* 183, 225–236.
- 1042 Heinze, T., Hamidi, S., Galvan, B., Miller, S.A., 2017. Numerical simulation of the 2008 West-
1043 Bohemian earthquake swarm. *Tectonophysics* 694, 436–443.
- 1044 INGV Seismological Data Centre, 2006. Rete Sismica Nazionale (RSN). Istituto Nazionale di Geofisica e
1045 Vulcanologia (INGV), Italy. <https://doi.org/10.13127/SD/X0FXNH7QFY>
- 1046 Jakoubková, H., Horálek, J., Fischer, T., 2018. 2014 mainshock-aftershock activity versus earthquake
1047 swarms in West Bohemia, Czech Republic. *Pure and Applied Geophysics* 175, 109–131.
- 1048 Jansen, G., Ruhl, C.J., Miller, S.A., 2019. Fluid Pressure-Triggered Foreshock Sequence of the 2008
1049 Mogul Earthquake Sequence: Insights From Stress Inversion and Numerical Modeling. *Journal of*
1050 *Geophysical Research: Solid Earth* 124, 3744–3765.

- 1051 Jeanne, P., Guglielmi, Y., Lamarche, J., Cappa, F., Marié, L., 2012. Architectural characteristics and
1052 petrophysical properties evolution of a strike-slip fault zone in a fractured porous carbonate
1053 reservoir. *Journal of Structural Geology* 44, 93–109.
- 1054 Jenatton, L., Guiguet, R., Thouvenot, F., Daix, N., 2007. The 16,000-event 2003–2004 earthquake
1055 swarm in Ubaye (French Alps). *Journal of Geophysical Research: Solid Earth* 112.
- 1056 Jeppson, T.N., Bradbury, K.K., Evans, J.P., 2010. Geophysical properties within the San Andreas Fault
1057 Zone at the San Andreas Fault Observatory at Depth and their relationships to rock properties and
1058 fault zone structure. *Journal of Geophysical Research: Solid Earth* 115.
- 1059 Kakurina, M., Guglielmi, Y., Nussbaum, C., Valley, B., 2019. Slip perturbation during fault reactivation
1060 by a fluid injection. *Tectonophysics* 757, 140–152.
- 1061 Kanamori, H., Anderson, D.L., 1975. Theoretical basis of some empirical relations in seismology.
1062 *Bulletin of the seismological society of America* 65, 1073–1095.
- 1063 Kato, A., Saiga, A., Takeda, T., Iwasaki, T., Matsuzawa, T., 2014. Non-volcanic seismic swarm and fluid
1064 transportation driven by subduction of the Philippine Sea slab beneath the Kii Peninsula, Japan.
1065 *Earth, Planets and Space* 66, 86.
- 1066 Kaviris, G., Elias, P., Kapetanidis, V., Serpetsidaki, A., Karakonstantis, A., Plicka, V., De Barros, L.,
1067 Sokos, E., Kassaras, I., Sakkas, V., 2021. The Western Gulf of Corinth (Greece) 2020–2021 seismic
1068 crisis and cascading events: First results from the Corinth Rift Laboratory network. *The Seismic
1069 Record* 1, 85–95.
- 1070 Kerckhove, C., 1969. La 'zone du Flysch' dans les nappes de l'Embrunais-Ubaye (Alpes occidentales).
1071 *Géologie alpine* 45, 5–204.
- 1072 Kerckhove, C., Cochonat, P., Debelmas, J., 1978. Tectonique du soubassement parautochtone des
1073 nappes de l'Embrunais-Ubaye sur leur bordure occidentale, du Drac au Verdon. *Géologie Alpine,
1074 Grenoble* 54, 67–82.
- 1075 Kurz, J.H., Jahr, T., Jentzsch, G., 2004. Earthquake swarm examples and a look at the generation
1076 mechanism of the Vogtland/Western Bohemia earthquake swarms. *Physics of the Earth and
1077 Planetary Interiors* 142, 75–88.
- 1078 Larroque, C., Baize, S., Albaric, J., Jomard, H., Trévisan, J., Godano, M., Cushing, M., Deschamps, A.,
1079 Sue, C., Delouis, B., 2021. Seismotectonics of southeast France: from the Jura mountains to Corsica.
1080 *Comptes Rendus. Géoscience* 353, 1–47.
- 1081 Le Goff, B., Bertil, D., Lemoine, A., Terrier, M., 2009. Systèmes de failles de Serenne et de la Haute
1082 Durance (Hautes-Alpes): évaluation de l'aléa sismique. *Rapport BRGM RP-57659-FR* 84.
- 1083 Leclère, H., Cappa, F., Faulkner, D., Fabbri, O., Armitage, P., Blake, O., 2015. Development and
1084 maintenance of fluid overpressures in crustal fault zones by elastic compaction and implications for
1085 earthquake swarms. *Journal of Geophysical Research: Solid Earth* 120, 4450–4473.
1086 <https://doi.org/10.1002/2014JB011759>
- 1087 Leclère, H., Daniel, G., Fabbri, O., Cappa, F., Thouvenot, F., 2013. Tracking fluid pressure buildup from
1088 focal mechanisms during the 2003–2004 Ubaye seismic swarm, France. *Journal of Geophysical
1089 Research: Solid Earth* 118, 4461–4476.

- 1090 Leclère, H., Fabbri, O., Daniel, G., Cappa, F., 2012. Reactivation of a strike-slip fault by fluid
1091 overpressuring in the southwestern French-Italian Alps. *Geophysical Journal International* 189, 29–
1092 37.
- 1093 Lohman, R.B., McGuire, J.J., 2007. Earthquake swarms driven by aseismic creep in the Salton Trough,
1094 California. *Journal of Geophysical Research: Solid Earth* 112.
- 1095 Ma, S., Eaton, D.W., 2009. Anatomy of a small earthquake swarm in southern Ontario, Canada.
1096 *Seismological Research Letters* 80, 214–223.
- 1097 Martínez-Garzón, P., Kwiatek, G., Bohnhoff, M., Dresen, G., 2016. Impact of fluid injection on fracture
1098 reactivation at The Geysers geothermal field. *Journal of Geophysical Research: Solid Earth* 121, 7432–
1099 7449.
- 1100 Martínez-Garzón, P., Kwiatek, G., Ickrath, M., Bohnhoff, M., 2014. MSATSI: A MATLAB package for
1101 stress inversion combining solid classic methodology, a new simplified user-handling, and a
1102 visualization tool. *Seismological Research Letters* 85, 896–904.
- 1103 Mathey, M., Doin, M.-P., André, P., Walpersdorf, A., Baize, S., Sue, C., 2021a. Spatial Heterogeneity of
1104 Uplift Pattern in the Western European Alps Revealed by InSAR Time-Series Analysis. *Geophysical
1105 Research Letters* e2021GL095744.
- 1106 Mathey, M., Sue, C., Pagani, C., Baize, S., Walpersdorf, A., Bodin, T., Husson, L., Hannouz, E., Potin, B.,
1107 2021b. Present-day geodynamics of the Western Alps: new insights from earthquake mechanisms.
1108 *Solid Earth* 12, 1661–1681.
- 1109 Mathey, M., Walpersdorf, A., Sue, C., Baize, S., Deprez, A., 2020. Seismogenic potential of the High
1110 Durance Fault constrained by 20 yr of GNSS measurements in the Western European Alps.
1111 *Geophysical Journal International* 222, 2136–2146.
- 1112 Mattila, J., 2015. Genesis and evolution of brittle structures in Southwestern Finland and Western
1113 South Africa. *Turun yliopisto, Maantieteen ja geologian osasto, Turku.* 110s.
- 1114 Mazzotti, S., Aubagnac, C., Bollinger, L., Coca Oscanoa, K., Delouis, B., Do Paco, D., Doubre, C.,
1115 Godano, M., Jomard, H., Larroque, C., Laurendeau, A., Masson, F., Sylvander, M., Trilla, A., 2021.
1116 FMHex20: An earthquake focal mechanism database for seismotectonic analyses in metropolitan
1117 France and bordering regions. *Bulletin de la Société Géologique de France* 192.
1118 <https://doi.org/10.1051/bsgf/2020049>
- 1119 Michael, A.J., 1984. Determination of stress from slip data: faults and folds. *Journal of Geophysical
1120 Research: Solid Earth* 89, 11517–11526.
- 1121 Miller, S.A., Collettini, C., Chiaraluce, L., Cocco, M., Barchi, M., Kaus, B.J., 2004. Aftershocks driven by
1122 a high-pressure CO₂ source at depth. *Nature* 427, 724–727.
- 1123 Nishikawa, T., Ide, S., 2017. Detection of earthquake swarms at subduction zones globally: Insights
1124 into tectonic controls on swarm activity. *Journal of Geophysical Research: Solid Earth* 122, 5325–
1125 5343.
- 1126 Parotidis, M., Shapiro, S.A., 2004. A statistical model for the seismicity rate of fluid-injection-induced
1127 earthquakes. *Geophysical Research Letters* 31.

- 1128 Perello, P., Marini, L., Martinotti, G., Hunziker, J.C., 2001. The thermal circuits of the Argentera Massif
1129 (western Alps, Italy): an example of low-enthalpy geothermal resources controlled by Neogene alpine
1130 tectonics. *Eclogae Geologicae Helvetiae* 94, 75–94.
- 1131 RESIF, 1995. RESIF-RLBP French Broad-band network, RESIF-RAP strong motion network and other
1132 seismic stations in metropolitan France. RESIF - Réseau Sismologique et géodésique Français.
1133 <https://doi.org/doi:10.15778/resif.fr>'.
- 1134 Ross, Z.E., Cochran, E.S., 2021. Evidence for latent crustal fluid injection transients in Southern
1135 California from long-duration earthquake swarms. *Geophysical Research Letters* e2021GL092465.
- 1136 Ross, Z.E., Cochran, E.S., Trugman, D.T., Smith, J.D., 2020. 3D fault architecture controls the
1137 dynamism of earthquake swarms. *Science* 368, 1357–1361.
- 1138 Ruhl, C.J., Abercrombie, R.E., Smith, K.D., Zaliapin, I., 2016. Complex spatiotemporal evolution of the
1139 2008 Mw 4.9 Mogul earthquake swarm (Reno, Nevada): Interplay of fluid and faulting. *Journal of*
1140 *Geophysical Research: Solid Earth* 121, 8196–8216.
- 1141 Sanchez, G., Rolland, Y., Schreiber, D., Giannerini, G., Corsini, M., Lardeaux, J.-M., 2010. The active
1142 fault system of SW Alps. *Journal of Geodynamics* 49, 296–302.
- 1143 Schoenball, M., Dorbath, L., Gaucher, E., Wellmann, J.F., Kohl, T., 2014. Change of stress regime
1144 during geothermal reservoir stimulation. *Geophysical Research Letters* 41, 1163–1170.
- 1145 Schultz, R., Skoumal, R.J., Brudzinski, M.R., Eaton, D., Baptie, B., Ellsworth, W., 2020. Hydraulic
1146 fracturing-induced seismicity. *Reviews of Geophysics* 58, e2019RG000695.
- 1147 Shapiro, S.A., Huenges, E., Borm, G., 1997. Estimating the crust permeability from fluid-injection-
1148 induced seismic emission at the KTB site. *Geophysical Journal International* 131, F15–F18.
- 1149 Shapiro, S.A., Rothert, E., Rath, V., Rindschwentner, J., 2002. Characterization of fluid transport
1150 properties of reservoirs using induced microseismicity. *Geophysics* 67, 212–220.
- 1151 Shelly, D.R., Hill, D.P., 2011. Migrating swarms of brittle-failure earthquakes in the lower crust
1152 beneath Mammoth Mountain, California. *Geophysical Research Letters* 38.
- 1153 Shelly, D.R., Moran, S.C., Thelen, W.A., 2013. Evidence for fluid-triggered slip in the 2009 Mount
1154 Rainier, Washington earthquake swarm. *Geophysical Research Letters* 40, 1506–1512.
- 1155 Shelly, D.R., Taira, T., Prejean, S.G., Hill, D.P., Dreger, D.S., 2015. Fluid-faulting interactions: Fracture-
1156 mesh and fault-valve behavior in the February 2014 Mammoth Mountain, California, earthquake
1157 swarm. *Geophysical Research Letters* 42, 5803–5812.
- 1158 Sibson, R.H., 1990. Rupture nucleation on unfavorably oriented faults. *Bulletin of the Seismological*
1159 *Society of America* 80, 1580–1604.
- 1160 Sira, C., Schlupp, A., Schaming, M., Granet, M., 2012. Séisme de Barcelonnette du 26 Février 2012.
1161 Rapport du BCSF 43p.
- 1162 Sira, C., Schulupp, A., Schaming, M., Chesnais, C., Cornou, C., Deschamps, A., Delavaud, E., Maufroy,
1163 E., 2014. Séisme de Barcelonnette du 7 avril 2014. Rapport du BCSF.

- 1164 Son, M., Cho, C.S., Choi, J.-H., Jeon, J.-S., Park, Y.K., 2021. Spatiotemporal patterns of the 2020
1165 Haenam earthquake sequence, South Korea: lineament and migration implying fluid-driven
1166 earthquake swarm. *Geosciences Journal* 25, 19–31.
- 1167 Stein, R.S., 1999. The role of stress transfer in earthquake occurrence. *Nature* 402, 605–609.
- 1168 Sue, C., 1998. Dynamique actuelle et récente des Alpes occidentales internes: Approche structurale
1169 et sismologique (PhD Thesis). Grenoble 1.
- 1170 Sue, C., Delacou, B., Champagnac, J.-D., Allanic, C., Tricart, P., Burkhard, M., 2007. Extensional
1171 neotectonics around the bend of the Western/Central Alps: an overview. *International Journal of*
1172 *Earth Sciences* 96, 1101–1129.
- 1173 Sue, C., Tricart, P., 2003. Neogene to ongoing normal faulting in the inner western Alps: a major
1174 evolution of the late alpine tectonics. *Tectonics* 22.
- 1175 Talwani, P., Chen, L., Gahalaut, K., 2007. Seismogenic permeability, ks. *Journal of Geophysical*
1176 *Research: Solid Earth* 112.
- 1177 Terakawa, T., Miller, S.A., Deichmann, N., 2012. High fluid pressure and triggered earthquakes in the
1178 enhanced geothermal system in Basel, Switzerland. *Journal of Geophysical Research: Solid Earth* 117.
- 1179 Terakawa, T., Zoporowski, A., Galvan, B., Miller, S.A., 2010. High-pressure fluid at hypocentral depths
1180 in the L'Aquila region inferred from earthquake focal mechanisms. *Geology* 38, 995–998.
- 1181 Thouvenot, F., Jenatton, L., Scafidi, D., Turino, C., Potin, B., Ferretti, G., 2016. Encore Ubaye:
1182 Earthquake Swarms, Foreshocks, and Aftershocks in the Southern French Alps. *Bulletin of the*
1183 *Seismological Society of America* 106, 2244–2257.
- 1184 Tricart, P., 2004. From extension to transpression during the final exhumation of the Pelvoux and
1185 Argentera massifs, Western Alps. *Eclogae Geologicae Helveticae* 97, 429–439.
- 1186 Tricart, P., Schwartz, S., 2006. A north-south section across the Queyras Schistes lustrés (Piedmont
1187 zone, western Alps): Syn-collision refolding of a subduction wedge. *Eclogae Geologicae Helveticae* 99,
1188 429–442.
- 1189 University of Genoa., 1967. Regional Seismic Network of North Western Italy [Data set]. *International*
1190 *Federation of Digital Seismograph Networks*. <https://doi.org/10.7914/SN/GU>
- 1191 Utsu, T., 1961. A statistical study on the occurrence of aftershocks. *Geophys. Mag.* 30, 521–605.
- 1192 Utsu, T., Ogata, Y., 1995. The centenary of the Omori formula for a decay law of aftershock activity.
1193 *Journal of Physics of the Earth* 43, 1–33.
- 1194 Vallée, M., Nocquet, J.-M., Battaglia, J., Font, Y., Segovia, M., Regnier, M., Mothes, P., Jarrin, P.,
1195 Cisneros, D., Vaca, S., 2013. Intense interface seismicity triggered by a shallow slow slip event in the
1196 Central Ecuador subduction zone. *Journal of Geophysical Research: Solid Earth* 118, 2965–2981.
- 1197 Vavryčuk, V., Bouchaala, F., Fischer, T., 2013. High-resolution fault image from accurate locations and
1198 focal mechanisms of the 2008 swarm earthquakes in West Bohemia, Czech Republic. *Tectonophysics*
1199 590, 189–195.

- 1200 Ventura-Valentin, W., Brudzinski, M.R., 2021. Characterization of swarm and mainshock-aftershock
1201 behavior in Puerto Rico article type: Focus Section—Puerto Rico Seismicity/2020 M6. 4 Earthquake
1202 Sequence. *Seismol. Res. Lett.*
- 1203 Wallace, R.E., 1951. Geometry of shearing stress and relation to faulting. *The Journal of geology* 59,
1204 118–130.
- 1205 Walpersdorf, A., Pinget, L., Vernant, P., Sue, C., Deprez, A., team, R., 2018. Does long-term GPS in the
1206 Western Alps finally confirm earthquake mechanisms? *Tectonics* 37, 3721–3737.
- 1207 Wessel, P., Luis, J.F., Uieda, L., Scharroo, R., Wobbe, F., Smith, W.H., Tian, D., 2019. The generic
1208 mapping tools version 6. *Geochemistry, Geophysics, Geosystems* 20, 5556–5564.
- 1209 Wynants-Morel, N., Cappa, F., De Barros, L., Ampuero, J.-P., 2020. Stress perturbation from aseismic
1210 slip drives the seismic front during fluid injection in a permeable fault. *Journal of Geophysical*
1211 *Research: Solid Earth* 125, e2019JB019179.
- 1212 Yoshida, K., Hasegawa, A., Okada, T., 2015. Spatial variation of stress orientations in NE Japan
1213 revealed by dense seismic observations. *Tectonophysics* 647, 63–72.
- 1214
- 1215