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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 18<sup>th</sup> 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 m<sup>2</sup>/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).

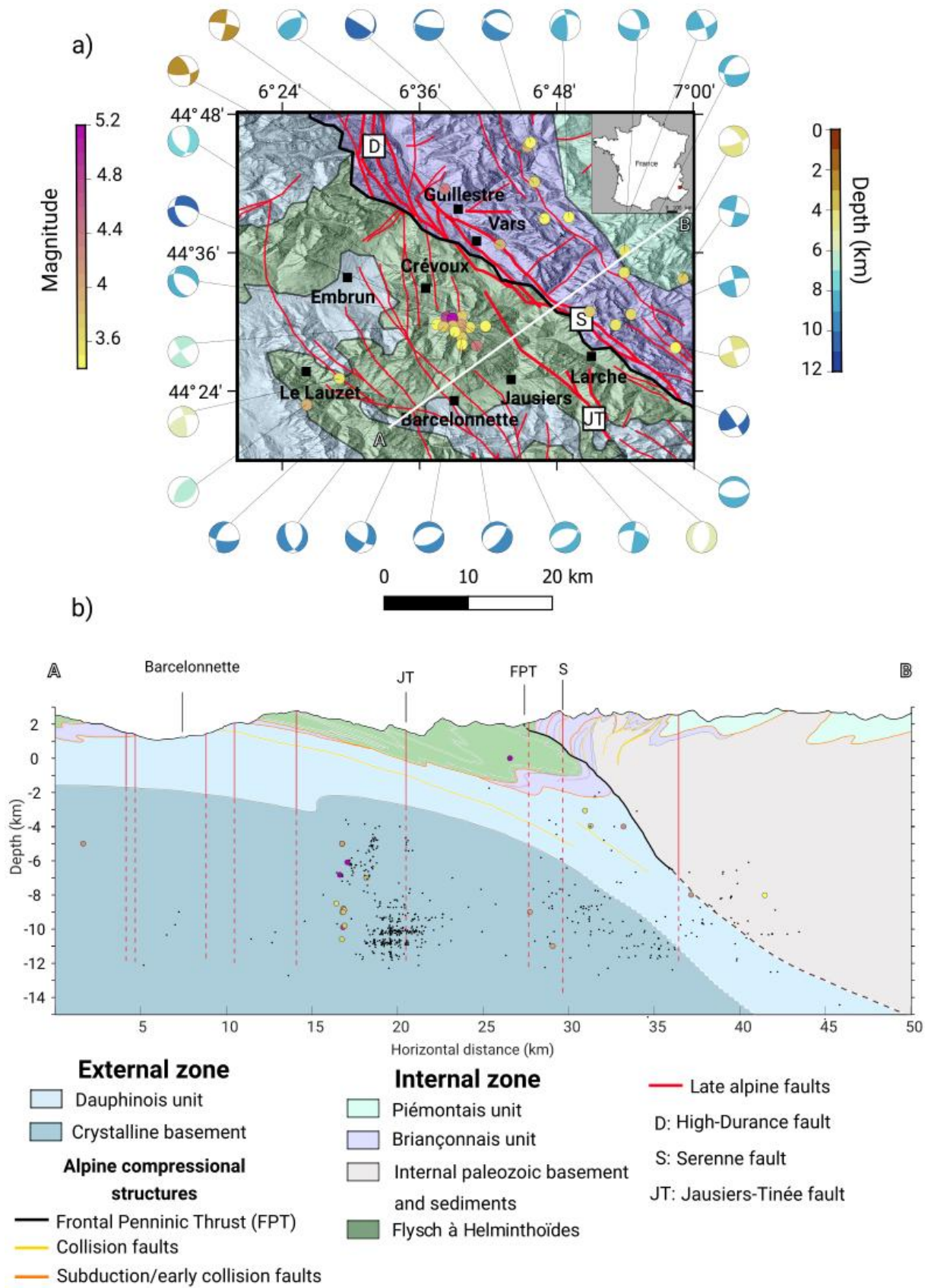
150 Crossing perpendicularly the Alpine structures, the Ubaye Valley gives us a view of the long  
151 geological history of compression and extension resulting from the setting of the Western Alps. The  
152 late Alpine faults as defined by Kerckhove (1969) and Tricart (2004) are the more recently recorded  
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



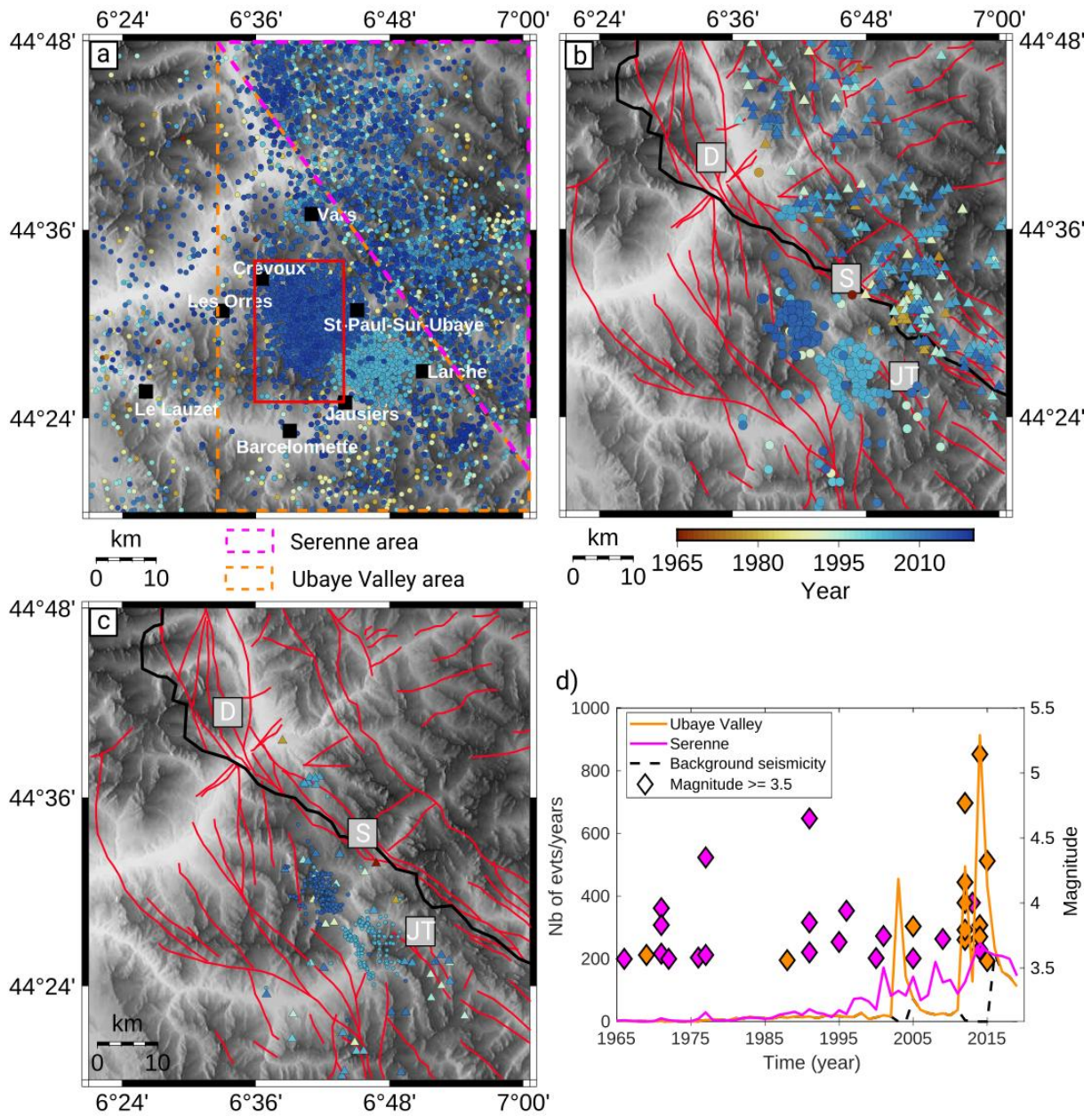
176

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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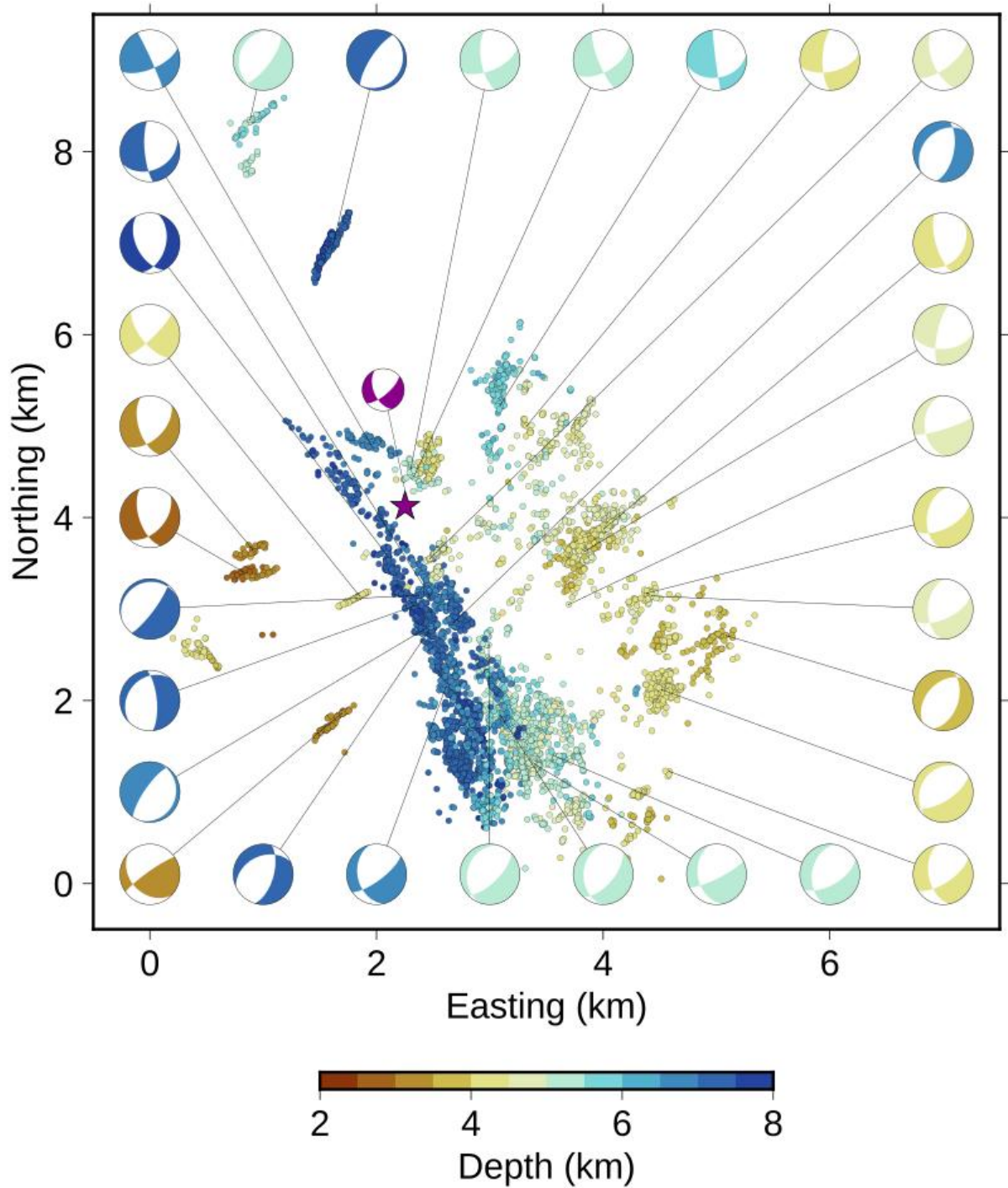
### 288 **3.2 2014 Aftershocks**

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  $\varphi$  (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 ( $\sigma_1$  and  $\sigma_3$ ) and one vertical component ( $\sigma_2$ ). However, in our case, the  
357 stress-state inversed for the 2012-2015 sequence (Fojtíková and Vavryčuk, 2018) has only one  
358 horizontal component ( $\sigma_3$ ) and two components ( $\sigma_1$  and  $\sigma_2$ ) 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  $\sigma_1$  nor  $\sigma_2$  are vertical, we first express the vertical stress ( $\sigma_v$ ) 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  $\theta_1$  and  $\theta_2$  corresponding to the angle between the vertical stress and the plunge of  $\sigma_1$  and  $\sigma_2$ .  $\rho$   
 364 is the rock density taken at  $2700 \text{ kg/m}^3$ ;  $g$  is the gravitational constant (rounded at  $10 \text{ m/s}^2$ ) 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  $\sigma_1$  and  $\sigma_2$  as function of  $\sigma_3$ :

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

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

368 We then assume that  $\sigma_3$  depends linearly on the depth ( $Z$ ), such that:

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

369 Therefore, the only unknown left to determine  $\sigma_1$  and  $\sigma_2$  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  $\tau$  and normal stress  $\sigma_n$  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  $\sigma_1$ ,  $\sigma_2$ , and  $\sigma_3$  respectively.

375 We can then define the slip-tendency  $\alpha$  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  $\alpha$  equal to 1, while it gets further from failure when  $\alpha$  decreases. If we first assume  
 378 that all faults are at a hydrostatic pressure,  $\alpha$  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  $\mu$  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  $\rho g$ , with a step of  $1 \text{ N/m}^3$ . 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,  $\sigma_1$ ,  $\sigma_2$ , and  $\sigma_3$  (eq. 4, 5, and 6), as well as the stress state  $\tau$  and  $\sigma n$  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  $40^\circ$ , but lower than  $75^\circ$ . 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  $\mu$ ,  
 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  $\tau_r$  and  $\sigma_r$  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).



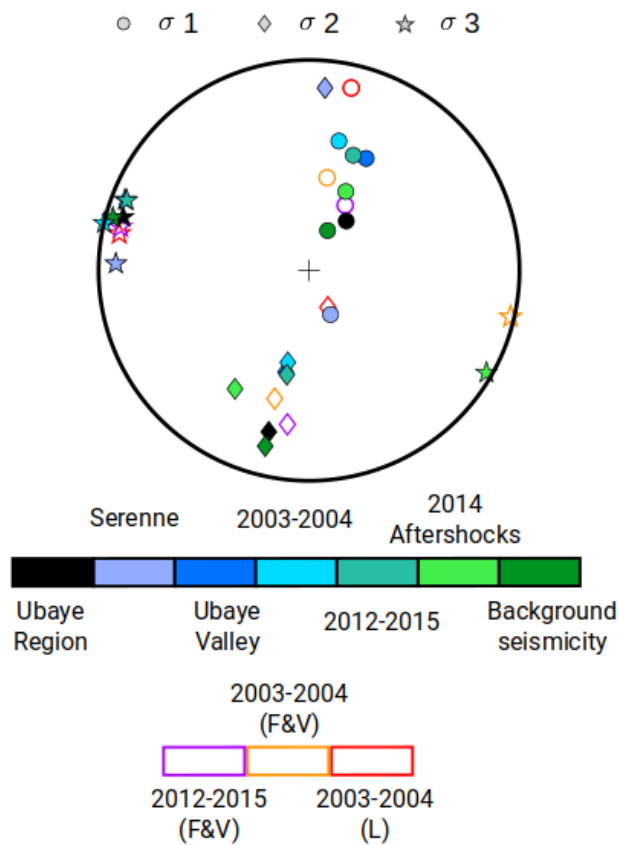
440 For the full Ubaye Region, we obtain a  $\sigma_1$  oriented  $N37^\circ \pm 8^\circ$  and plunging  $66^\circ \pm 6^\circ$ , a  $\sigma_2$  oriented  
441  $N192^\circ \pm 2^\circ$  plunging  $22^\circ \pm 6^\circ$ , and a sub-horizontal  $\sigma_3$  oriented  $N285^\circ \pm 1^\circ$ . The  $\sigma_3$  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,  $\sigma_1$  and  $\sigma_2$ , 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  $\sigma_1$  and  $\sigma_2$  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,  $\sigma_3$  orientation is found similar to the Ubaye Region one while there are  
450 differences in the plunge for the other two components. Both  $\sigma_1$  and  $\sigma_2$  are within the  $N190^\circ$  vertical  
451 plane, but  $\sigma_1$  is found subvertical ( $71^\circ$ ) in the Serenne area, while it shows a  $41^\circ$  plunge in the Ubaye  
452 Valley. Accordingly,  $\sigma_2$  is sub-horizontal ( $14^\circ$ ) and more vertical ( $49^\circ$ ) 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  $\sigma_1$  and  $\sigma_2$   
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  $\sigma_2$ , while they use the same  
464 catalogue of mechanisms as Fojtíková and Vavryčuk (2018) who found plunging of  $49^\circ$  (Tab.S2).

465 We also retrieve similar R-ratio  $\approx 0.31$ , 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  $\approx 0.29$ . This value  
 468 indicates extension and strike-slip regime in this region, which is in agreement with geodetic  
 469 observations.



470

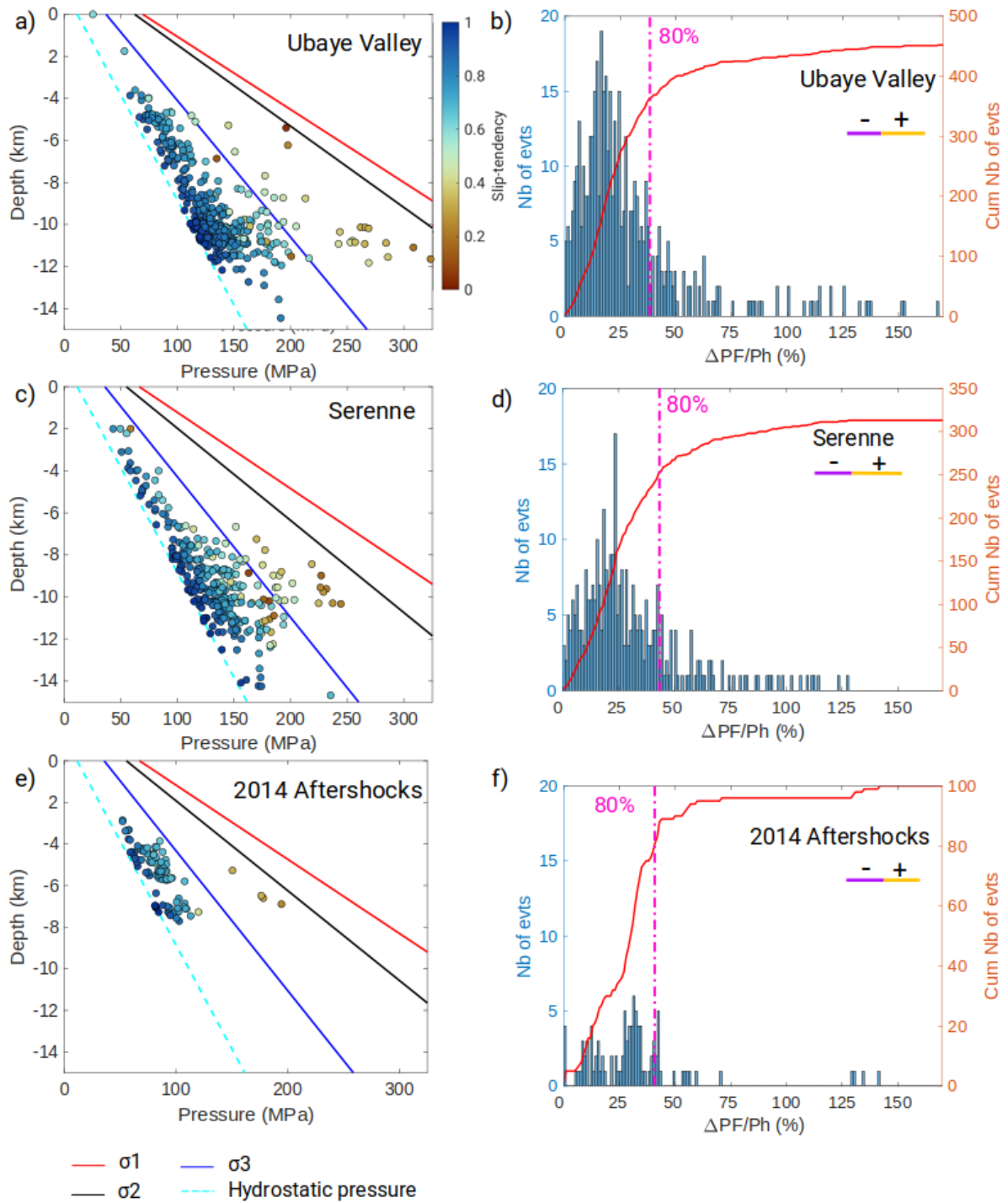
471 Figure 4: Polar view of the stress-state orientations (circle for  $\sigma_1$ , diamond for  $\sigma_2$ , and stars for  $\sigma_3$ ).  
 472 Computed stress-states are shown as filled symbols coloured by regions (full Ubaye Region, and its  
 473 two sub-areas: Serenne and Ubaye Valley, Fig. 2b); 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,  $\sigma_1$  is close to  $\sigma_2$ , 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,  $\sigma_3$ ,  $\sigma_2$ , and  $\sigma_1$ , 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).

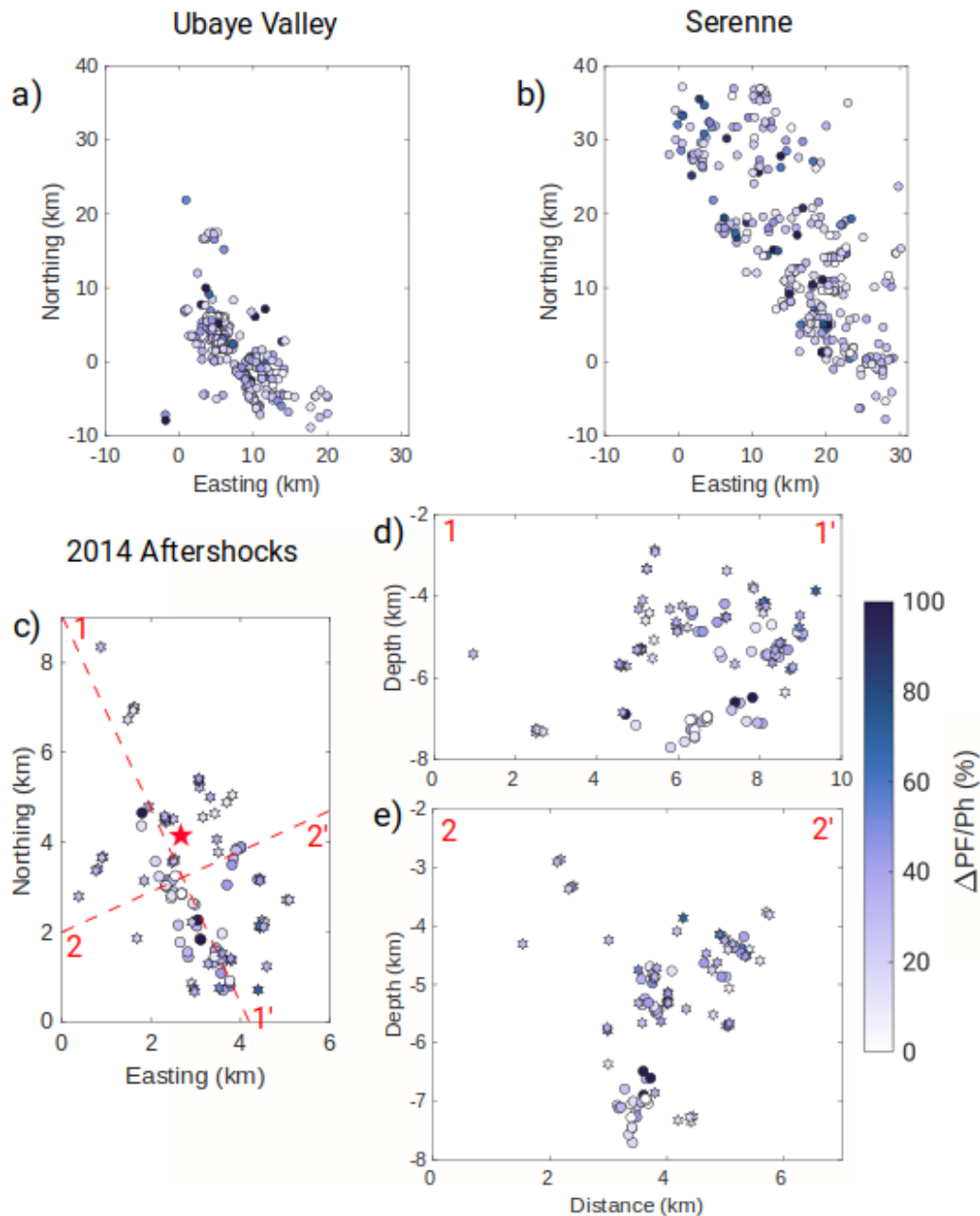
502

## 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  $\sigma_3$ . 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  $\sigma_3$  have the lowest quality index (Figs S11b and d). We then discard events with a  
512 fluid-pressure value higher than  $\sigma_3$  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

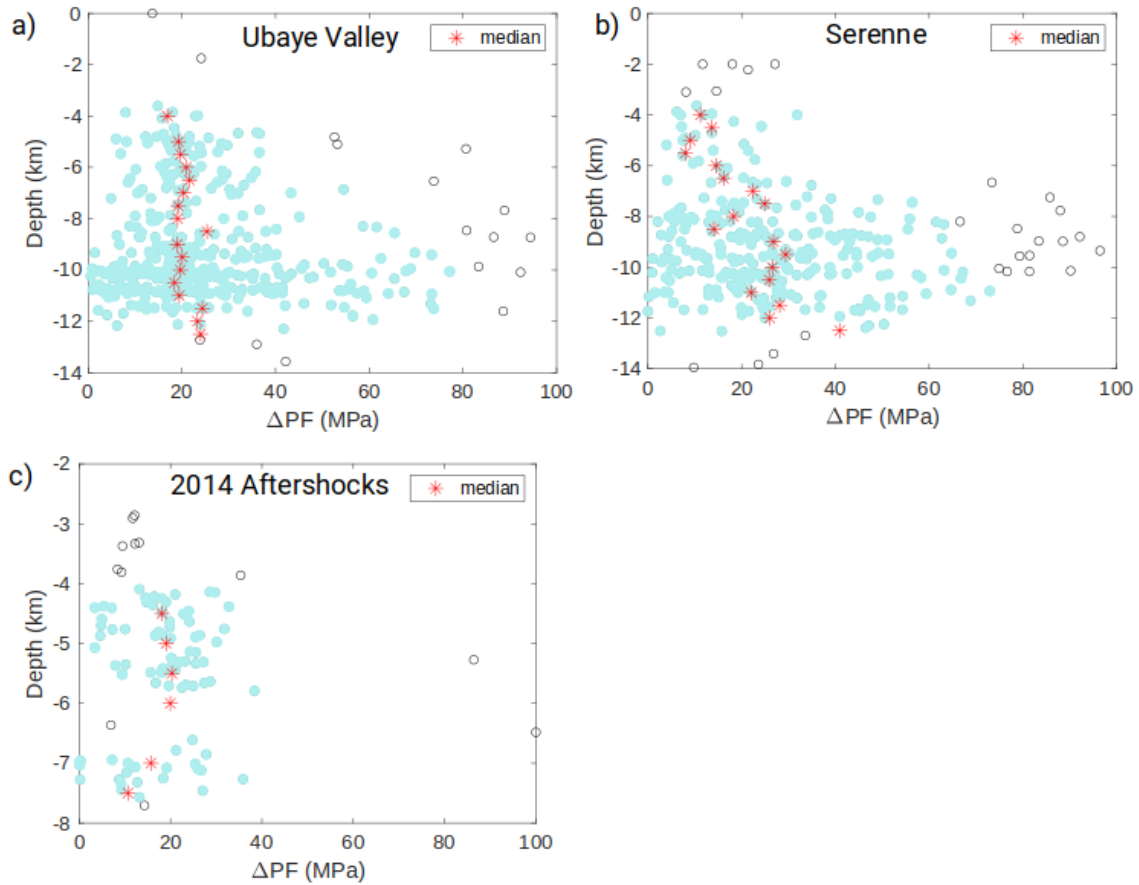
529 (Thouvenot *et al.* 2016). The dashed red-lines show the cross-sections that are used in d) and e). In  
530 c), d), and e), the hexagrams and circles correspond to events belonging to swarm-like clusters and  
531 aftershock-like clusters, respectively. The depth is below sea level. The same figure, 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  $\sigma_3$ , and are thus discarded for the median  
 559 calculation. The depth is below sea level.

560

561

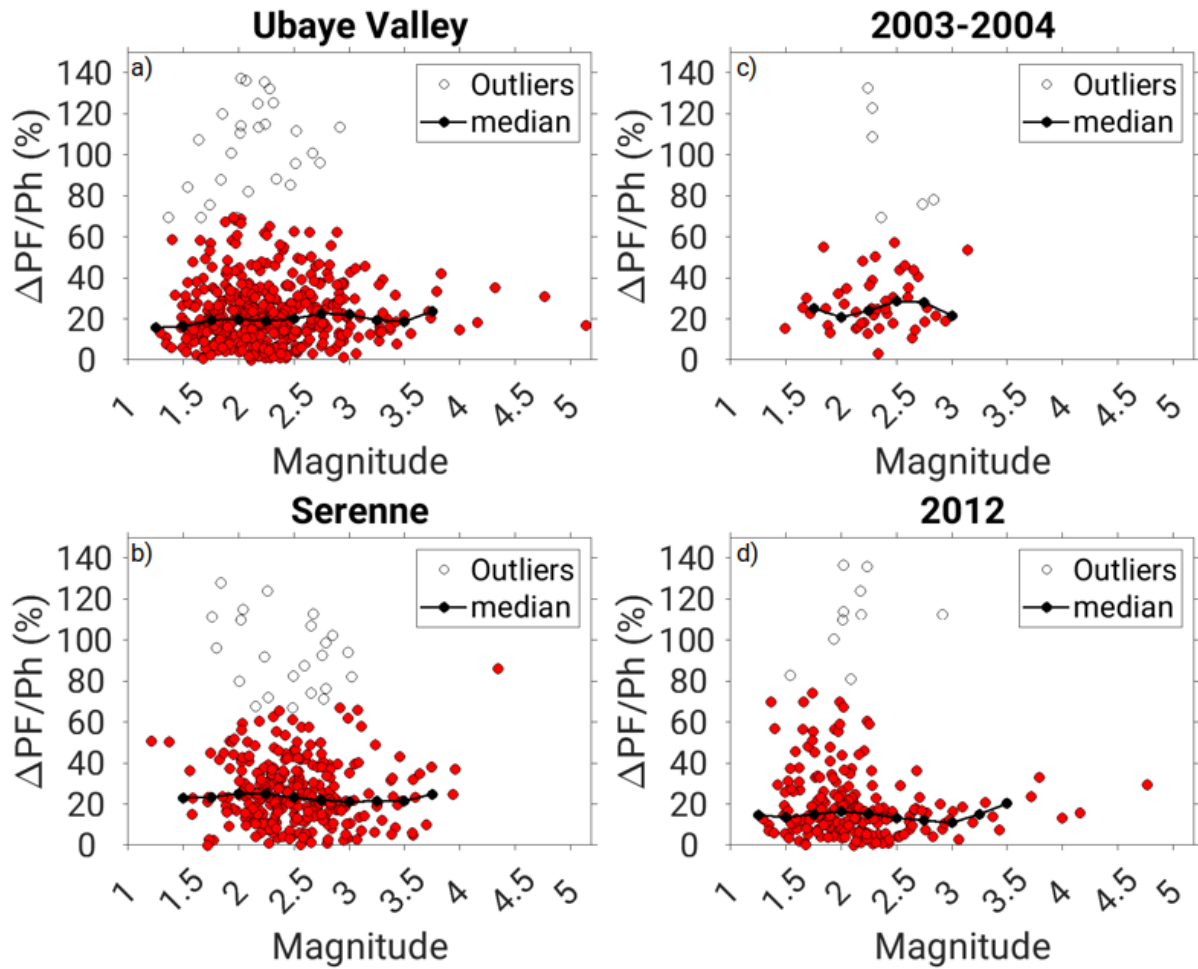
562 Fig. 8 represents the overpressure ratio as a function of the event magnitude (ML) for the different  
 563 areas and seismic sequences. For the sake of homogeneity, we use the local magnitude (ML)  
 564 computed by the LDG (Duverger *et al.* 2021) for all events in our focal mechanism catalogues. 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  $\sim 35$  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  $\sigma_3$ , which are therefore discarded from the analysis. The black line shows the floating

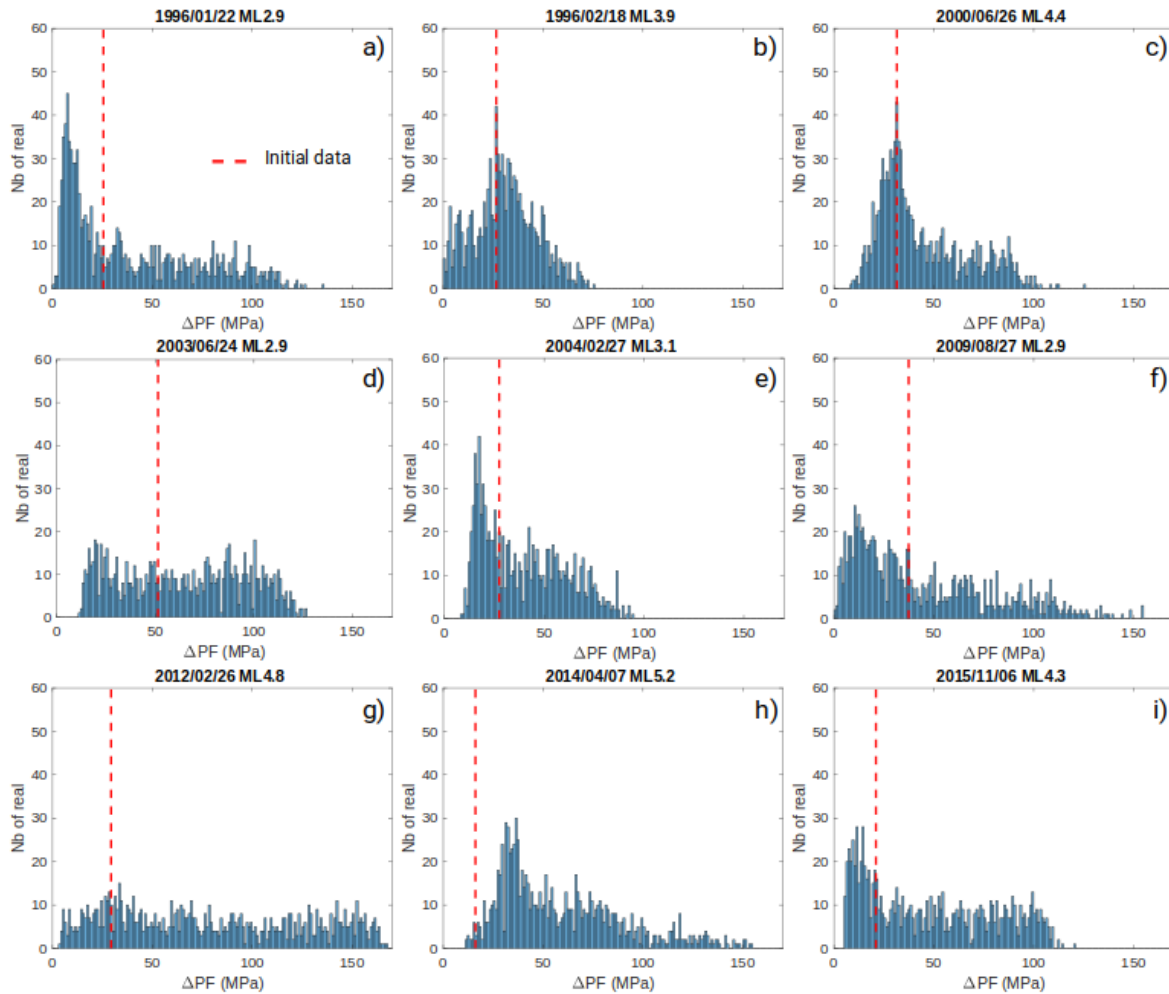
610

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).

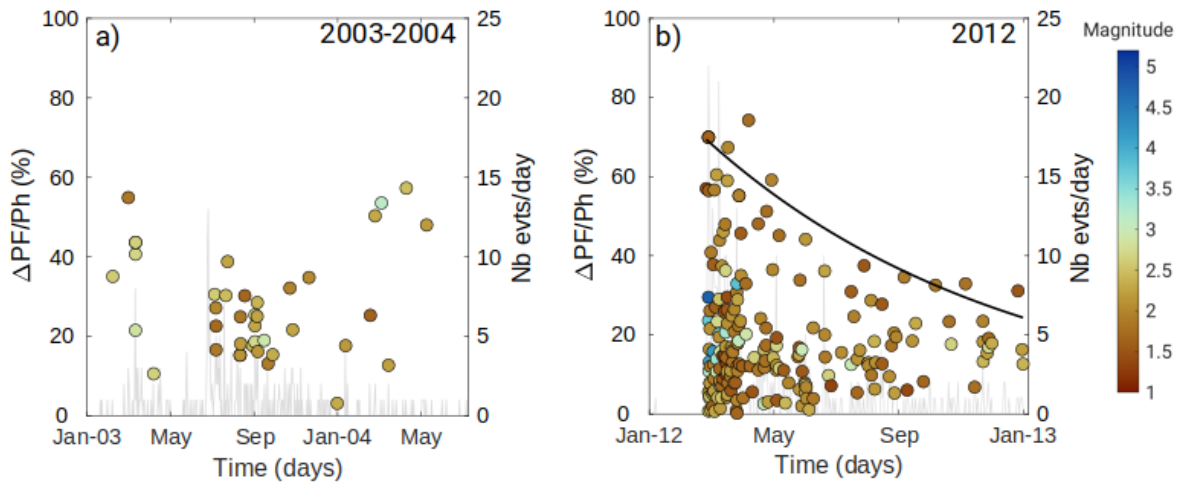
612



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  $\sigma_3$  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 ( $\Delta PF$ ). 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  $\Delta PF$ :

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

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  $\sigma_3$ , striking  $\sim N280^\circ$ , and  $\sigma_1$  and  $\sigma_2$  in a vertical  
679 plane oriented  $\sim N190^\circ$ . Their plunge in this plane varies, as  $\sigma_2$  is found between sub-horizontal and  
680  $45^\circ$  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 ( $\sigma_3$  vertical) is dominant while high altitude areas show a  
683 strike-slip regime ( $\sigma_2$  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  $\sigma_1$  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

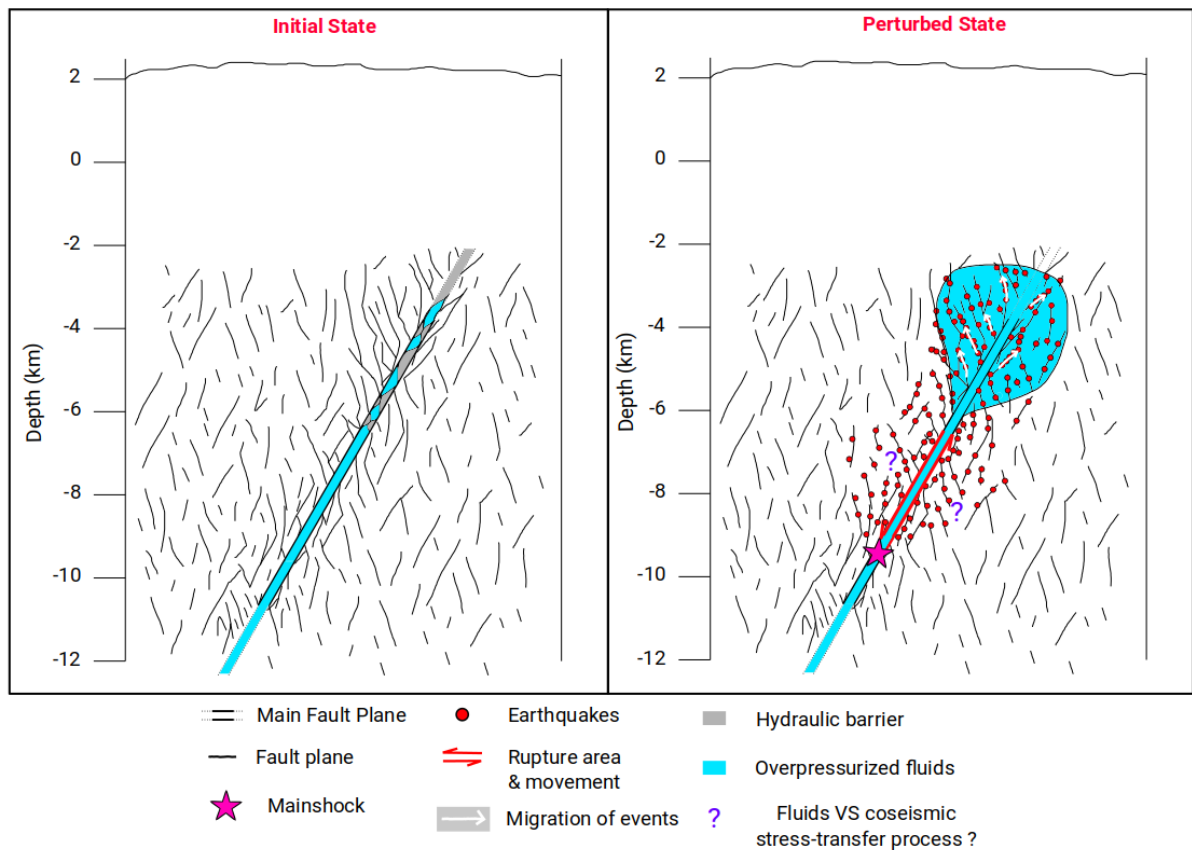
The fault valve behaviour is the model we use to explain the different types of seismicity, and particularly the high rate of aftershocks observed after the 2014 events. As the required fluid-pressure decreases with time, the mainshock is likely to initiate the fluid-pressure diffusion. Initially, the 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  $32 \text{ km}^3$  (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  $\sigma_1$ , N201°E, 49° for  $\sigma_2$ , and N293°E, sub-horizontal for  $\sigma_3$ .



887 Similar values were obtained for the Serenne area, with some differences in the plunge of  $\sigma_1$  and  $\sigma_2$ ,  
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).

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## 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

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929

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