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1 Respective roles of the weathering profile and the tectonic fractures in the structure and functioning
2 of crystalline thermo-mineral carbo-gaseous aquifers

3

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12

13 *Abbreviations:* BRGM: Bureau de Recherches Géologiques et Minières; CGG: Compagnie Générale de
14 Géophysique; CTMCG: Crystalline Thermo-Mineral and Carbo-Gaseous; EMMA: End-Member Mixing
15 Analysis; GW: GroundWater; MARTHE: Modelling of Aquifers with Rectangular grid in Transient state
16 for Hydrodynamic calculations of hEads and flows; PCA: Principal Component Analysis

17

18 1. Introduction

19 Thermo-mineral and carbo-gaseous hydrosystems are well known for their economic importance and
20 are exploited for centuries for thermal and spa activities. Sparkling natural mineral waters are
21 bottled in Europe since the early XVIIth century (Lopoukhine, 1998) and constitute a dynamic industry
22 (e.g. Renac et al., 2009; Cinti et al., 2014). These hydrosystems are associated with specific geological
23 structures, such as faults and fractures in crystalline bedrock, allowing deep (hot) fluids and/or gases
24 travelling to the surface. Case studies are scarcer in geological sedimentary context where gases
25 fluxes are often masked by the sediments. These high heat/gas flow areas are generally associated
26 with recent volcanism and/or extensional tectonics such as graben or back-arc basins (e.g. Barnes et
27 al., 1984; Kerrick et al. 1995; Matthews et al. 1987; Weinlich 2005).

28 Despite their economic importance, due to their structural complexity, crystalline thermo-mineral
29 and carbo-gaseous (CTMCG) hydrosystems are rarely characterized in detail (see nevertheless for
30 instance Maréchal et al., 2014). In the absence of a better alternative, groundwater fluxes are often
31 considered to be limited to fractures and faults related to the tectonic regime (e.g. Forster and Smith
32 1989, Stober and Bucher 1999) in hydraulically active faults whose geometry is rarely known.
33 Moreover, in many descriptions, faults are assimilated to pervious structures, often without any
34 hydrodynamic argument, even if several studies demonstrate that they are often impervious due to
35 rock crushing and various kind of sealing materials (e.g. Mohamed and Worden, 2006; Lachassagne
36 et al., 2011; Petrella et al. 2015; Tokan-Lawal et al. 2015).

37 However, groundwater flow in crystalline aquifers is not limited to tectonic fractures. Such aquifers
38 more commonly derive their hydrodynamic properties from weathering processes (Acworth, 1987;
39 Chilton and Foster 1995; Dewandel et al. 2006; Lachassagne et al., 2011), resulting in weathering
40 profiles composed of several stratiform layers. From top to bottom, where not partially or totally
41 eroded, they are i) several tens of metres thick unconsolidated saprolite layer and laminated
42 saprolite layer, together forming the “regolith”. Given its clayey–sandy composition, this superficial
43 layer may have a high porosity (up to several percents, depending on the lithology of the mother
44 rock) but a low hydraulic conductivity. ii) The underlying “fissured layer” (Lachassagne et al., 2011)
45 that constitutes the transmissive part of the aquifer. It has been shown that its fracturing is induced
46 by prolonged in-situ weathering processes, mostly as a consequence of the stresses induced by
47 certain swelling minerals such as biotite (e.g. Wyns et al., 2004; Dewandel et al., 2006; Lachassagne
48 et al. 2011). Below the weathering profile (saprolite + fissured layer), (iii) the fresh basement is only
49 permeable where discontinuities (ancient tectonic fractures, joints, veins, dykes, lithological contacts,

50 etc.) induce locally a deepening of the weathering front and, consequently, the development of some
51 pervious fractures along these ancient structures (Dewandel et al., 2011; Lachassagne et al., 2011).
52 Although the previously developed conceptual models based on groundwater flow through faults
53 and tectonics fractures only can be relevant, the hydrodynamic properties of such fractured aquifers
54 are rarely characterized in details, nor their geometry and functioning described in 3-D. Moreover,
55 fissured layers of large extension are rather a generality in shallow crystalline aquifers, and faults are
56 not always, or even rarely, permeable structures.

57 The aim of this work is to characterize in details a CTMCG hydrosystem associated with a peri-Alpine
58 graben in order to understand its geological structure and hydrogeological functioning. A
59 multidisciplinary approach is developed with various complementary methods: geological modelling
60 with geophysics and geological data from outcrops and boreholes, hydrodynamic data,
61 hydrochemistry, hydrogeological and geochemical modelling. The converging results allow
62 conceptualization of the system. The strengths of this work rely, among others, on the high density of
63 information, rarely available in such a context, and the transferability of the resultant conceptual
64 scheme.

65 **2. Material and methods**

66 **2.1. Study site. Location, geological and hydrogeological description of the study area**

67 The CTMCG system of Saint-Galmier (Saint-Galmier, France) is located at the boundary between two
68 distinct geomorphological and geological areas: the Forez plain to the West, with outcropping
69 Oligocene to Quaternary sediments, and the Hercynian crystalline massif of the Monts-du-Lyonnais
70 to the East (Figure 1), separated by a major roughly N-S oriented tectonic fault.

71 From a structural point of view, the Forez Plain is a N-S elongated graben, resulting from the
72 Cenozoic rifting associated with an early stage of the Alpine orogenic event (Figure 1a). This
73 extensive phase is highlighted by the normal faults bordering the plain on both sides (Figure 1b).

74 At Saint-Galmier, the N-S oriented border fault separates the Forez plain from the Monts-du-
75 Lyonnais plutonic and metamorphic rocks. The city of Saint-Galmier lies on porphyritic calc-alkaline
76 granite with biotite. The granite often appears weathered, both at outcrop and in boreholes, and
77 shows:

- 78 i) an often thin layer of sandy saprolite, a clay-rich material with coarse sand-size clasts, mainly
79 washed away by Plio-Quaternary erosion,
- 80 ii) a very thick laminated saprolite layer (from 15 m to more than 30 m-thick). This layer is
81 constituted by a relatively consolidated highly weathered parent rock with a coarse sand-size clasts
82 texture and a millimetre-scale dense horizontal lamination crosscutting the biggest minerals,
- 83 iii) and a fissured layer, observed from boreholes but rarely at the surface. This layer is characterized
84 by several fractured zones with a depth-decreasing density. Some of these zones are filled with
85 highly weathered materials similar to the sandy saprolite.

86 The rift-type basin of the Forez plain is a relic of the system of the Loire's Limagne that belongs to the
87 European Cenozoic peri-Alpine grabens (Figure 1a). It is filled up with Late Eocene to Miocene clayey-
88 sandy sedimentary deposits which can reach up to 800 m in the central part of the basin (Ech-Cherif
89 El Khetani, 1996; Briot et al., 2001). Deposits are continental and show a strong spatial heterogeneity
90 at the basin scale with:

- 91 i) a dominant clayey-conglomeratic and -sandy facies close to the borders and in the northern part,
92 with piedmont facies inherited from border reliefs,
- 93 ii) claystone and marl dominant facies in the central and southern areas with floodplain and lake
94 facies (Gerbe et al., 1998). Quaternary alluvium of the Loire River and its tributaries – both recent
95 and old – covering the Forez plain are masking the Tertiary sediments described above (Figure 2).

96 Several sparkling natural mineral waters (NMW) are known around the basin, the south-eastern
97 border fault of the graben (Figure 1) being the discharge point of the studied deep mineral carbo-
98 gaseous system, the Saint-Galmier natural mineral water. The historical “Fontfort” spring (see
99 location on Figure 4) was exploited since the Antiquity, and has been bottled at Saint-Galmier since
100 1837. The development of its commercial use, with the digging and exploitation of deep large

101 diameter wells, and then the drilling of borewells during the second half of the XXth Century,
102 progressively lowered the piezometric level, and dried up the spring during the XIXth century. A new
103 hydrodynamic steady state was reached during the XXth century, with now stable piezometric levels
104 since several decades at least. As it is ancient, the previous unsteady state is not documented with
105 data chronicles.

106 **2.2. Methods and data**

107 In this research, a multidisciplinary approach was applied to gather clues obtained from each
108 method. Results from these complementary methods were compiled to conceptualize the aquifer
109 structure and its functioning. These various methods are described here below.

110 a) Geological data (re)interpretation

111 Data from 254 boreholes (including 60 boreholes located in the narrow study area – Figure 1 and
112 Figure 2) were analyzed for geological and structural interpretation. Additional geological,
113 weathering and structural (e.g. fractures, joints and faults) observations were carried out on 27
114 outcrops and on a 70 m-deep cored drill (Figure 4). A simplified geological map with stratigraphic
115 succession for the basin was also established by (re)analysing and confronting bibliographic data (e.g.
116 Duclos, 1967; Duclos et al., 1974; Ech-Cherif El Kettani, 1996; Gerbe et al., 1998; Barbarin et al.,
117 2012a & b; and Couëffé et al., 2014).

118 b) Geophysics

119 Six existing seismic profiles (CGG, 1990) covering the narrow study area were reprocessed and
120 reinterpreted. A detailed geometrical analysis was performed on seismic reflectors to identify (i) the
121 main stratigraphic unconformities and (ii) the vertical displacement of successive reflectors which are
122 assumed to highlight tectonic discontinuities.

123 Stratigraphic attributions of seismic units were assigned on the basis of geometrical relationships
124 between reflectors and validated by the calibration with sedimentary succession recognized at near
125 subsurface in closest deep boreholes (17 boreholes with depth ranging between 70 and 350 m).
126 Additionally, check-shot survey on Frarie borehole and analysis of P-wave velocities on samples (from
127 6 outcrops and the cored-well) were used to define mean interval velocities of each seismic unit and
128 to calibrate the conversion from time seismic to depth geological sections.

129 The analysis of reflectors located below Cenozoic deposits highlights the existence of distinctive
130 internal seismic units into the St-Galmier granite weathered layers. These distinctive units are
131 characterized by reflection with bedded geometries that do not display sedimentary features. Their
132 estimated thicknesses were compared to those defined from saprolite and fissured layers from
133 borehole lithologs. The good consistency between thicknesses leads to consider that the
134 distinguished of seismic units below Cenozoic sedimentary cover can be attributed with the St-
135 Galmier granite layers of the weathering profile. Additional geophysical data (electrical resistivity
136 profiles, gravimetric maps) were also used to constraint the geological scheme.

137 c) 3D-geological and structural model

138 Borehole logs and seismic profiles reinterpretation allowed identification of the main
139 lithostratigraphical and structural elements of the area; i.e. faults, stratigraphic succession, geometry
140 of sedimentary deposits, paleo-morphology of the Paleozoic substratum and geometry of its
141 weathered layers. Two areas (Figures 1 and 2) were used for the construction of the geological
142 model: the “narrow study area”, with a high density of data, and an “extended area” with scarcer
143 data, but necessary to define a precise scheme at the boundaries of the “narrow area”.

144 These main geological and structural elements were gathered in a 3-D geometrical numeric model
145 using the Geomodeller© software developed by BRGM and Intrepid Geophysics. Geomodeller© uses
146 the co-kriging method for spatial interpolation (Lajaunie et al., 1997; Calcagno et al., 2006 ;
147 Courrioux et al., 2006 ; Calcagno et al., 2008).

148 d) Hydrodynamics: pumping tests and airlift measurements

149 A total of 36 pumping tests is available within the “narrow area” (see Fig. 6), 24 were performed in
150 the granite (in 18 different boreholes) and 12 in the Cenozoic deposits (11 boreholes). Test durations
151 range from a few days to several months with flowrates between 1.2 and 13 m³/h in the Tertiary
152 sediments and between 1.8 and 40 m³/h in the granite.

153 Data interpretation was conducted in two phases: analysis of the tests and modelling. After
154 normalizing the drawdown curves according to pumping rate changes, the log–log diagnostic plot
155 developed by Bourdet et al. (1983) was used. This analysis of derivative drawdown facilitates the
156 identification of a conceptual model without preliminary or “a priori” hypothesis on the aquifer’s
157 structure. Derivative curves were computed numerically based on the algorithm proposed by
158 Bourdet et al. (1989) and type curves were used for determining flow regimes (Bourdet et al., 1983,
159 1989; Ehlig-Economides, 1988; Deruyck et al., 1992; Spane and Wurster, 1993; Renard, 2005;
160 Renard et al., 2009; Dewandel et al., 2014). The diagnosis allows identifying the properties of the
161 borehole (e.g. borehole storage, skin effects), of the aquifer (e.g. isotropy/anisotropy, fractures, dual
162 porosity), its geometry (boundaries of various types), the possible relationships with other aquifers
163 (e.g. leakage), etc. Then, the mathematical model respecting the identified flow-regimes was used to
164 compute drawdown and to estimate hydrodynamic parameters. To this end a simultaneous double
165 fitting on both drawdown and derivative drawdown is necessary. This methodology has been, for
166 example, successfully applied for the hydrodynamic characterization of hard rock aquifers (Dewandel
167 et al., 2011, 2014; Roques et al., 2014; Maréchal et al., 2014).

168 According to the diagnoses of the Saint-Galmier’s pumping tests (see the “results” section), two
169 horizontal multi-layers aquifer models that respect the identified flow-regimes were used: one
170 considers that the piezometric variation in the overlying aquifer due to leakage is negligible (model
171 derived from Hantush, 1961); the other considers that the piezometric level decreases in the upper
172 aquifer because of pumping in the deeper aquifer (model derived from Hunt and Scott, 2007). As
173 diagnoses showed evidences of no-flow boundaries, the models also consider up to 4 orthogonal no-
174 flow boundaries (i.e. up to a laterally closed reservoir). A compartmentalized aquifer model, i.e. the
175 pumped aquifer is laterally limited by two other aquifers with various hydrodynamic properties
176 (Dewandel et al., 2014), was also used.

177 Air-lift flow measurements carried out during the drilling of down-to-the-hole-hammer boreholes
178 were treated statistically particularly to provide the depth of productive fracture zones (Maréchal et
179 al., 2004; Dewandel et al., 2005; Courtois et al., 2010, Roques et al., 2016). We compiled air-lift data
180 from 29 boreholes (measurements every 3 metres in average) and also added the depth of
181 productive fracture zones deduced from flowmeter measurements (every meter in average) in 4
182 other boreholes from which air-lift flows were not available. 195 productive zones were identified.
183 The discharge and location of the productive zones were referenced with respect to the interface
184 between the saprolite and the granite’s underlying fissured layer in order to explore if there is any
185 vertical trend of productive fractures distribution in the fissured layer (e.g. Courtois et al., 2010).

186 **e) Piezometric level and discharge monitoring**

187 As the natural mineral water is pumped for bottling, piezometric level and borehole pumping rate
188 are recorded at a hourly time step in the 10 abstraction boreholes, equipped either within the
189 granite or within the Tertiary sedimentary aquifers, since their construction, and in 14 unpumped
190 observation boreholes since 2010 (although previous monitoring was also existing). In the granite, six
191 nested piezometers in the weathering profile were drilled and equipped in 2009. Several campaigns
192 of piezometric measurements, including non-monitored wells, are also available across the studied
193 area. Discharge of the Coise stream is also measured at a hourly time step.

194 **f) Geochemistry**

195 Hydrochemical parameters of the abstraction boreholes are monitored weekly. In addition,
196 groundwater punctual analyses for major ions are available from 1962 to 1998 on 67 other boreholes
197 from the Saint-Galmier area, and 10 additional mineralized water samples at the scale from the Forez
198 plain area. A sampling campaign was also performed in 2010: 22 groundwater and surface water
199 (Coise stream) samples were analysed for major ions. This geochemical database was used to define
200 the Saint-Galmier NMW geochemical facies and its main characteristics as compared to other NMW
201 types found locally and regionally. Water type classification was performed using Piper diagrams,
202 biplots, Principal Component Analysis (PCA) and End-Member Mixing Analysis (EMMA) following the
203 Christophersen and Hooper’s method (1992). Distinction among groundwater characteristics was
204 defined using the computation of the sodium-pole indicator: Sodium-pole (%) = $\frac{[HCO_3^-]}{[HCO_3^-] + [Ca^{2+}]}$

205 $[\text{Mg}^{2+}]/[\text{HCO}_3^-]$ with element concentrations expressed in meq/L. This indicator, varying between 0
206 and 1, allows a clear distinction between Na and Ca groundwater types, regardless of the total
207 mineralization and the HCO_3^- content.

208 EMMA was used to determine proportions of end-members contributing to the NMW pumped in the
209 abstraction boreholes. 580 data (major ions) from January 2000 to December 2012 were used. For
210 each abstraction borehole, the hydrochemical time series and the potential end-members, i.e. water
211 samples from the geochemical database, were standardized for all major ions using the mean and
212 standard deviation of the hydrochemical time series. The main steps can be summed up as follow
213 (see Burns et al., 2001 for details), each water sample being described in n -dimensions, n being the
214 number of ions used for the EMMA: i. From these n -dimensions, a PCA is used to find a new 2-D
215 space in which most of the variance of the chemistry dataset is kept; ii. The potential end-members
216 are plotted in this new space, which allows graphically selection of up to 3 water samples to
217 characterise the water chemistry pumped in the exploited borehole; iii. The relative position of these
218 3 end-members compared to the water samples from the abstraction borehole allows quantifying 3
219 mixing ratios, which are used to re-compute the series of the n major ions; iv. A statistical (least
220 square linear regression) goodness of fit is computed to validate the choice of the end-members. If
221 needed, steps 2 and 3 are repeated until the best fit is obtained.

222 From these approaches, different poles of groundwater were identified and indicators were
223 developed to allow mixing rate quantification within each groundwater body.

224 **g) Numerical modelling of an aquifer compartment**

225 A deterministic hydrogeological model of the best known aquifer compartment (A3) was developed
226 using the MARTHE_7.4© BRGM finite differences code (Thiéry, 1990, 2010). The 2D multilayer model
227 comprises the four main layers i.e. the sedimentary deposits, a confining layer, the saprolite and the
228 fractured layers of the granite). The modeled area is a rectangle of 2.05 km by 1.66 km (Figure 12)
229 composed of 1353 square cells (50 x50 m) per layer. The model solves in steady state both
230 hydrodynamic and mass transport equations. Boundary conditions were defined according to the
231 available knowledge synthesized in the conceptual model. The main objective of this modelling is to
232 validate the hydrogeological conceptual model for this compartment, to confirm and spatialize the
233 geochemical results about mixing rates, and to locate the deep fractures providing deep highly
234 mineralized fluid.

235 **h) Synthesis and hydrogeological conceptual model**

236 All the information obtained from these various approaches was compiled and synthesized in a
237 realistic hydrogeological conceptual model that explains the geological structure and hydrogeological
238 functioning of the aquifer system. Each method provides clues which individually may not be
239 significant but, when converging towards a similar scheme, they allow building up a consistent
240 hydrogeological conceptual model.

241 **3. Results and interpretation**

242 **3.1. Geological model**

243 **3.1.1. Lithological model**

244 The main geological layers described in the lithological model are i) Quaternary deposits (alluvium
245 and colluvium); ii) Cenozoic sedimentary deposits with distinction between syn-rift and post-rift
246 units; weathered granite including iii) the saprolite (sandy regolith and laminated layer) and iv) the
247 underlying fissured layer.

248 The geological map (Feybesse et al., 1996) shows the stratigraphic sequence from top to bottom:
249 i) the recent and ancient Quaternary alluvium and colluvium covering the Forez plain until the
250 foothills of the crystalline massifs of the Monts du Lyonnais and ii) the Cenozoic deposits of the post-
251 rift and syn-rift series (see below their description), the first ones forming a narrow band along the
252 contact between the Forez plain and the Monts du Lyonnais (Figure 2), iii) the crystalline massifs of
253 the Monts du Lyonnais mainly formed by the plutonic St-Galmier's granite and its weathering facies.

254 The seismic profiles (Figure 3), calibrated using borehole geological logs, cores and P-waves
255 measurements, were used to locate the thickness of the sedimentary cover and to discriminate the

256 unconsolidated saprolite from the laminated layer within the sandy regolith. The main fractured zone
257 within the fissured layer (upper part of the fissured layer) was readily identified on seismic profiles
258 (Figure 3). In the Cenozoic sedimentary deposits an internal unconformity separating syn-rift and
259 post-rift series (Duboeuf et al.,1991) divides the Cenozoic sediments into two units with differing
260 hydraulic properties: (i) a low hydraulic conductivity mostly clayey Late Eocene - Oligocene “synrift”
261 series and ii) a Miocene (post-Aquitainian) “post-rift” series characterized by a higher hydraulic
262 conductivity due to its mostly clayey-sandy to sandy composition.

263 **3.1.2. Structural model**

264 From the reinterpretation of geological and geophysical data, the main structural elements identified
265 and located are: i) mainly normal faults highlighted by shifts of seismic reflectors in the synrift series
266 that do not affect post-rift deposits; ii) strike-slip accidents, inducing only small shifts on seismic
267 reflectors and characterized by complex geometries ascribed to flower structures; iii) depocenters of
268 the synrift deposits; iv) positive palaeomorphologies of the top of the Paleozoic bedrock, often
269 corresponding to hanging wall related to main normal faults; v) zones of higher fracture density in
270 the granite (for which reflectors belonging to the weathering profile are more discontinuous and of
271 lower frequency).

272 A structural model was built locating the two main types of faults (Figure 4): i) normal faults sealed
273 by the post-rift deposits and ii) on the eastern edge, normal faults affecting both the synrift and post-
274 rift series also related to the Cenozoic extension but being lately reactivated.

275 **3.1.3. The 3-D geological model**

276 The 3-D geological model, built with the Geomodeller© software (Figure 4), allowed the spatial
277 representation of the faults and the isohypses of the main geological layers and interfaces, i.e., from
278 bottom to top: undifferentiated unweathered granite and fissured layer, laminated layer,
279 unconsolidated saprolite, syn-rift and post-rift deposits. Quaternary sediments, as they are very thin,
280 were integrated in the post-rift layer. From a hydrogeological standpoint, this model informs the
281 depth and geometry (top) of the potential aquifers in the granite and location of faults that may
282 convey deep mineral water and CO₂ to these aquifers. The model shows where aquifers (the granite’s
283 fissured layer, aquifers within Cenozoic layers) can laterally be placed by faults opposite to
284 impervious formations. Finally, it allows assessing the spatial extension of the low hydraulic
285 conductivity syn-rift Cenozoic deposits that may act as aquitards as well as the spatial extension of
286 post rift sediments that may allow a connection between granite aquifers and the subsurface.

287 **3.2. Hydrodynamics**

288 **a) Interpretation of pumping tests**

289 Data from 36 pumping tests were analysed and modelled. To illustrate, Figure 5 shows the type
290 curves and derivative curves of pumping tests performed on the A3 borehole as well as those
291 corresponding to A1 and A2 boreholes located at 42.5 m and 19.7 m from A3, respectively. These
292 three wells are located in the granite under the sedimentary cover (about 30 m thick) and are
293 hydraulically connected. The thickness of the saprolite is variable, 41 m for A1, 10 m for A2 and 52 m
294 in A3. Drawdown curves are shifted because of the well effects of each borehole (wellbore, skin,
295 head losses). However their derivatives are characterized with a similar pattern, demonstrating that
296 these three boreholes belong to the same granite aquifer. The succession of flow behaviours –
297 (i) radial flow, (ii) linear flow, (iii) leakage, and (iv) pseudo-steady state flow with a slope of 1 -,
298 respectively shows (i) that the granite aquifer has a transmissivity ranging between 1.4 and 1.8x10⁻³
299 m²/s, (ii)&(iv) that it is limited in space by several impervious limits and that its shape is a priori
300 “rectangular” and relatively elongated. Additionally (iii) it shows evidence of leakage quite surely
301 from an overlying aquitard (saprolite and/or post-rift sediments). The used model (Hunt and Scott
302 (2007) model with no-flow boundaries) successfully matches the observed drawdowns.

303 Derivative curves from 34 tests, regardless of the geological formation, exhibit similar behaviours i.e.
304 an aquifer limited in space by 2 to 4 impervious limits (symbols B, C and D in Figure 6). Depending on
305 the duration of the tests and on the distance to the limits, all limits were not necessarily identified, as
306 well as leakage effects. Only one test (Charpinière borehole; Cenozoic deposits) shows a different
307 diagnosis: a laterally compartmentalized aquifer.

308 The bore wells where 4 impervious limits were reached belong to aquifer “compartments” which
309 surface area was then estimated (with an accuracy depending on the reliability of the estimate of the
310 storage coefficient). All computed surface areas are of the order of one square kilometre. Each of
311 these so-identified no-flow boundaries was then successfully attributed to one of the faults from the
312 structural scheme (Figure 6). They appear to correspond with the graben faults, i.e. N-S and N130
313 (Figure 9). Therefore, these faults appear to have a low permeability. This is due either to impervious
314 fault gouge, or because of the contact between permeable (fissured granite) and low-permeability
315 formations (Cenozoic deposits, saprolite), due to fault throws.

316 **b) Air-lift data analysis**

317 Figure 7a shows the average air-lift flow and flowmeter measurements from boreholes, according to
318 the depth below the bottom of the saprolite. No flowrate was observed in the saprolite layer. Figure
319 7b provides the density of productive fracture zones according to the depth. The percentage of
320 available data according to depth is also presented (Figure 7c). Statistics from Figures 7a and 7b are
321 considered representative at a given depth where at least 30% (Dewandel et al., 2005) of the data
322 are available, as not all wells reach the deepest depths. Consequently, data below 150-175 m below
323 the saprolite bottom are not considered significant. The productive fractures’ density (Figure 7b) was
324 corrected according to this observation density (for each depth interval, corrected density equals raw
325 density multiplied by the percentage of observation).

326 The average of the total air-lift flowrate by borehole is 30 m³/h, with 30% of the boreholes reaching
327 higher flowrates, which is rather high for crystalline rock aquifers. Cumulated average flowrate
328 increases with depth below the saprolite in the first 100-125 metres below the base of the saprolite,
329 then tends to stabilize down to 150-175 m and becomes erratic deeper because of the lack of data.
330 Similarly and logically, the productive fracture density is the highest in this 100-125 m interval; this
331 was also observed on raw data, i.e. not corrected from observation density (not shown on Figure 7).
332 The thickness of the fissured layer of the granite aquifer is then about 100 to 125 m (see also Figure 3).

333 **c) Hydrodynamic properties of the aquifers (from pumping tests)**

334 This 100-125 m thick granite aquifer shows high transmissivity values for a fractured granite aquifer
335 ($\sim 10^{-3}$ m²/s, Figure 8). This is about 10 to 100 times the values usually expected for such a lithological
336 context. The transmissivity of the granite aquifer is lower (down to 10^{-5} m²/s) in the deepest part of
337 the hydrosystem located in the western part of the study area under a thick sedimentary cover
338 (Figure 8); this result will be discussed later. Storage coefficients range between 10^{-3} and 5×10^{-5} .
339 The granite aquifers also exhibit minor leakage effects (E symbol on Figure 6) that are attributed to
340 the low hydraulic conductivity Cenozoic sediments, and/or to the saprolite layer, and particularly its
341 laminated bottom. The estimated hydraulic conductivity of leaky layers is about 10^{-9} - 10^{-8} m/s.
342 The transmissivity of Cenozoic deposits aquifers ranges from 2×10^{-5} to 8×10^{-3} m²/s. Much more than
343 for granite aquifers, their hydrodynamic parameters are highly heterogeneous and they are
344 compartmented, due to the small lateral and vertical extension of sand bars. Despite its estimation
345 was limited by the absence of piezometers, the spatial extension of these compartments, of about
346 0.1 km² each, appears to be lower than the one of the granite aquifers. Leakage effects (10^{-10} - 10^{-7}
347 m/s) were found for a few pumping tests, highlighting the multi-layered structure of the Cenozoic
348 deposits.

349 **3.3. Piezometric data**

350 The piezometric data (Figure 9) logically show that, in each exploited aquifer compartment, the
351 groundwater flows converge towards the abstraction borehole (logically, only one well is pumped in
352 each aquifer compartment). This scheme is particularly clear in the A3 compartment where a
353 detailed piezometric data set is available. The high piezometric differences between the different
354 aquifer compartments confirm the low permeability of the boundaries that limit them that were
355 identified and located from both pumping tests and the geological model.

356 A “Multiple Input Single Output” temporal analysis of the piezometric time series (e.g. Maréchal et
357 al., 2014) from the granite aquifer (fissured layer) was used to discriminate the impact of pumping,
358 rainfall and water level from the Coise stream and sedimentary formations. Impulsional responses
359 associated to each input highlight that piezometric variations in the granite aquifer are i) weakly

360 related to the climatic variability and water levels in sedimentary formations, and not at all to the
361 Coise stream, ii) quite exclusively influenced by the pumping rate in each compartment and iii) not
362 influenced by the pumping rate changes in adjacent compartments, which confirms the no-flow
363 boundaries between them.

364 **3.4. Geochemical tools**

365 *3.4.1 GW chemical facies*

366 The spatial distribution of GW mineralization is highly variable in space, at both local (Saint-Galmier)
367 and regional (Forez plain) scale, ranging from few hundreds to more than 6000 $\mu\text{S}/\text{cm}$. Graphical
368 major ions analysis using Piper diagrams and biplots were first used to check if the lithology of the
369 GW reservoir (Cenozoic deposits/unconsolidated saprolite/fractured granite) could explain the GW
370 chemical facies. No relationship was found, showing that local water-rocks interaction is not the main
371 process that explains the GW chemistry. Mixing processes can thus be proposed, which implies to
372 identify different poles to describe the mineralization, regardless of the nature of the geological
373 reservoir.

374 Highly mineralized water ($>4000 \mu\text{S}/\text{cm}$) is mainly of $\text{HCO}_3\text{-Na}$ type in the Forez plain but mineralized
375 water of $\text{HCO}_3\text{-Ca-Mg}$ type can also be found in the Saint-Galmier area. This is why the Ca/Na ratio
376 was used to identify the relative contribution of these two mineralized poles. A third type of GW is
377 defined by its low mineralization ($<500 \mu\text{S}/\text{cm}$), which is representative of aquifers that are not fed
378 by deep mineral inputs. The 2 mineralized poles are clearly highlighted on Figure 10 where the Ca/Na
379 ratio is compared to the electrical conductivity: Ca-Mg -type for which the Ca/Na ratio is higher than
380 0.6 and Na -type for which the Ca/Na ratio is below 0.2. Low-mineralized water ($<500 \mu\text{S}/\text{cm}$) appears
381 as a dilution pole and shows a large range of Ca/Na ratio. Mixing processes occurring between the
382 different groundwater poles can thus explain the variability of chemical facies and electrical
383 conductivities of groundwater within the system (Figure 10).

384 *3.4.2. Spatial analysis*

385 As the samples are well distributed across the area, the sodium-pole indicator helps in characterizing
386 the spatial representation of the sodium pole within each aquifer compartment through kriging
387 interpolation and also to compare the chemistry of the various compartments (Figure 13). In this
388 representation, the groundwater samples from the “dilution pole” were excluded to focus on the
389 location of the input of deep mineralized water within the aquifer system. Then, the spatial
390 distribution of the chemical facies is consistent with the aquifer compartments previously identified.
391 Figure 13 shows that the Na -type water is associated with the mineralized province observed around
392 Violes where EC reaches $5000 \mu\text{S}/\text{cm}$, whereas the mineralized area of Pétillante, Généreuse and the
393 surrounding boreholes also shows highly mineralized water ($>5000 \mu\text{S}/\text{cm}$) associated to a relatively
394 strong calcium facies. The area of the Ca-Mg water type (Ca/Na ratio > 0.6) therefore appears to be
395 moderately mineralized compared to Na -type groundwater observed elsewhere. This Ca-Mg water
396 type also remains spatially limited, in the Saint-Galmier area.

397 Beyond the spatial repartition of the different types of mineralized water within specific
398 compartments based on the former analysis, the distribution of electrical conductivities highlights
399 the distribution of the groundwater from the dilution pole within compartments. It is clearly visible
400 that the least mineralized groundwater (e.g. Aubignat, Cote Rouge and Hippodrome) is found in
401 specific aquifers compartments (Figure 13). Mineralization gradients in aquifer compartments are
402 induced by pumping boreholes location as well as the location of the sources of deep mineral water
403 and dilution poles. To precisely identify these locations, deterministic modelling was performed on
404 the best known compartment (A3; cf. §3.5).

405 *3.4.3. Temporal analysis.*

406 Chemical data from the A3 compartment were analysed with the EMMA statistical method. Firstly, a
407 PCA (Principal Component Analysis) was performed using the 6 major ions HCO_3^- , Ca^{2+} , Mg^{2+} , Na^+ , Cl^-
408 and NO_3^- . 84% of the total variance was explained by 2 factorial axes. The F1 axis, linked to overall
409 mineralization, highlights the mixing processes between deep mineralized groundwater and the
410 dilution pole in the A3 borehole (Figure 10) and within the A3 compartment. The best candidate for
411 the “deep” pole was the composition of the A3 borehole groundwater in 1985 (i.e. just after its

412 drilling, at the beginning of its abstraction). There was no clear chemical distinction between dilution
413 pole from the saprolite and from the Cenozoic sedimentary aquifer, but a mixing of two distinct
414 dilution end-members is needed. The best fit was given using two water samples from boreholes
415 drilled in the unconsolidated saprolite.

416 Contribution rates from the deep pole and the dilution pole were simulated over the time using the
417 time series of chemistry of A3 groundwater and the corresponding pumping rates (Figure 11). The
418 EMMA allows reproducing the hydrochemical time series of the 6 major ions with a correlation
419 coefficient between 0.86 and 0.95. The deep contribution is relatively stable. It highlights that
420 pumped water in A3 is composed of about 60% of the deep component and 40% of the dilution pole.
421 This result is used to better constrain the deterministic modelling of the A3 compartment (cf. § 3.5.3).

422 **3.5. Deterministic modelling of the A3 compartment**

423 The calibration of the numerical model was carried out by simulating the system both under natural
424 conditions before pumping started and under abstraction in both granite (abstraction of 18 m³/h in
425 A3) and Cenozoic aquifers. The main criteria to assess the validity of the model were i) the location of
426 the the historical Fontfort spring that was flowing out from this compartment (Figure 4) in the
427 natural state and its absence after several centuries of borehole abstraction, ii) the consistency of the
428 calibrated parameters and boundary conditions compared to values of reference data (i.e.
429 permeability, recharge, deep mineralized water flow) and iii) the calculated groundwater levels
430 compared with available measurements.

431 *3.5.1. Boundary conditions and hydraulic properties*

432 To reproduce the East-West regional flow in the Cenozoic formations, a constant hydraulic head
433 condition equal to the topographic level minus 1.5 metres was imposed on the cells of the western
434 limit of the model. For the other lateral boundaries of the model, it is assumed that the modelled
435 area is sufficiently large compared to that of the A3 reservoir for boundary conditions not to affect
436 the flows in the A3 area. A no-flow condition is thus imposed along the perimeter of the modelled
437 area for all layers. In the fissured granite layer, the aquifer compartment exploited with the A3
438 borehole is laterally limited by impervious boundaries, shaping the reservoir limits drawn on Figure 9
439 (Figure 12). The bottom of the model is impervious with a no-flow condition. However to allow the
440 rise of deep mineralized water, a constant vertical flow of 8 m³/h uniformly distributed over twenty
441 cells is imposed. This upward flow is located in the area where a sub-vertical deep fault is expected
442 from the hydrochemical spatial analysis (cf. § 3.4.1) and from the geological model. A uniform and
443 constant recharge is imposed on the outcropping cells of the model, i.e. those of the Cenozoic
444 formations located on the west of the fault separating Cenozoic formations from the granite, and
445 those of the saprolite on the eastern part.

446 The permeability of the different layers of the model was set in agreement with the results of
447 pumping test interpretations which notably highlighted that transmissivity values found for granite
448 aquifer in the area are relatively high (i.e. about 10⁻³ m²/s; hydraulic conductivity: 1.5x10⁻⁵ m/s)
449 (Figure 8). The sub-vertical deep fault feeding the aquifer with deep mineralized fluid, which was not
450 intercepted by any borehole, is defined as a more permeable zone than the surrounding layers with a
451 vertical anisotropy ($K_z = 10 \cdot K_{xy}$).

452 *3.5.2. Calibration and results*

453 The hydrodynamic model calibration consisted in multiple steady state simulations with various
454 parameters in order to fit at best piezometric levels and feature the observation data.

455 The recharge rate of the saprolite is the main calibrated parameter of the model. The applied
456 recharge is 142.5 mm/year over the saprolite. A recharge rate of about 10% of this value is imposed
457 (i.e. 15 mm/year) on the Cenozoic sedimentary deposits, which present a lower infiltration rate
458 because of their low hydraulic conductivity. Other parameters as hydraulic conductivities of layers
459 and location of boundaries are close to the ones defined by the analysis of pumping tests (i.e.
460 hydraulic conductivity of post-rift sediments and saprolite 10⁻⁷m/s and 5x10⁻⁹m/s respectively).

461 The extension and orientation of the sub-vertical fault in the granite and the value of the upflow of
462 mineralized fluid were also adjusted during the calibration phase in order to match with the
463 observed chemical contents at the various boreholes located within the A3 compartment.

464 In agreement with the main features of regional hydrogeology, the calibrated model in natural
 465 regime reveals generally east-west oriented flows, drained by the Coise stream. At the east of the
 466 boundary fault, groundwater from granitic layers overflows along the Coise stream and from two
 467 cells located on the right bank of the stream close to the historical Fontfort spring. In the central part
 468 of the A3 compartment along the Coise, the piezometric levels are less than 10 metres deep from the
 469 ground level. These elements are fully consistent with the description of the system at the beginning
 470 of the industrial abstraction as proposed by Archambault (1947). The sum of the overflow of the A3
 471 reservoir on the right bank of the Coise is 15.3 m³/h, i.e. 4.25 L/s, which seems to be a reasonable
 472 order of magnitude for this type of spring.

473 Under abstraction condition, the hydraulic heads are rather well reproduced (Table 1). The simulated
 474 hydrogeological system presents no overflow in the saprolite, which is consistent with current field
 475 observations. Outputs from the Cenozoic deposits to the Coise stream are less important (1.6 m³/h)
 476 than in the natural state (4.6 m³/h). Exchanges between the Cenozoic sediments and the
 477 unconsolidated saprolite are locally upwards, locally downwards, with a consistent spatial
 478 distribution, and are balanced on the entire modelled area. Considering only the A3 reservoir area,
 479 the Cenozoic aquifer feeds the unconsolidated saprolite (3.4 m³/h), which had been previously
 480 highlighted by the interpretation of pumping tests (i.e. leakage effect).

481 Table 1: Comparison of modelled and measured hydraulic heads (metres above mean sea level) (steady state
 482 and under abstraction conditions) for A3 and observation boreholes in the A3 compartment. See location of
 483 the boreholes on Figure 9

	Modelled (m)	Measured (m)	Delta (m)
A3	307.88	307.2	+ 0.68
SG13	314.34	314.53	- 0.19
SG3bis	312.09	311.25	+ 0.84
Badoit IV	312.94	312.11	+ 0.83
Flute 70	311.37	311.64	- 0.27
A1	311.7	311.05	0.65
A2	310.43	310.48	- 0.05

484

485 3.5.3. Coupling with hydrochemical tool

486 After hydrodynamic calibration, the model was applied to validate and enhance with spatialized
 487 results the mixing model derived from the statistical analysis on the A3 groundwater. Indeed, the
 488 EMMA analysis showed that the water mineralization in the A3 compartment and the dynamic
 489 evolution of 6 major ions can be explained by a combination between two poles, one deep and
 490 mineralized, and a freshwater dilution pole. Steady state non reactive transport simulations were
 491 performed with the model in order to set boundary conditions regarding the location and
 492 composition of the inflows of water representative of these two poles. The final configuration allows
 493 simulating concentrations of major ions to be as close as possible to measured data in the boreholes
 494 (Figure 12). The deep pole is entering the system through the deep sub-vertical fault crossing the A3
 495 compartment and the dilution pole is associated to the recharge over the outcropping saprolite to
 496 the east of the border fault. Figure 12 presents the simulated spatialized concentration of the Ca²⁺
 497 ion which is well representative of the mixing rate between deep mineralized Ca²⁺ rich water and the
 498 dilution pole which is poorer in Ca²⁺.

499 Table 2: Comparison of measured and modelled electrical conductivity and concentration of the main major
 500 ions (mg/L) in A3, SG3bis and SG13 boreholes from the A3 compartments (see Figure 9 for location of
 501 boreholes)

	μS/cm	mg/l							
	EC at 25°C	SiO ₂	Ca ²⁺	Mg ²⁺	Na ⁺	K ⁺	HCO ₃ ⁻	Cl ⁻	SO ₄ ²⁻
A3 measured	1584	26	173	69	126	11	1115	71	54
A3 modelled	1451	23	162	61	104	10	938	66	53

<i>Delta</i>	<i>8.4%</i>	<i>9.8%</i>	<i>6.6%</i>	<i>12.1%</i>	<i>17.5%</i>	<i>9.2%</i>	<i>15.9%</i>	<i>7.6%</i>	<i>1.9%</i>
SG3Bis measured	1328	22	147	50	81	9	724	70	52
SG3bis modelled	1471	25	161	60	104	9	919	76	53
<i>Delta</i>	<i>-10.8%</i>	<i>-12.7%</i>	<i>-9.2%</i>	<i>-20.1%</i>	<i>-28.6%</i>	<i>-4.9%</i>	<i>-26.9%</i>	<i>-8.5%</i>	<i>-2.5%</i>
SG13 measured	508	16	53	20	36	4	223	43	32
SG13 modeled	426	13	30	6	40	10	105	36	22
<i>Delta</i>	<i>16.1%</i>	<i>18.1%</i>	<i>43.7%</i>	<i>71.7%</i>	<i>-12.3%</i>	<i>-161%</i>	<i>52.8%</i>	<i>16.4%</i>	<i>31.8%</i>

502
503 According to the proposed configuration of inflows into the system, the model water budget shows
504 that 44% of recharge of the A3 compartment comes from the deep mineralized pole, 45% from the
505 saprolite and 11% from leakage of Cenozoic formations. Sensitivity tests were carried out during the
506 model construction and after the calibration phase to explore its capabilities. One will retain that:
507 (i) the position and shape of the deep fault into the A3 compartment naturally affect results,
508 especially on calculations of chemical concentration. An extension of the fault to the North is
509 required to reproduce the high mineralization in SG3bis and the extension to the South-East to
510 explain the lesser mineralization of SG13 (Figure 12); (ii) overall, a recharge on the Oligocene
511 between 0 and 50 mm per year provides similar results.

512 The numerical model shows the consistency of all hydrodynamic and geochemical interpretations
513 presented above and thus confirms the assumptions about the hydrogeochemical functioning of the
514 A3 compartments. Also, it allowed the rather precise location of the deep fracture providing Ca-Mg-
515 type deep mineralized water.

516 **3.6. Hydrogeological conceptual model**

517 The very good consistency between all the clues obtained from the various approaches implemented
518 in the frame of this research (geological model, hydrodynamics, hydrogeochemistry, modelling, etc.)
519 allows formulation of a realistic and robust hydrogeological conceptual model (Figs 13 and 14). Such
520 a conceptual model is rather different from the ones already published for such CTMCG
521 hydrosystems.

522 The Saint-Galmier's CTMCG aquifer is hosted within the stratiform fissured layer of the granite with a
523 thickness of about 100-125 m, which lies below the low hydraulic conductivity of both Forez graben
524 sediments and granite's saprolite. This stratiform fissured layer belongs to a weathering profile that
525 developed during several phases of weathering. Indeed, the thin layer of unconsolidated saprolite
526 and the unusual thickness of the laminated saprolite layer can be interpreted as resulting from a
527 poly-phased weathering process (Dewandel et al., 2006). The abnormally thick laminated horizon
528 surely developed on an ancient fissured layer inherited from a first weathering phase, re-weathered
529 during a second weathering phase. There is no clear indication at Saint-Galmier about the age of
530 these two phases, however, as known regionally (Lachassagne et al., 2015), the first one probably
531 occurred during the Early Cretaceous and the second one during Early to Middle Eocene period, prior
532 to Late Eocene to Oligocene graben filling. On the raised compartment of the graben (east part of the
533 N-S regional fault), where the crystalline rocks now outcrops, the weathering processes are more or
534 less continuous since the regional uprising during Late Miocene and erosion of Tertiary sedimentary
535 deposits. Contrarily, weathering process on the granite below the graben's sediments surely stopped
536 since Eocene.

537 The density of hydraulically conductive fractures of the stratiform fissured layer decreases with
538 depth within the weathering profile, as classically observed in such profiles (see Dewandel et al.,
539 2006 for instance). The fracture hydraulic conductivity of this aquifer appears to be mostly related to
540 weathering processes; however tectonic structures or other discontinuities such as veins, joints, etc.
541 (see Dewandel et al., 2011 or Lachassagne et al., 2011; Roques et al., 2016) may also contribute to
542 the hydraulic conductivity of the aquifer. Compared to other data on hydrodynamical properties of
543 such fissured layer in granite (Maréchal et al., 2004; Dewandel et al., 2006; 2011), the St-Galmier
544 granite is characterized by high transmissivity values (between $8 \cdot 10^{-5}$ and $3 \cdot 10^{-3}$ m²/s in most

545 boreholes; Figure 8). These high values are believed to result from weathering processes, with a
546 weathering-induced fissures density enhanced by the fracturing of the area. However, transmissivity
547 of the fissured layer strongly decreases where it is buried below a high thickness of Oligocene
548 sediments (down to 10^{-5} m²/s). The origin of this decrease is yet to be investigated and understood:
549 closing of the granite's subhorizontal fractures caused by the sedimentary deposits weight?
550 Diagenetic clogging of the fractures?

551 As shown by boreholes and geophysics, this aquifer extends over a surface area of several square
552 kilometres, which explains its overall long term productivity. It is however divided into several (at
553 least 10), kilometre in length, hectometres in width, compartments. These compartments are all
554 bounded by impervious boundaries. Most of these impervious boundaries -for instance for the Côte
555 Rouge, Violes-Primevères, Aubignat, SB/SB', and quite surely for the Joyeuse-Frarie, Besacieux,
556 Richelande and A3 compartments- are associated with a great confidence to identified faults. These
557 faults all act as impervious boundaries either due to their impervious clayey fault gouge and/or as
558 they offset the aquifer compartments one to the others and put in contact the aquifer with
559 impervious layers such as, from bottom to top, the unfractured granite, the saprolite and the
560 Oligocene sediments.

561 This aquifer is bounded at its bottom by the unfractured impervious granite, and at its top by the
562 weathering profile's saprolite and, where present, by the Oligocene sedimentary cover, mostly also
563 of very low hydraulic conductivity. The geological synthesis and modelling however allowed
564 identifying two main lithological facies within the sediments: the clayey syn-rift sediments of very
565 low hydraulic conductivity and the clayey-sandy post-rift sediments that locally exhibit higher
566 hydraulic conductivities. The absence of the syn-rift sediments between the granite aquifer and the
567 overlying post-rift sediments, but also locally the absence, due to erosion, of the saprolite, locally
568 allows flow and/or leakage, and chiefly explains that in such specific areas the post-rift sediments are
569 invaded by natural mineral water.

570 The aquifer is fed on the one hand by deep carbo-gaseous highly mineralized mineral water rising up
571 only by a few permeable subvertical fractures or parts of fractures, and on the other hand by water
572 from local aquifers recharged by rainfall. The latest are mostly granite aquifers slowly recharged as a
573 consequence of the low hydraulic conductivity of the saprolite cover and Oligocene sediments. In the
574 southern part of the CTMCG hydrosystem, it is a NW-SE subvertical fracture that feeds the A3,
575 Pétilante and Généreuse compartments with carbogaseous Ca-Mg rich deep water. Nevertheless,
576 the hydrogeological modelling of the A3 compartment showed that a short length of a N-S fracture,
577 neighboring the NW-SE fracture, is also required. In the northern part of the CTMCG hydrosystem,
578 only a segment of one N-S fracture is enough to explain the carbogaseous Na-rich pattern of water
579 from the Violes-Primevères compartment. All the other compartments are interpreted to be either
580 fed by waters arriving from other compartments (for instance "Joyeuse-Frarie" where a mixing
581 between Ca-Mg and Na-rich waters occurs) or are not fed by deep mineral water (e.g. Côte Rouge
582 and Aubignat) but only by the local recharge. The particularity of this case study is that the
583 compartments are undergoing constant pumping. This provides an advantage for the interpretation
584 because a flux toward the abstraction boreholes is created, and the pumped water is there
585 representative of a larger area, which can overcome the low density of observations given the
586 complexity of such a system (in average hardly a bit more of one borewell/piezometer per identified
587 compartment). The study of geochemistry and piezometry allows validating the
588 compartmentalization of the aquifer, as the chemical facies is different for each box and samples
589 integrate the waters of each compartment by convergent fluxes. This hypothesis of flux converging
590 towards abstraction boreholes within each compartment is largely supported by the few available
591 piezometric data and the hydrochemistry.

592 Before the exploitation of the aquifer by pumping (till the XIXth Century), the natural mineral water
593 was naturally flowing out at the Fontfort spring which is the lowest place in the area, in the Coise
594 stream valley, where the granite aquifer outcrops near the contact with the low permeability
595 Oligocene sediments. The natural mineral water was also leaking upwards through the saprolite
596 and/or the sedimentary cover, particularly in areas where the saprolite and/or the syn-rift low

597 permeability sediments were eroded. The exploitation of the aquifer by pumping modified the flux
598 directions, and also mixings, in several compartments of the aquifer, particularly in the A3 and
599 Joyeuse-Frarie compartments, and explains the current distribution of the natural mineral water
600 hydrochemical facies in the aquifer.

601 4. Discussion

602 A realistic and robust hydrogeological conceptual model of a CTMCG hydrosystem is proposed based
603 on a detailed characterization of the system: geology, hydrogeochemistry, hydrodynamic, modelling.
604 In fact, to our opinion, only a multidisciplinary approach such as the one performed here (see also for
605 instance Lachassagne et al., 2009; Maréchal et al., 2014) allows such an in-depth characterization for
606 achieving a realistic conceptual model. It is the convergence of the clues obtained from the various
607 methods, and not only clues of one method, that ensures the robustness of each part of the
608 conceptual model and, overall, of the conceptual model globally.

609 This conceptual model is of course of great use for the day-to-day operational management of this
610 NMW resource and its protection. It opens up further operational perspectives such as the numerical
611 modelling of all the aquifer's compartments. However, the results of this research also open-up
612 larger fundamental perspectives.

613 In such CTMCG hydrosystems, “the aquifer” at the origin of the carbogaseous natural mineral water
614 is rarely identified, and most authors vaguely consider that, as the hydrosystem is fractured, the
615 tectonic fractures that provide the deep groundwater also constitute “the main or the only possible
616 aquifer” (e.g. Stober and Bucher 1999). For St-Galmier CTMCG, we clearly demonstrate that:

617 (i) the fissured stratiform layer of the crystalline rock's (here a granite) weathering profile constitutes
618 “the aquifer”. This large and thick aquifer explains the high hydrogeological inertia of the studied
619 boreholes and can be numerically modelled similarly to a porous aquifer (see also Lachassagne et al.,
620 2001);

621 (ii) as a consequence, this aquifer has a large extension (several square kilometres) and is “easily”
622 reached by any vertical borehole, which explains the quite high quantitative success of the drillings in
623 this area, as compared for instance to other CTMCG aquifers where this weathering profile was
624 totally eroded (see for instance the case study of Vals-les-Bains, France; Ledoux et al., 2016);

625 (iii) this aquifer is compartmented by faults that all act as impervious boundaries. The low hydraulic
626 conductivity of most faults was already largely demonstrated (see the references cited in the
627 Introduction). However, it does not prevent to continue to advertise about this fact as several
628 hydrogeologists “naturally” tend to consider the contrary;

629 (iv) this aquifer is only very locally fed by deep highly mineralized carbo-gaseous water originating
630 from a few segments of subvertical tectonic fractures that only constitute a few percents of the total
631 length of identified faults.

632 This duality (fractures locally pervious but mostly impervious) has been previously mentioned (e.g.
633 Bense et al. 2013). More recently, in the same structural context (i.e. peri-Alpine graben), Gumm et
634 al. (2016) highlighted the rise of deep fluid along faults that also act as barrier for the shallow
635 compartments. Of course, some tectonic fractures may locally “boost” the hydraulic conductivity of
636 the aquifer. It will be one of the future challenges of research on this site to try to identify if such
637 areas exist, to find a mean to distinguish both types of fracturing (weathering and tectonic
638 respectively) and their associated hydrodynamic properties.

639 As a research perspective and with the objective to generalize the concepts developed in this paper,
640 particularly the one from § (ii) above, it would also be of high interest to re-examine, on the one
641 hand the data from CTMCG located in a similar context where the weathering profile may have been
642 preserved from erosion (for instance the CTMCG bordering grabens filled with sediments, see for
643 instance Figure 1), and, on the other hand, CTMCG where the weathering profile was eroded, to
644 deepen the interpretation presented in this paper. Moreover, this re-analysis may help to better
645 understand their geological structure and hydrogeological functioning and further to help their
646 management and protection.

647 Moreover, the approach presented in this paper also allowed the precise location and the
648 characterization (orientation, hydrochemistry of the deep water produced by the fracture) of some,
649 or quite surely all, of the subvertical tectonic fractures feeding the aquifer with deep highly
650 mineralized carbo-gaseous water. In other studies, such a trial to characterize these fractures is often
651 only resulting from a geological structural analysis or, worse, only by a “lineament study”, without
652 any strong and robust demonstration. We here also consider that such a multidisciplinary approach
653 comprising, among others (of course a strong structural and geological model), a hydrodynamic and
654 hydrochemical approach (synthesized, in the A3 well compartment, by the numerical modelling) is
655 much more robust to identify and characterize such feeding fractures. In a previous similar study
656 (Maréchal et al., 2014), we showed that such fractures weren’t detected at all by the only classical
657 geological and structural approach.

658 Analysis of the spatial distribution of mineralization of groundwater shows that the deep fractures
659 play a major role on the facies of mineral waters. The gas enriched waters rise up under hydraulic
660 and thermal gradient effect thanks to some segments of the sub-vertical faults. They drain deep
661 fluids and mantellic CO₂ (Blavoux, 1993). However, mineral waters have variable chemical facies and
662 appear either Ca-Mg-type (typical Saint-Galmier facies) or Na-type. To date, the most reasonable
663 explanation for the variability of the chemical facies is that uprising fluids are enriched in various
664 cations according to the nature of the granite and the encountered secondary minerals within the
665 faults throughout the ascent of the fluid.

666 5. Conclusion

667 This multidisciplinary study (geology, hydrogeology, geochemistry and modelling) has clarified the
668 hydrogeological functioning of the Saint-Galmier carbo-gaseous system thanks to the confrontation
669 using results from complementary methods.

670 The geological approach was based upon numerous available data (logs drilling, seismic, gravimetric
671 and field surveys) and allowed the realization of a 3D geological and structural model. Then, on the
672 basis of this geological model and, hydrodynamic and geochemical data, a robust and realistic
673 conceptual hydrogeological model was built. It highlights the high compartmentalization of the
674 aquifer system and the mixing between deep mineralized water ascending through deep faults and
675 more superficial dilution pole within each compartment.

676 The permeability of the crystalline aquifer is found to be relatively high, especially when the
677 sedimentary cover is thin. The fissured layer of the weathering profile plays a major hydrodynamic
678 role compared to tectonic faults as it hosts thick and extended aquifers, delimited by low hydraulic
679 conductivity faults.

680 A deterministic hydrogeological model of the A3 compartment has been implemented with the aim
681 to validate the functioning hypotheses that emerged from these complementary approaches. The
682 hypotheses were tested successfully and the numerical model proposes a very sound solution for the
683 orientation and the location of the fault segments providing deep mineral fluids. The comparison of
684 the information from the transport simulations with data from geochemical analysis (EMMA) makes
685 possible the estimation of the signature of the Ca-Mg-type mineralized pole of this major reservoir.

686 The chemical variability of the mineralized water in such a restricted area is, to date, explained by the
687 probable variability of the hydrothermally dissolved minerals, but we are not able to clearly state on
688 the origin of the very localized Saint-Galmier facies (i.e. Ca-Mg type carbo-gaseous mineral water).
689 The original composition of the deep mineralized poles and the associated mineralization processes,
690 the transit time and mixing rates within the compartments still remains largely unknown. With this
691 goal in mind, additional sampling were run in 2015 for a wide range of tracers analyses (e.g. major
692 ions, stable isotopes of the water molecule, $\delta^{13}\text{C}$, $^{87}\text{Sr}/^{86}\text{Sr}$, $^{234}\text{U}/^{238}\text{U}$, ^{226}Ra , ^{222}Rn , noble gases,
693 $^3\text{He}/^4\text{He}$, SF₆, ^3H , ^{39}Ar and ^{85}Kr) and should provide additional information.

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697 comments, remarks and proposals that helped improving the manuscript.

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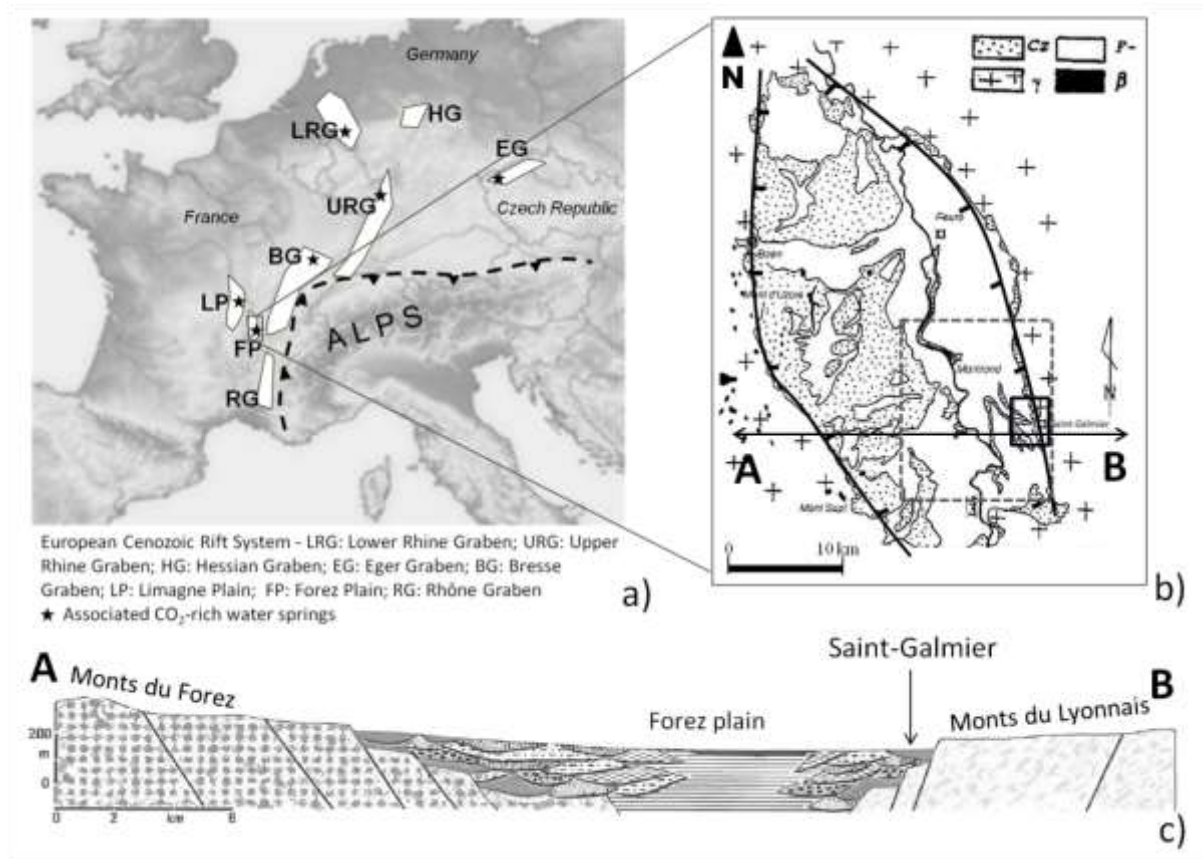
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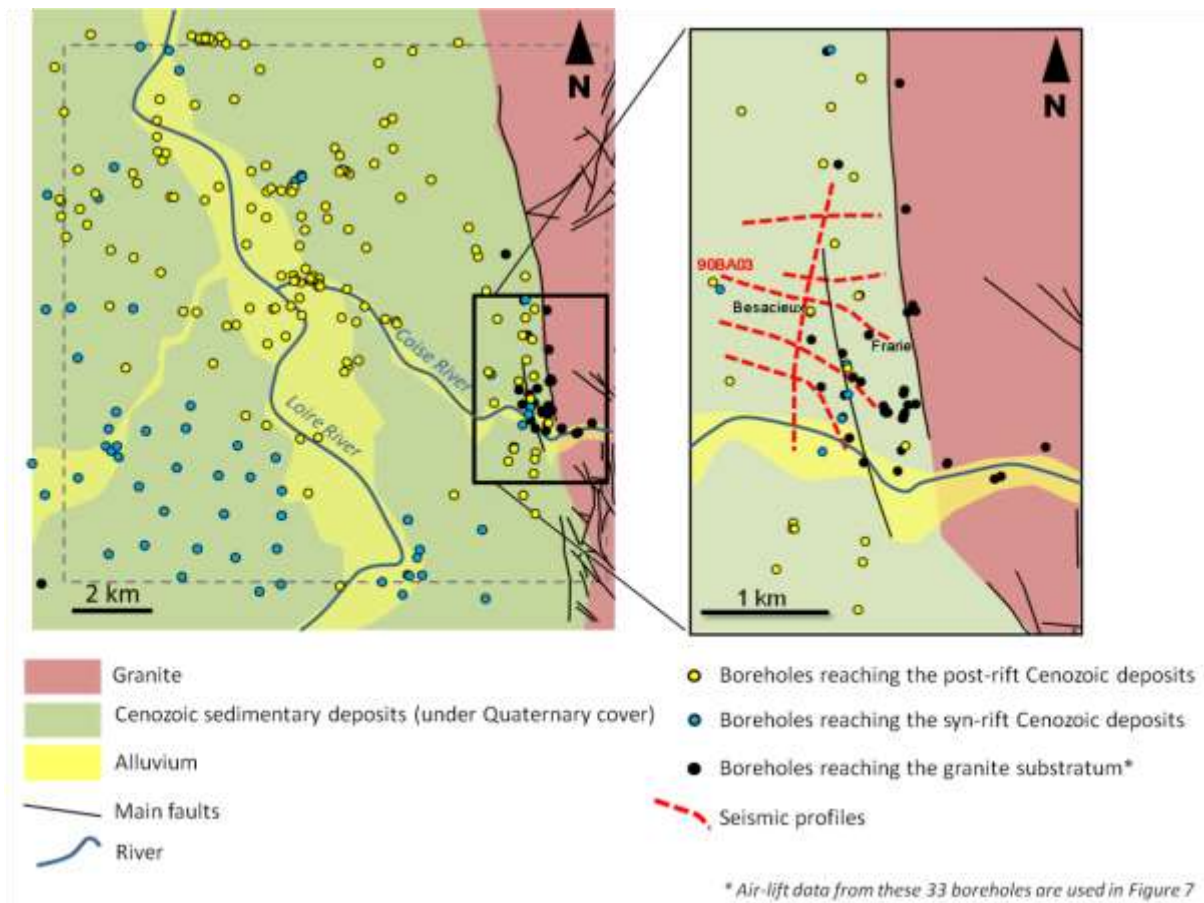
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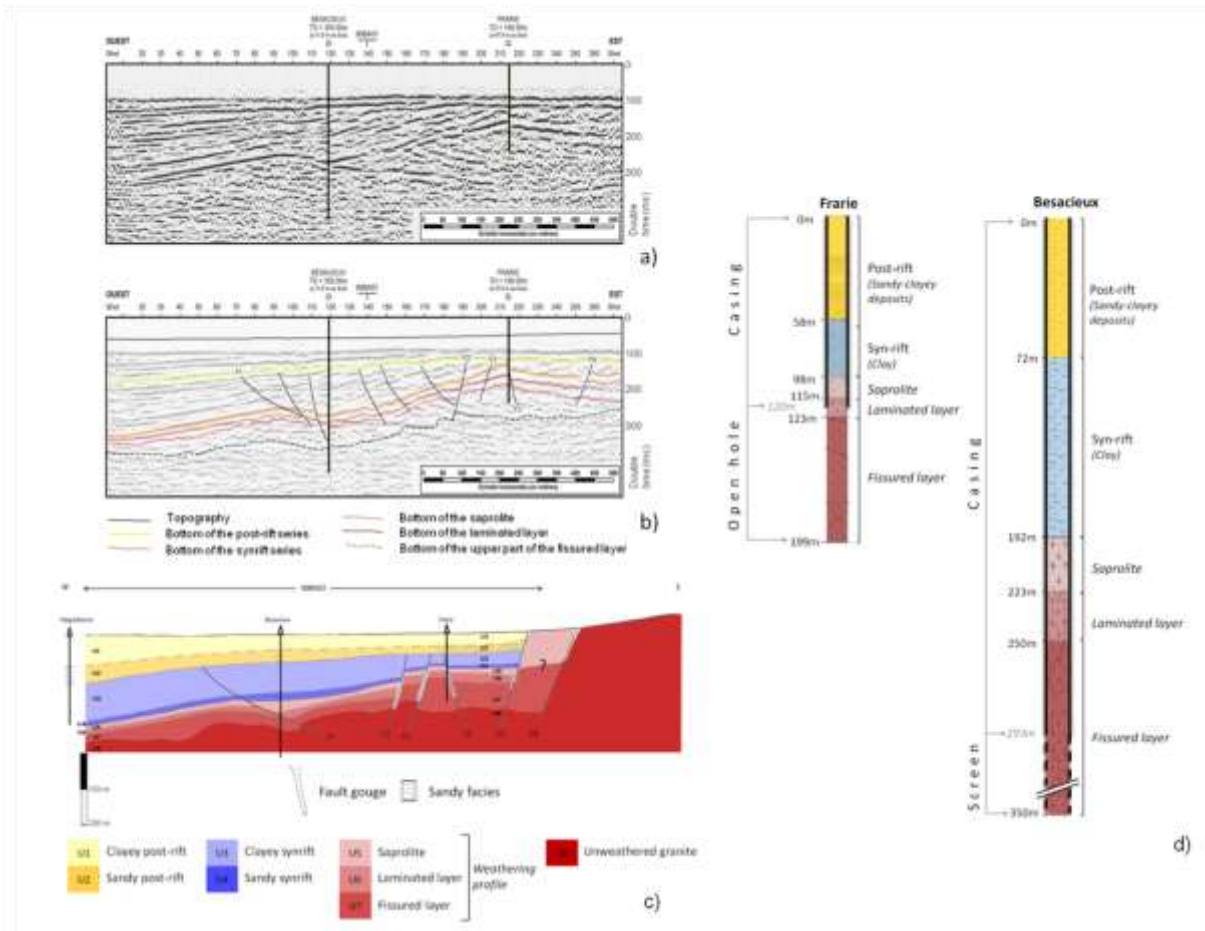
858 **Figures**



859

860 Figure 1: a) Location of the main peri-alpine grabens of the European Cenozoic Graben System; b)
 861 Main geological outcrops of the Forez Graben delimited by the main regional normal faults, cz:
 862 Cenozoic sedimentary deposits; F: Quaternary alluvium and colluvium; γ: Paleozoic bedrock; β
 863 Miocene volcanic formations (blue frame: geological model extended area and red frame: narrow
 864 geological model area) (Gerbe et al., 1998 modified); c) Schematic profile of the Forez Graben along
 865 the AB section. From West to East: the Monts du Forez massif (granite), the Forez plain
 866 (heterogeneous continental deposits) and the Monts du Lyonnais massif (metamorphic bedrock
 867 intruded by the granitic pluton of Saint-Galmier) (adapted from Le Griel, 1984).

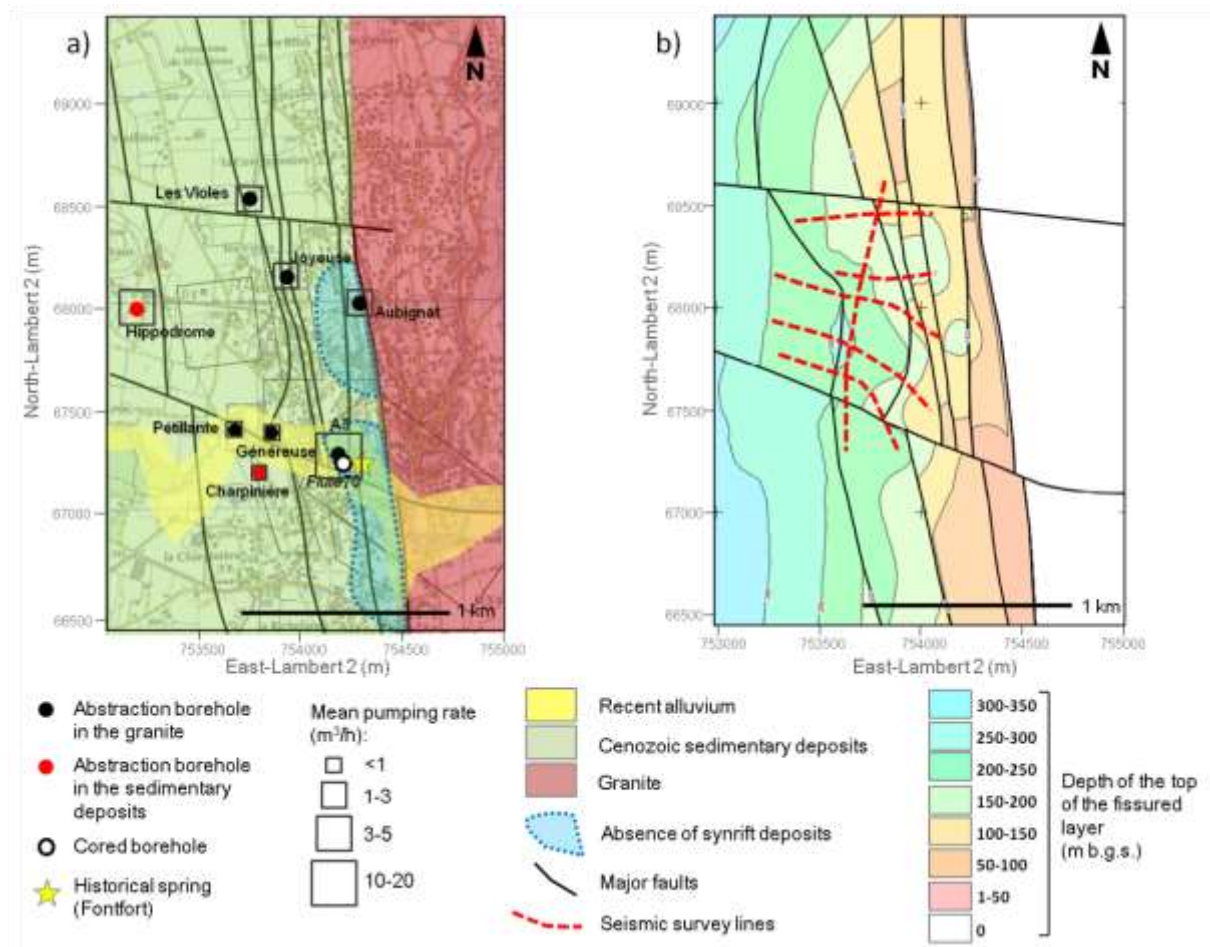




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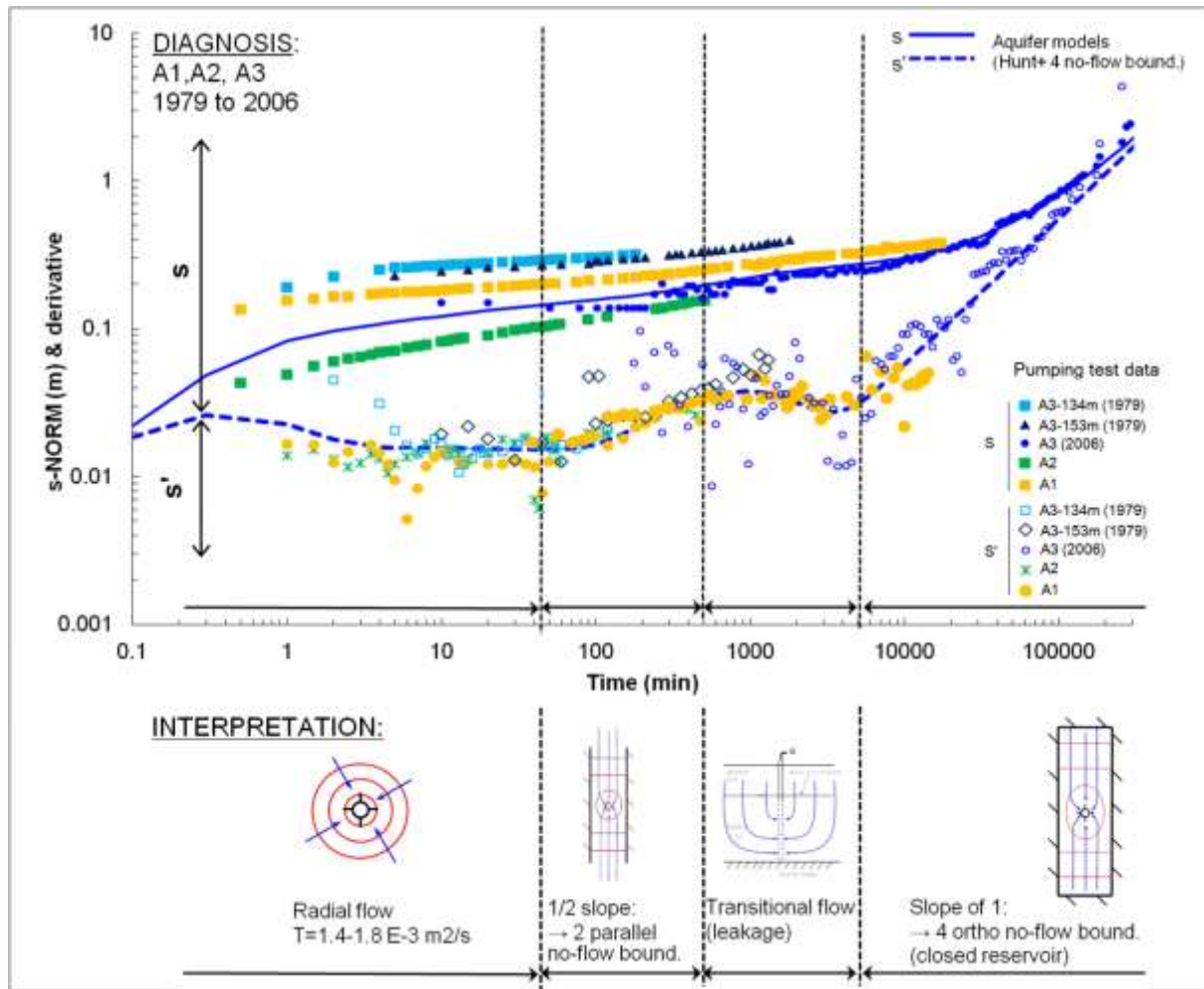
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Figure 3: a) Digitized raw data of the 90BA03 seismic profile (see location on Figure 2); b) Line drawing (note the sealing of syn-rift normal faults by post-rift deposits); c) Geological cross section obtained by conversion from seismic times to depths and complementary borehole data d) Simplified technical and geological logs of Besacieux and Frarie boreholes.



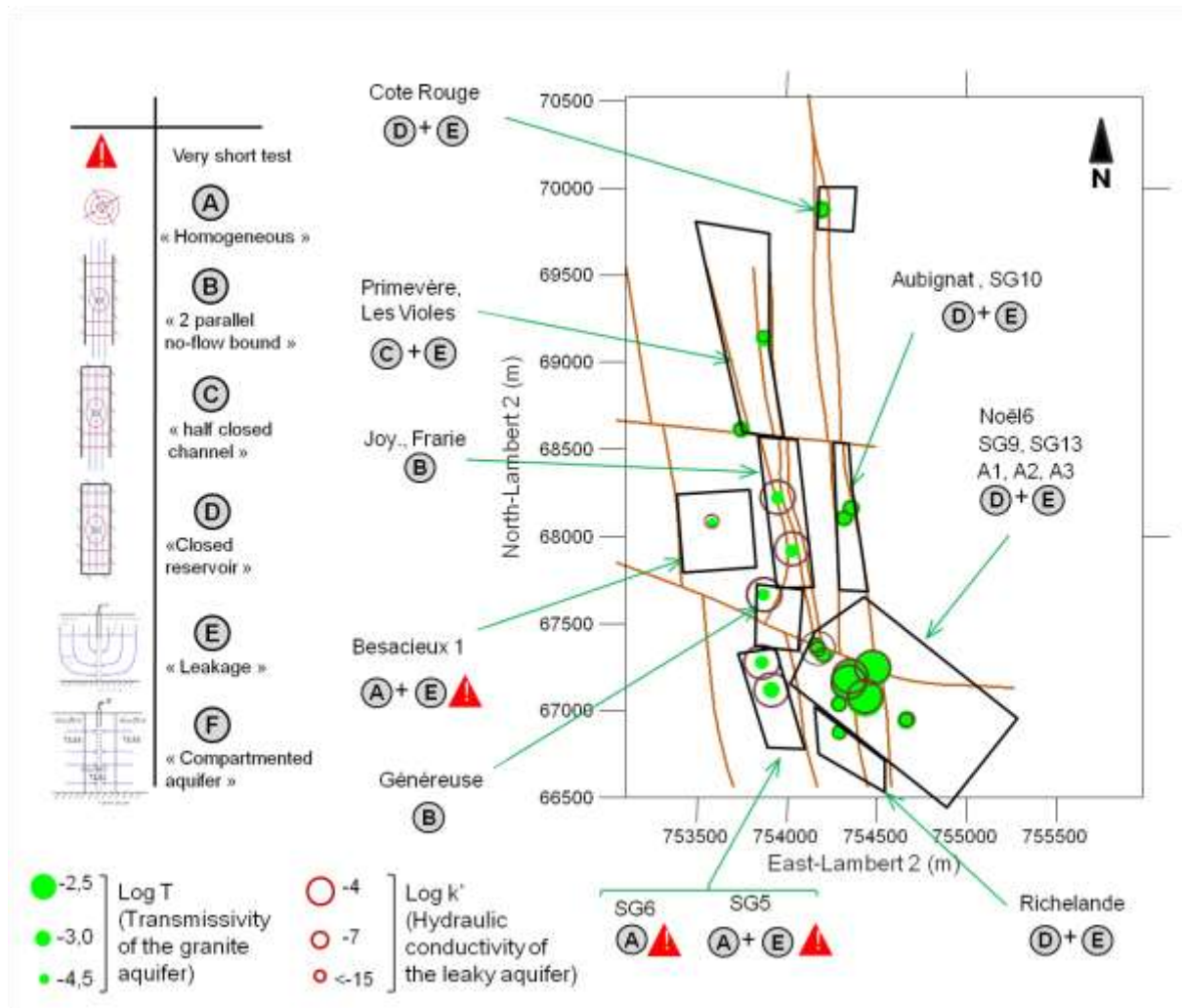
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879 Figure 4: a) Simplified geological and structural sketch of the narrow area deduced from the 3-D geological
 880 model. Location of the abstraction boreholes within the area and their mean pumping rates. Location of the
 881 historical spring (Fontfort) and the cored borehole (Flute 70); b) depth of the top of the granite' fissured layer
 882 and structural sketch. Location of the seismic survey lines. Scale in meters (Lambert 2 coordinate system).



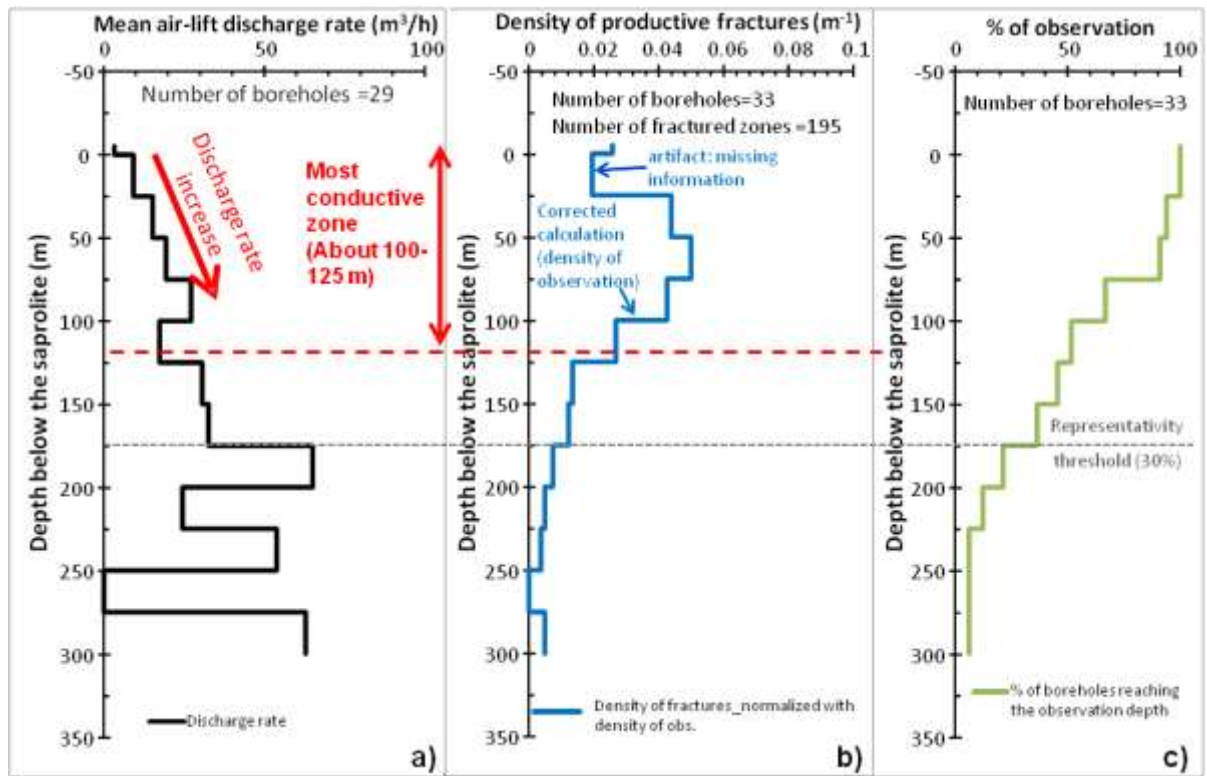
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884 Figure 5: Example of pumping test interpretation in granite. Diagnosis and modelling of the tests performed in
 885 A1 (1 test), A2 (1 test) and A3 (2 tests at variable depth, 134 and 153 m; and 1 long duration test in 2006)) from
 886 1979 to 2006. Model (A3-2006 only): Hunt and Scott (2007) and 4 no-flow boundaries and well effects.



887

888 Figure 6: Synthesis of pumping tests interpretation in the granite aquifer: Transmissivity (T) of the
 889 granite aquifer and hydraulic conductivity (K') of the leaky aquifer values (Log₁₀T scale). Diagnosis for
 890 each borehole (A to F), and identification of theoretical aquifer compartments limited by impervious
 891 boundaries (black boxes). Scale in meters (Lambert 2 coordinate system).

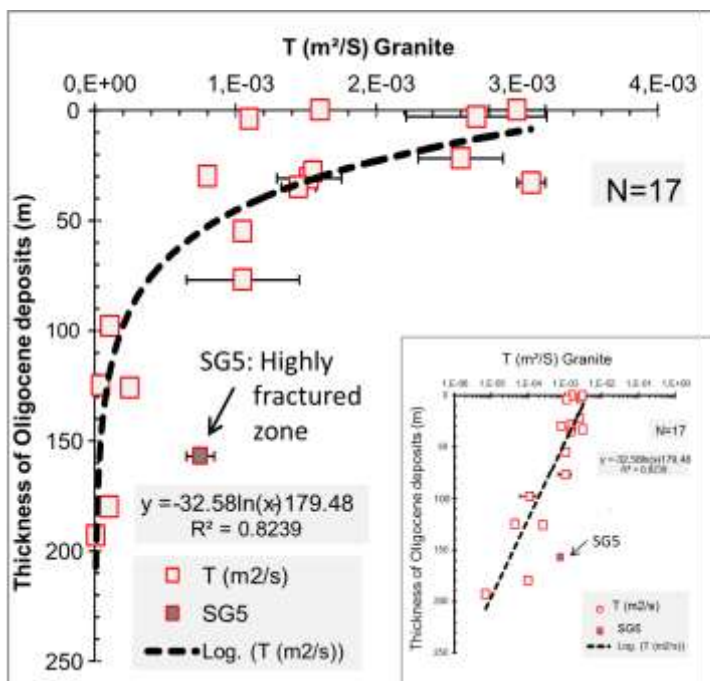


892

893 Figure 7: Air-lift analysis results for all the boreholes located into the granite aquifers (See Figure 2
 894 for the location of boreholes). The discharge rates are represented as a function of the depth below
 895 the bottom of the saprolite. Part c) allows to define the representativeness of the analysis (density of
 896 observation in % of the number of boreholes vs. depth). Parts a) and b) allow identifying the most
 897 permeable area. a) mean discharge rate, b) corrected fracture density (according to the density of
 898 observation, see explanation in the text). NB: these graphs are significant for at least a density of
 899 observation of 30%, i.e. until a 150-175 m depth under the bottom of the saprolite.

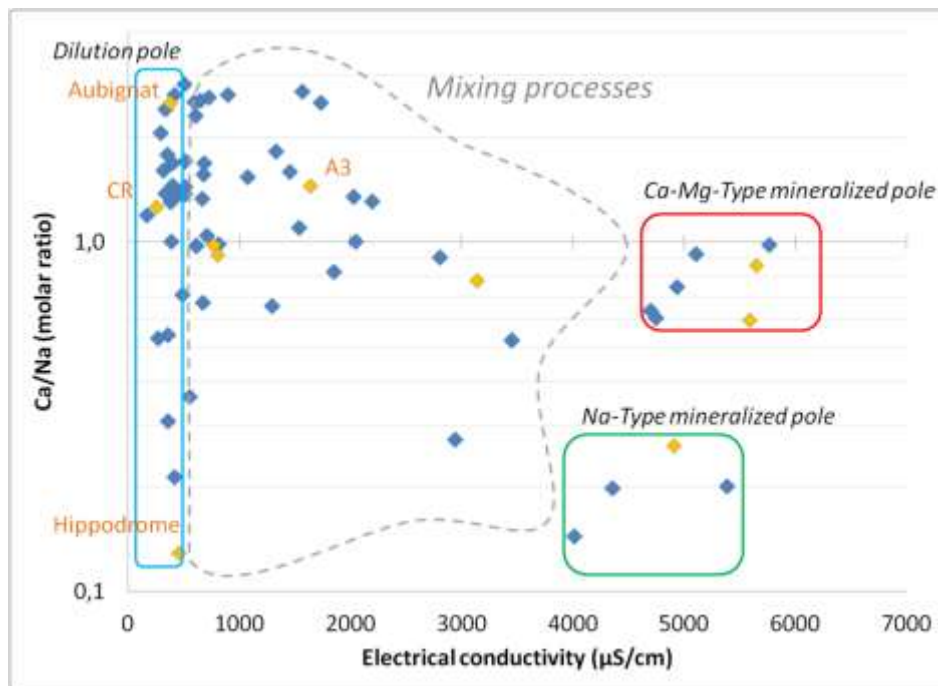
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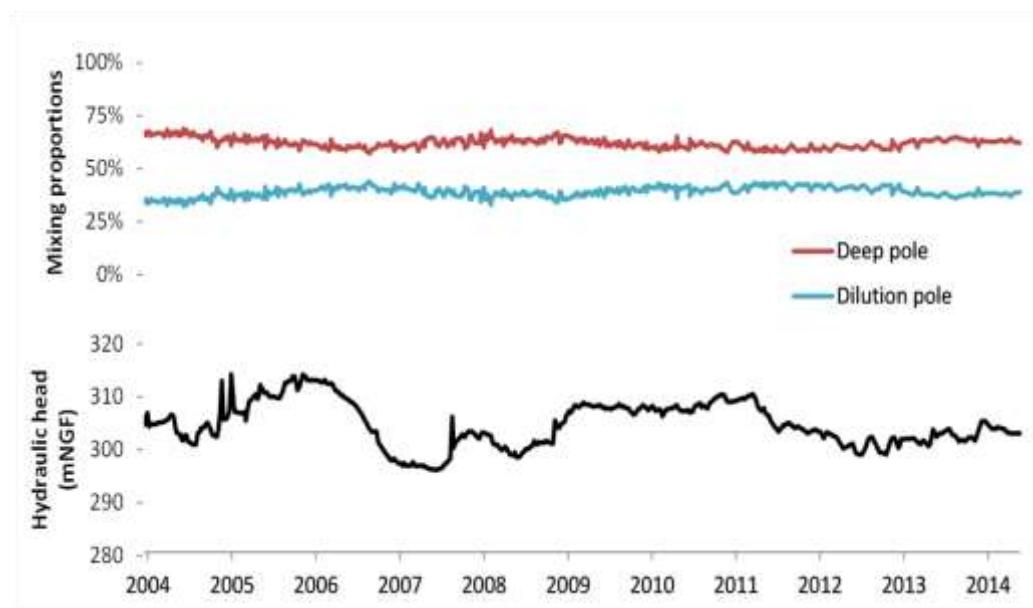
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903 Figure 8: Estimated transmissivity of the granite aquifer vs. thickness of the Cenozoic sedimentary
 904 deposits (Log scale representation for horizontal axis (T) in the lower right corner; N= 17, SG5 data
 905 not comprised). The granite's transmissivity decreases as a function of the sedimentary cover
 906 thickness.



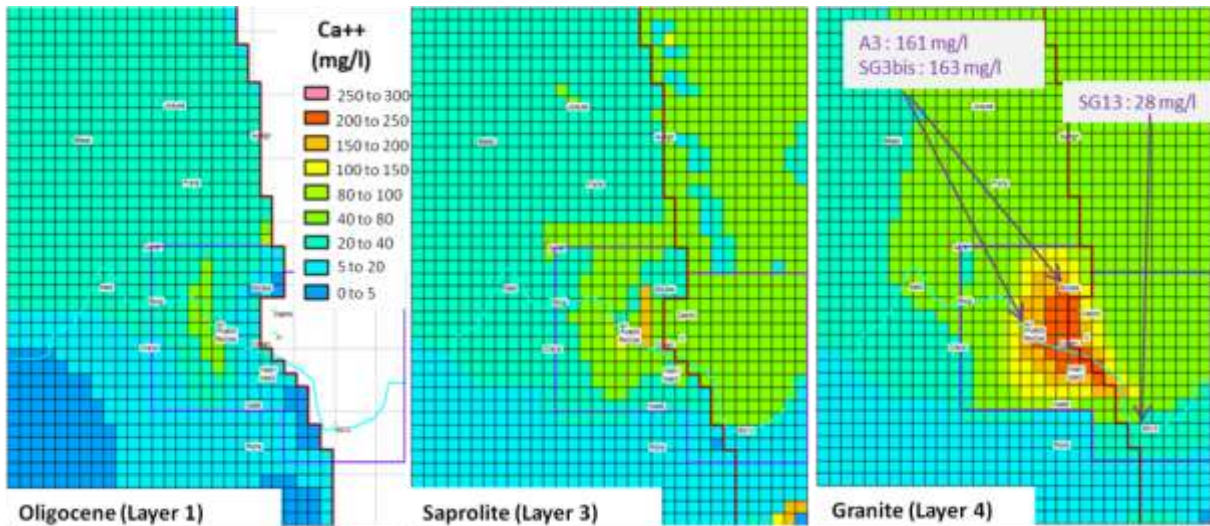
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913 Figure 10: Molar Ca/Na ratio vs. electrical conductivity of groundwater of the Saint-Galmier aquifer system. Ca-
 914 Mg-type and Na-type deep mineralized poles, dilutive pole and mixing processes are identified. Exploitation
 915 boreholes are represented in orange.



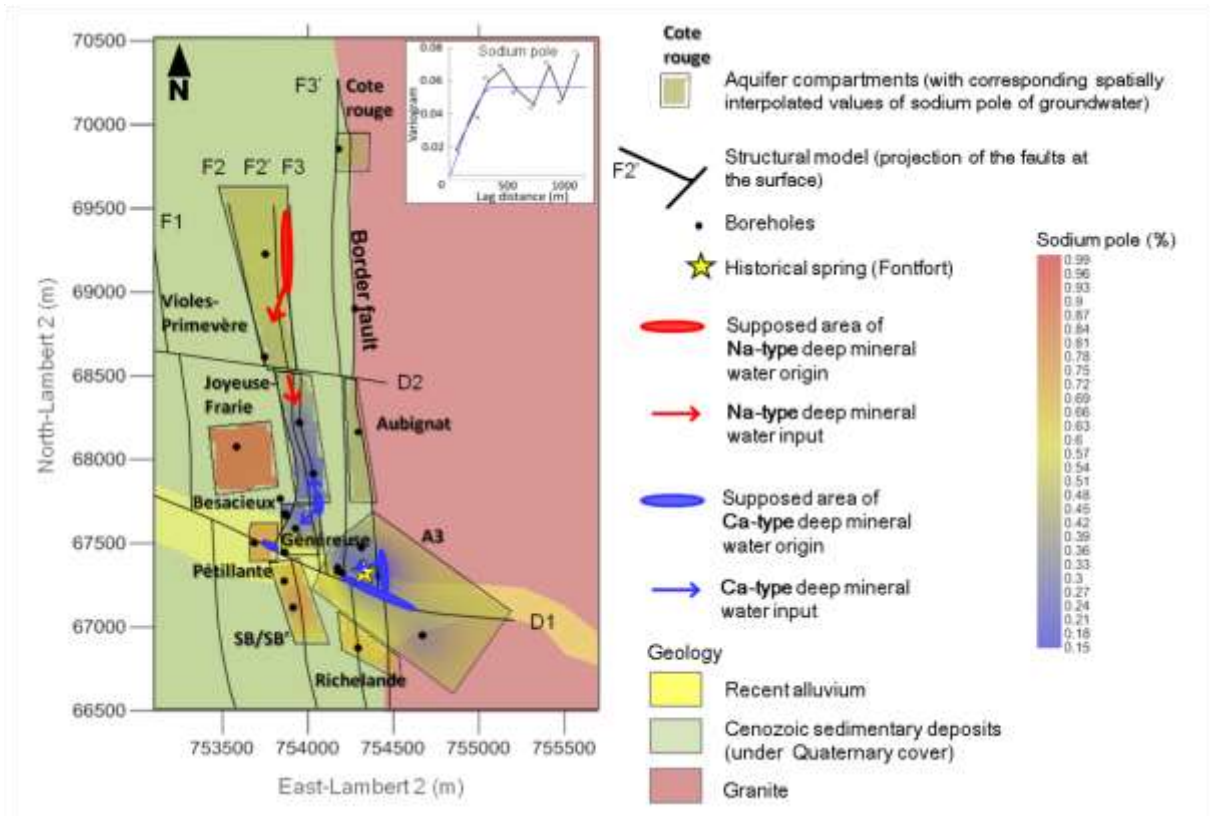
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917 Figure 11: Time series of the mixing proportions of the deep pole and the dilution pole, and corresponding
 918 hydraulic head of A3 borehole over the 2004-2014 period.



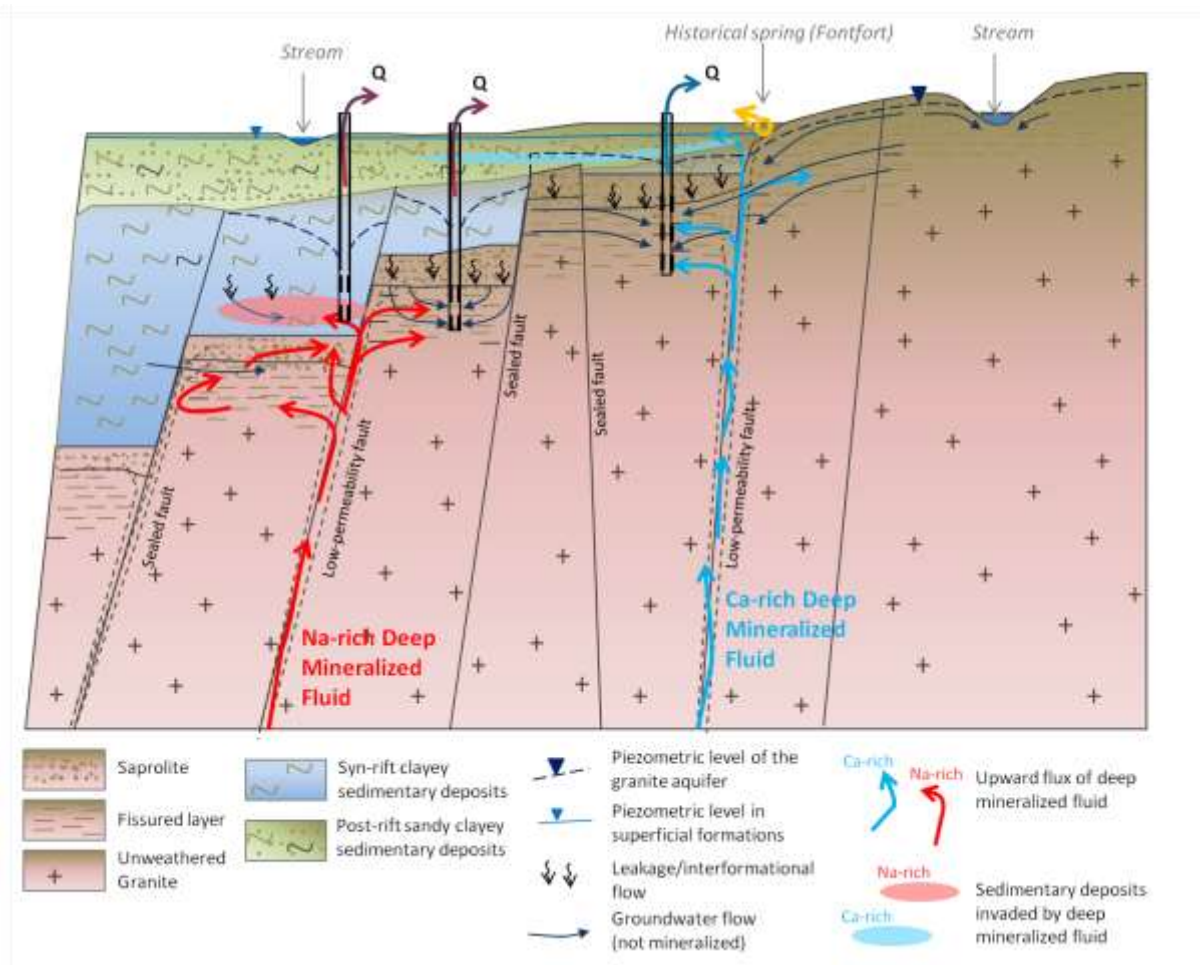
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920 Figure 12: Modelling results: Ca^{2+} content resulting from the mixing and transport of the two poles (i.e. deep
 921 pole and dilution pole) in the 3 aquifer layers, from left to right Cenozoic sediments, saprolite and fractured
 922 granite. See the location of the modelled area on Figure 9. The observed data are reported in the layer 4.



923

924 Figure 13: Conceptual model of the Saint-Galmier aquifer system. 10 aquifer compartments were identified
 925 (black boxes) with spatial interpolation (kriging) of the sodium pole values. The variogram of the sodium pole is
 926 represented in the upper right corner of the map. Impermeable fractures deduced from the structural model
 927 are represented by black lines. Few of these structures allow upflowing of Ca-Mg type (blue arrows) and Na-
 928 type (red arrows) deep mineralized water (See also Figure 14).



929

930 Figure 14: Cross-sectional view (E-W) of the conceptual model of the Saint-Galmier aquifer system.