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1 **Hydrogeological assessment of a deep-seated coastal landslide based on a** 2 **multi-disciplinary approach**

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11

12 **1. Introduction**

13 Over the last decades, the joint use of multi-disciplinary investigation methods has been widely disseminated in
14 the study of landslide behaviour (Caris and Van Asch, 1991; Jomard et al., 2007; Fressard et al., 2016; Denchik
15 et al., 2019; Di Maio et al., 2020). It is known that no single method, whether direct or indirect, is able to reliably
16 image the structure and dynamics affecting an unstable slope. A wide range of methods are available for
17 researchers or practitioners to achieve this purpose, such as borehole descriptions, laboratory tests, geophysical
18 surveys, and remote sensing techniques (Jongmans and Garambois, 2007; Chambers et al., 2011; Perrone et al.,
19 2014). For hydrologically controlled landslides, a major issue is also the characterization of the water circulation,
20 often affected by spatial and temporal variability, beyond the sole structural aspect (Malet et al., 2005).

21 Geophysical surveys are classically used for this type of study. The most common methods are seismic refraction
22 and reflection, geoelectrical methods (electrical resistivity and self-potential), electromagnetic methods, and
23 gravimetry. Each method has its own set of advantages and drawbacks (Frappa and Lebourg, 2001; Pazzi et al.,
24 2019). Depending on the method, the use of geophysics allows the imaging of the subsurface from one to four
25 dimensions (Jongmans and Garambois, 2007). These methods are most often used to study landslide structure,
26 including lithostratigraphic alternation, bedrock depth, geological boundaries, and slip surfaces (Bogoslovsky
27 and Ogilvy, 1977; Meric et al., 2005; Godio et al., 2006; Jomard et al., 2007; Göktürkler et al., 2008; Erginal et
28 al., 2009; Schmutz et al., 2009; Chambers et al., 2011; Coulouma et al., 2012). Geophysics is also of real interest
29 in determining the hydrogeological characteristics associated with a given structure (Grandjean et al., 2006;
30 Jomard et al., 2007; Lee et al., 2008; Niesner and Weidinger, 2008; Chambers et al., 2011; Gance et al., 2016).

31 Geoelectrical methods are classically used for landslide studies (Grandjean et al., 2006; Jongmans and
32 Garambois, 2007; Travelletti and Malet, 2012; Ekinci et al., 2013; Gance et al., 2016; Palis et al., 2017; Whiteley
33 et al., 2019). Electrical resistivity tomography (ERT) is classified as an active method and is most often used in
34 the form of two-dimensional profiles. It is designed to measure the electrical potential difference between two
35 pairs of electrodes by injecting a direct current into the soil (Kearey et al., 2002; Whiteley et al., 2019). ERT can
36 be used to assess the internal structure of landslides and highlight the moisture content of materials (Telford and
37 Sheriff, 1990; Jongmans and Garambois, 2007; Shevnin et al., 2007; Chambers et al., 2011; Denchik et al., 2019;
38 Pazzi et al., 2019). This method is particularly suitable to characterize the saturation of clay slides, which are
39 very sensitive to the presence of water.

40 Self-potential (or spontaneous polarization) is a passive geoelectric method. This technique has been
41 increasingly used in recent years in the field of hydrogeological studies and allows the characterization of the
42 solid-liquid interface of porous media (Colangelo et al., 2006; Jardani et al., 2006; Roubinet et al., 2016;
43 Cherubini, 2019) and highlighting of underground flows and drainage patterns (Revil et al., 1999; Chambers et
44 al., 2011). This method is defined as the measurement of the electrical potential distribution at the soil surface
45 without current injection (Naudet, 2004). SP measurements characterize a sum of potential electrical charges
46 (superposition principle), including thermo-electric, electrochemical and electrokinetic mechanisms (Corwin and
47 Hoover, 1979; Titov et al., 2002; Colangelo et al., 2006; Chambers et al., 2011). In the absence of telluric or
48 chemical disturbances, the electrokinetic phenomenon (or streaming potential) is expressed in the study of
49 landslides and is related to water circulation (Revil et al., 1999; Meric et al., 2006; Chambers et al., 2011). The
50 advantage of this method is the low weight of the device and the possibility of covering large areas in a short
51 time. Depending on the charge, electrical anomalies can be characteristic of either a draining and infiltration
52 zone (negative charge) or an accumulation and upwelling of an impermeable zone (positive charge) (Aubert,
53 1997; Perrone et al., 2004; Chambers et al., 2011). In particular, previous works have shown that negative SP
54 anomalies could correspond to a downward water movement in fractured surfaces or be related to an aquifer
55 drawdown (Jardani et al., 2009; Sujitapan et al., 2019).

56 ERT and SP are frequently used together to delineate preferential flow paths (Titov et al., 2002; Lapenna et al.,
57 2003; Jardani et al., 2006; Suski et al., 2006; Robert, 2012; Santoso et al., 2019). From a spatial point of view,
58 the scope covered by a 2D electrical tomography survey is often limited to two-dimensional transects, which are
59 often discontinuous despite possible overlaps. It does allow for in-depth imaging of the landslide structure and
60 the limits of the vadose zone, even in complex formations and on hilly slopes. The SP allows the characterization

61 of the flow dynamics affecting a landslide over a continuous area, under the sole condition of accessibility.
62 Therefore, it has the ability to validate and spatially extend the hydrogeological hypotheses made from ERT
63 surveys by providing hydrokinetic information.

64 However, geophysical data must be used in conjunction with other investigative methods (Chambers et al., 2011;
65 Perrone et al., 2014). In particular, they must be combined and calibrated with geological, geotechnical, and
66 geomorphological data and observations to improve the interpretation reliability (Jongmans and Garambois,
67 2007). It is therefore important to couple data sets that are sometimes very heterogeneous. Geological and
68 geomorphological knowledge is crucial for validating survey interpretations but also for assessing the observed
69 phenomena and their functioning (Hearn and Hart, 2011). The study of hydrogeological behaviour also requires
70 a focus on surface and subsurface flows. In this respect, hydrogeochemical analyses serve to highlight the
71 internal mechanisms and the flow of water by studying, for example, major ions or physicochemical
72 characteristics (Bogaard et al., 2004, 2007; Montety et al., 2007; Bertrand et al., 2013; Marc et al., 2017).
73 Artificial and isotopic tracing techniques are also classically used to help dissociate the origin of waters in an
74 unstable slope (Binet et al., 2007; Deiana et al., 2020).

75 The datasets to be integrated can be numerous and are sometime heterogeneous. The consistency of temporally
76 and spatially disparate multi-source data with sometimes uneven reliability is an important issue for the
77 characterization and conceptualization of landslide functioning (Bichler et al., 2004; Travelletti and Malet, 2012;
78 Lissak et al., 2014b; Fressard et al., 2016; Denchik et al., 2019). This is particularly the case for long-term
79 studied landslides, for which the data are sometimes discontinuous and scattered.

80 The study site discussed in this paper is a typical case of a complex landslide under hydrological control with a
81 high sensitivity due to the presence of clays. One of its significant specificities is due to its coastal character.
82 This site was the subject of several studies for more than 40 years, has been well documented and described from
83 structural and kinematic points of view (Maquaire, 1990; Lissak, 2012). Although the hydrological influence on
84 landslide activity has been proven, the hydrogeological functioning of the slope still begs many questions. The
85 present research aims to improve the understanding of internal flows by the joint use of existing data sets
86 (electrical tomography, geotechnical and geomorphological data), completed and supplemented by new
87 investigations (self-potential and physicochemical studies of water). Thus, through an integrative multi-
88 disciplinary approach, the objective is to homogenize this set of qualitative geotechnical, geophysical and
89 hydrological data and analyse them to propose a new conceptual hydrogeological model.

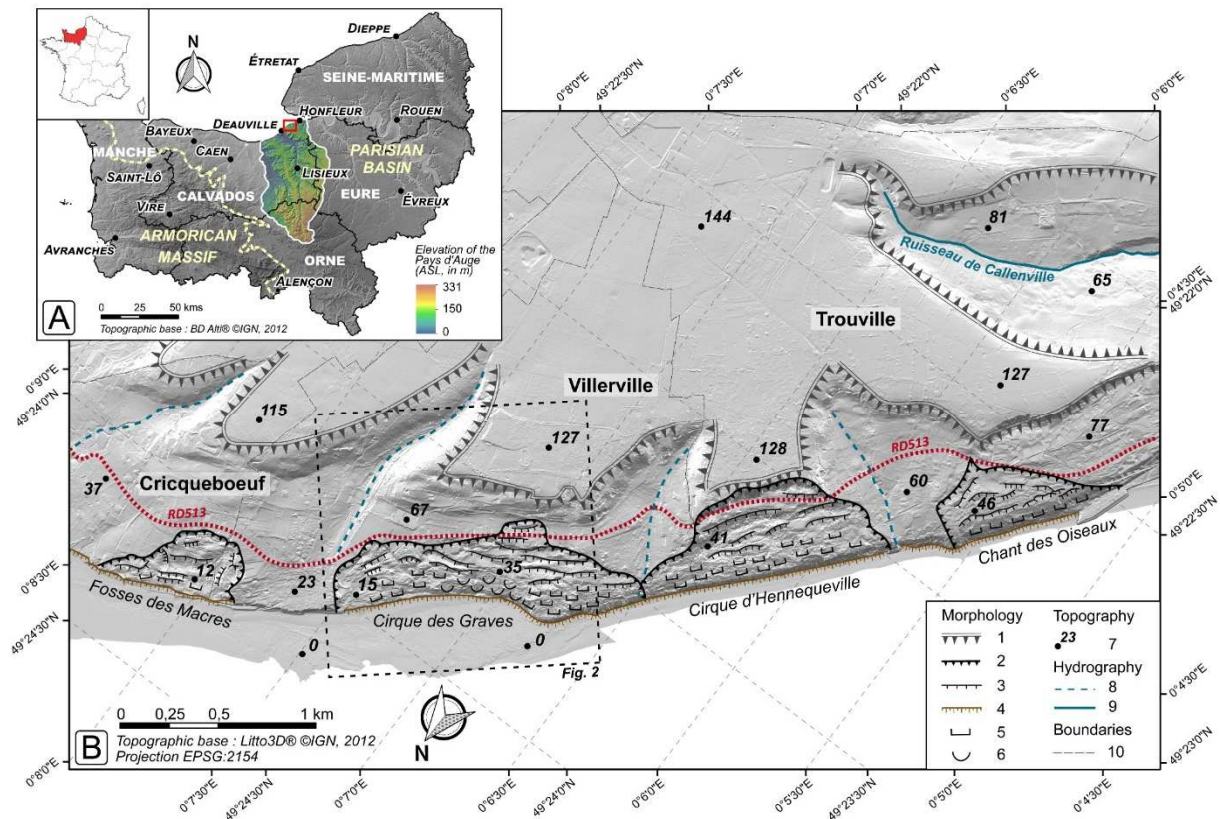
90 The expected results from this work are manifold. It aims to (1) identify the preferred discharge locations from
91 the continental water table; (2) understand the preferential water flows in the landslide; (3) highlight the degree
92 of heterogeneity of the hydrogeological functioning of the site; and (4) establish a link between morpho-
93 structural specificities and the behaviour of the landslide with regard to internal flows.

94

95 **2. Description of the site and prior knowledge**

96 **2.1. Study site**

97 The Cirque des Graves is located in the northern part of the Pays d'Auge (Normandy, France), along the coast of
98 the English Channel. It is the largest of the four large landslides affecting the coastline between the
99 municipalities of Trouville and Cricquebœuf (Fig. 1). These landslides are deep and rotational-translational and
100 were initially triggered during the last Weichselian glaciations (Elhaï, 1963; Juignet et al., 1967; Journaux, 1971;
101 Fressard, 2013). They are also all active due to the erosive action of the sea, exerting a continuous undermining
102 effect and not allowing the slope to reach an equilibrium profile (Maquaire, 1990). This landslide of 47 ha
103 (1 600 m long and 450 m wide) is adjacent to the Pays d'Auge plateau and surrounded by two non-permanent
104 watercourses at its southwest and northeast boundaries. Its anthropization is recent as it dates back to the end of
105 the 19th century (Lissak, 2012). Despite the destruction of a large part of the historical infrastructure due to
106 previous accelerations, approximately twenty individual houses and a major tourist road (RD513) are still
107 located on the unstable area.

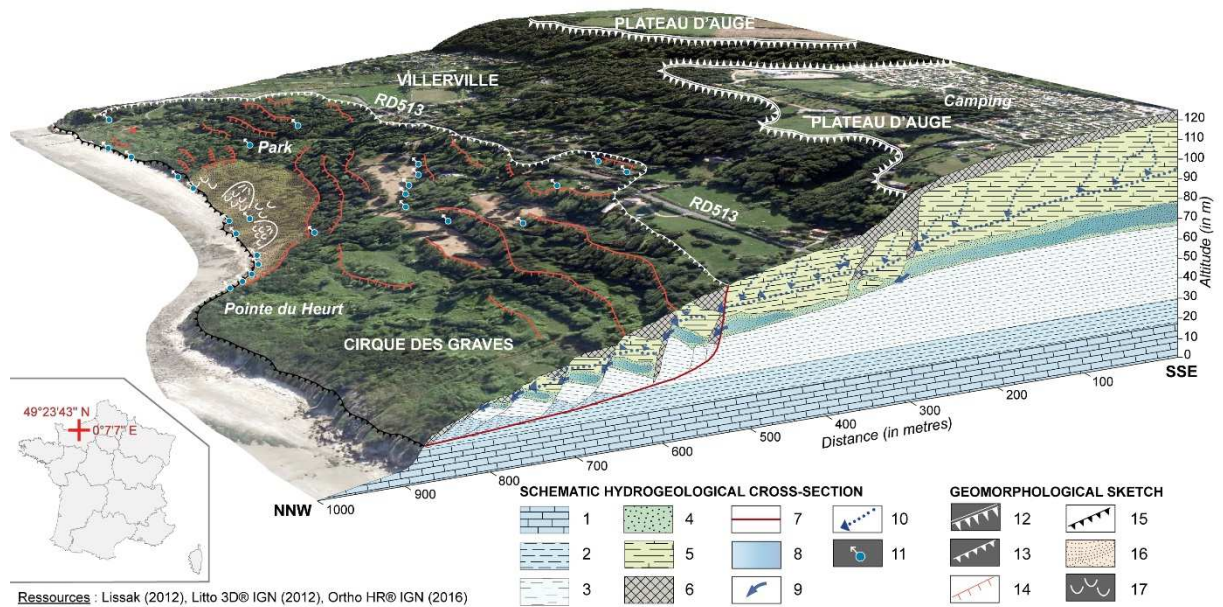


108

109 *Fig. 1 – A: Location of the study field in Normandy; B: Morphological sketch of the four major coastal landsides of the Pays*
 110 *d’Auge, between Deauville and Honfleur. 1. Border of the Plateau d’Auge; 2. Main scarp of the landslides; 3. Secondary*
 111 *scarp; 4. Basal scarp; 5. Bumpy morphology; 6. Solifluction lobes morphology; 7. Spot height (in meters); 8. Temporary*
 112 *river flow; 9. Permanent river flow; 10. Municipality boundary.*

113 The lithostructure of the unstable area is more complex than on the plateau due to the retrogressive kinematics of
 114 the slope generating discontinuities. From the plateau boundary to the sea, dissociated chalk slabs sink into
 115 Albian-Aptian sands and slide over Kimmeridgian and Oxfordian marly clay formations (Fig. 2). These
 116 formations, with fairly poor geotechnical characteristics, are underlain by a bedrock composed of Oxfordian
 117 gritty limestone (Maquaire, 1990). In the light of its size, the sharpness of the main scarp, the processes affecting
 118 the slope, and the landforms, this landslide can be classified as complex with rotational and translational
 119 components (Varnes, 1978; Cruden and Varnes, 1996; Lissak, 2012).

120



121 Ressources : Lissak (2012), Litto 3D© IGN (2012), Ortho HR© IGN (2016)

122 Fig. 2 – Schematic functioning of the Cirque des Graves proposed by Maquaire (1990) and Lissak (2012). 1. Oxfordian gritty

123 limestone of Hennequeville; 2. Oxfordian clay of Villerville; 3. Kimmeridgian marl; 4. Albian sands; 5. Cenomanian chalk;

124 6. Surficial deposits (loess, colluvium, head); 7. Limit of the active landslide; 8. Albian sand groundwater; 9. Albian

125 groundwater flow; 10. Cenomanian karstic flow; 11. Main springs; 12. Main scarp of the plateau; 13. Main scarp of the

126 cirque (uphill limit of the landslide); 14. Secondary scarp; 15. Foot scarp; 16. Shallow landslide/muddy flow; 17. Denting

127 and solifluxion.

128

129 2.2. Hydromechanical background

130 The first studies carried out in the Cirque des Graves date back to the 1930s, but a real research interest emerged

131 after the major event of 1982 (Fig. 3 h) and focused on the geotechnical and kinematic response for better

132 management of the infrastructure facilities (housing, networks, etc.). Since this first recent reactivation, three

133 other events impacted the cirque in 1989, 1995 and 2001 (Maquaire, 1990; Lissak et al., 2013, 2014a; Costa et

134 al., 2019).

135

136 *Morpho-structural knowledge*

137 The global structure of the landslide was determined thanks to numerous boreholes and geomorphological

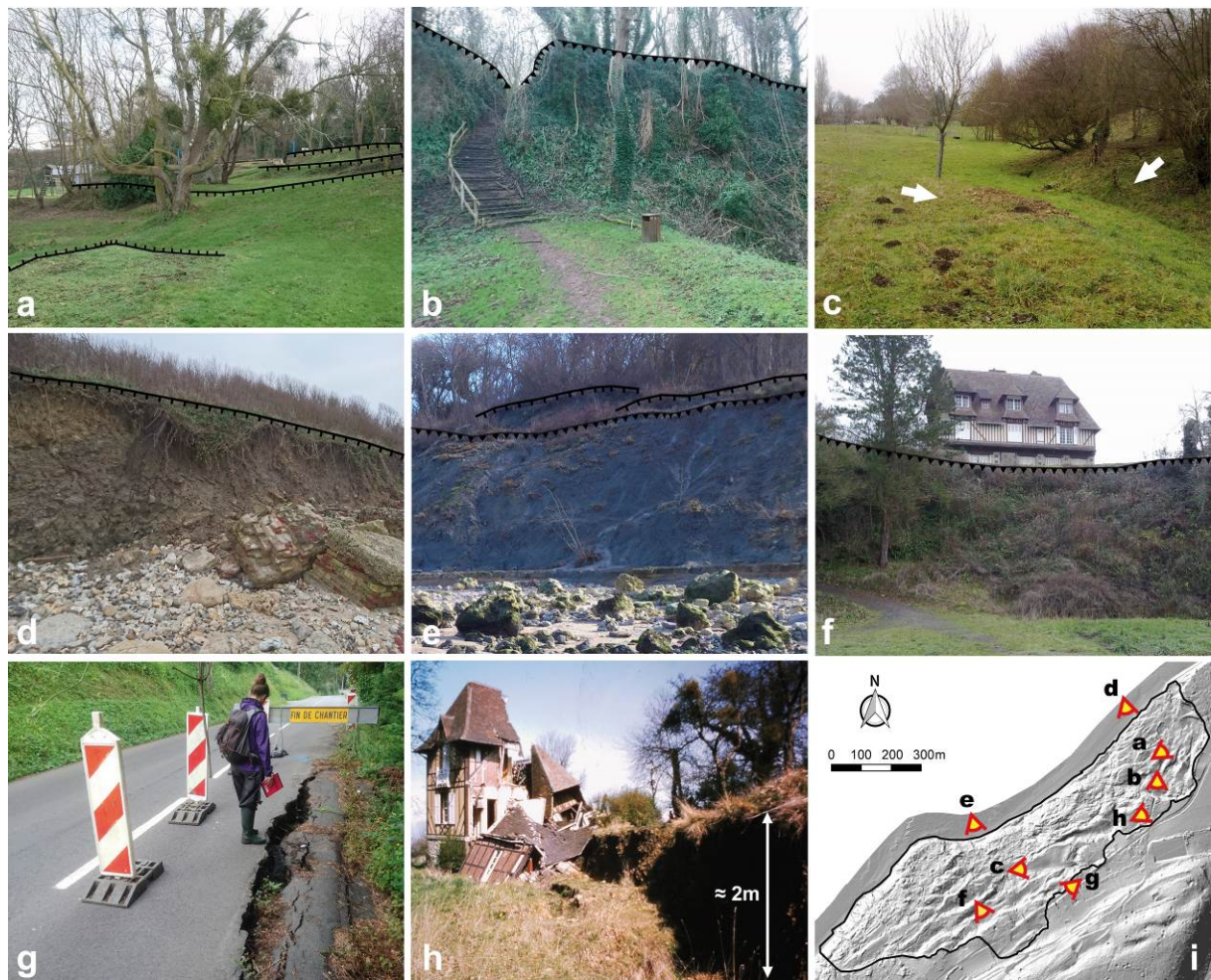
138 analyses (Maquaire, 1990; Lissak, 2012; Lissak et al., 2014b). In addition, electrical tomography surveys

139 confirmed the thickness of the different features over 18 transects using vertical and horizontal variations in

140 resistivity (Lissak et al., 2014b).

141 The litho-stratigraphy is generally consistent with the morphological observations (Lissak et al., 2014b). The
142 numerous main and secondary scarps, ranging in height from a few decimetres (Fig. 3 a) to more than 15 m
143 upstream of the western part (Fig. 3 b, f), correspond to a segmentation of the chalk by weathering, inducing a
144 dissociation into approximately 100 discrete slabs (Lissak, 2012; Lissak et al., 2014b). Their size increases from
145 east to west (Lissak et al., 2014b). Tensile zones between the slabs are filled with silty-clay formations.
146 Downwards, the thickness of the Albian-Aptian sands is uneven due to the importance of creep and the sinking
147 of the chalk slabs in this loose and poorly cohesive horizon.

148



149
150 *Fig. 3 – Main geomorphological features of the Cirque des Graves. (a) Minor scarp and stepped morphology in the park; (b)*
151 *Major scarp (chalk outcrop) at the upstream part of the park (c) Slab tilted into a counter-slope at the centre of the landslide;*
152 *(d) Heterogeneous soil matrix with blocks (heads) at the basal scarp of the park; (e) Mudflows in the Oxfordian marls at the*
153 *foot of the landslide; (f) Building near a main scarp in the middle of the landslide; (g) Recent crack on the RD513 due to the*

154 *retreat of the main scarp; (h) Destruction of the villa Chanteclair after the acceleration of 1982 (Maquaire, 1984); (i)*
155 *Location and orientation of the pictures.*

156

157 The site is characterized by a chaotic morphology with a succession of scarps, counter-slopes, horst-graben
158 structures, and general hummocky topography (Lissak et al., 2014b) (Fig. 3, b, c). The downstream part of the
159 central and western areas is also characterized by the absence of chalk slabs (Fig. 8) and a solifluction dynamic
160 of Oxfordian grey clays (Fig. 3 e).

161 The global volume of the slide was estimated at approximately $3e10^7$ m³ (Lissak, 2012; Lissak et al., 2014b).
162 The slip surfaces are deep, nested, and affect all of the horizons down to the Kimmeridgian and Oxfordian clays
163 (Maquaire, 1990, 2000; Maquaire and Malet, 2007). The slip surface was characterized by inclinometric surveys
164 to the east of the slide and measured at a depth of 23 m b.g.l. upstream and 5.5 m b.g.l. at the bottom of the slope
165 (Maquaire, 1990). However, no inclinometer was able to characterize the deepest slip surface in the middle and
166 western zones of the cirque because the devices were not implemented deep enough.

167

168 ***Kinematic knowledge***

169 From a kinematic point of view, topographic monitoring was carried out using (1) a total station (Maquaire,
170 1990), (2) GPS monitoring on fixed benchmarks and (3) the implementation of permanent GNSS receivers
171 (Lissak, 2012; Lissak et al., 2014a). These follow-ups revealed continuous activity with periodic accelerations
172 and uneven dynamics (Maquaire, 1990; Lissak, 2012; Lissak et al., 2014b). The mean annual velocity ranges
173 from 1-2 cm. yr⁻¹ near the main scarp to 10-12 cm. yr⁻¹ at the foot of the slope (Lissak, 2012). The most active
174 zone is located at the foot of the middle zone, which is subject to clay mudflows (Fig. 3 e). The general dynamic
175 is retrogressive with morphological downstream readjustments (Fig. 3 g), accentuated by basal sea excavation
176 (Fig. 3 d, e).

177

178 ***Hydrological knowledge***

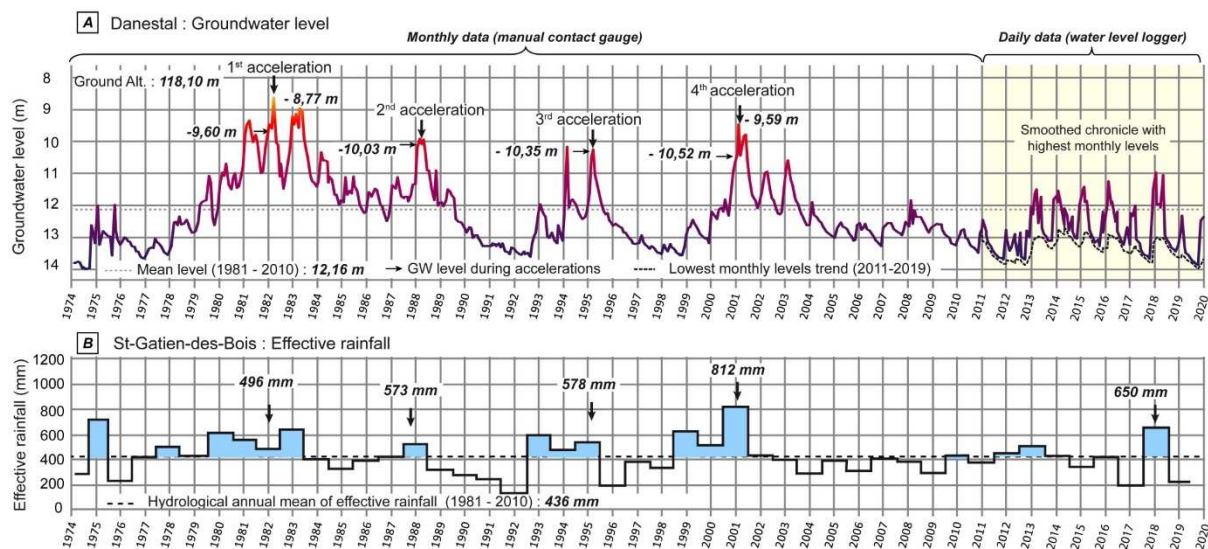
179 Previous work has shown a direct link between the rise of the chalk water table on the Plateau d'Auge (at the
180 Danestal piezometer, 17 km away from the slide) and the four major accelerations of 1982, 1989, 1995 and 2001

181 (Maquaire, 1990; Lissak, 2012; Bogaard et al., 2013; Lissak et al., 2014b; Costa et al., 2019) (Fig. 4). Although
 182 the levels between the plateau and the slope appear to be correlated, there is a delay of a few days to weeks due
 183 to the transit time of water through the matrix before the continental water table discharges on the coast (Lissak
 184 et al., 2014b).

185 Exceeding the trigger threshold to achieve a major acceleration requires several years with higher than normal
 186 annual rainfalls (Fig. 4 B). Thus, the major event of 1982 was preceded by six years of high annual precipitation.
 187 This expresses a cumulative effect of the precipitation on groundwater levels from one year to the next and
 188 underlines the major role of the continental discharge in exceeding the trigger levels (Maquaire, 2000).

189 Site-wide water flows are considered erratic with interconnected aquifers due to numerous structural
 190 discontinuities (Lafenetre, 2010; Lissak, 2012).

191



192

193 *Fig. 4 – Correlation between the four major landslide accelerations in the Pays d'Auge and (A) the water table level at the*
 194 *Danestal piezometer. Because the chronicle is irregular from 2011 onwards (change of acquisition method), the data have*
 195 *been smoothed to show the extreme monthly levels; (B) the effective annual rainfall at the Météo-France weather station in*
 196 *St-Gatien-des-Bois, from 1974 to 2020 (updated from Lissak, 2012 and Costa et al., 2019).*

197 The site has been equipped with a network of piezometers since the 1980s to allow groundwater monitoring.
 198 There is a slight temporal variability between high and low waters despite a significant spatial variability
 199 between some neighbouring piezometers that is directly related to the geological structure (Lissak, 2012).

200 Indeed, the hydraulic conductivity is variable depending on the material. Cohesive chalk and sandy formations
201 are most permeable. Their values range from $1.5e^{-6} \text{ m.s}^{-1}$ to $1.8e^{-4} \text{ m.s}^{-1}$, while clays and surface flint clays have
202 the weakest hydraulic conductivity of between $3.9e^{-9} \text{ m.s}^{-1}$ and $5.9e^{-7} \text{ m.s}^{-1}$ (Lafenetre, 2010; Lissak, 2012).

203 Hydromechanical modelling work has shown that a 1-m rise in the water table results in a 6% decrease in the
204 overall safety factor of the slope (Maquaire, 2000; Maquaire and Malet, 2007). This demonstrates a real
205 importance of the internal circulation of water on the landslide activity and confirms the hydrologically driven
206 functioning of the slide.

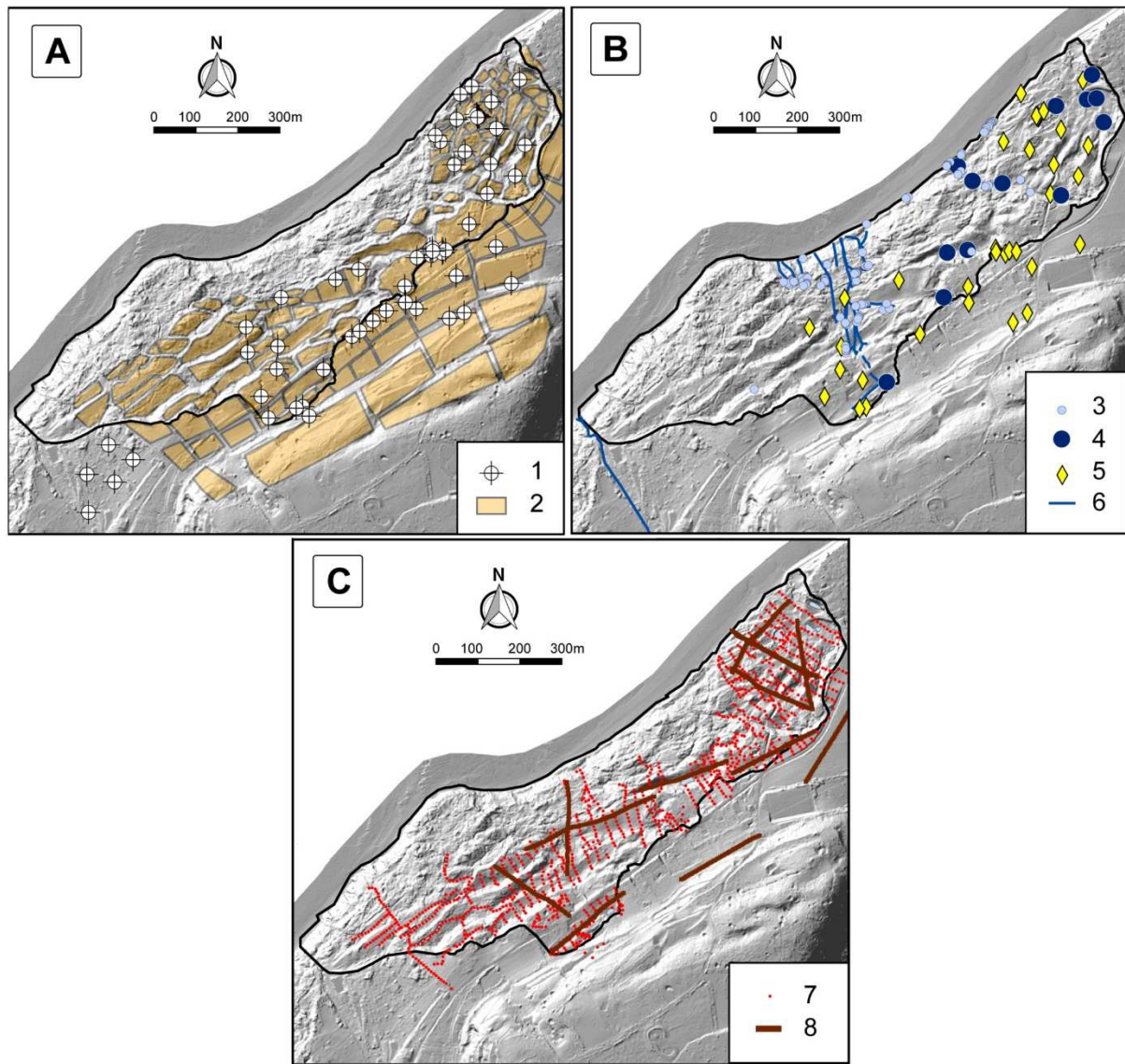
207

208 **3. Materials and methods**

209 In light of the work described in the previous section, some data could be reused for joint analysis with the new
210 investigations carried out. Fig. 5 shows the nature and location of all of the input data. The uneven spatial
211 distribution of the data is partly due to the inaccessibility of part of the cirque (high scarps, fallow land, mudflow
212 areas, etc.).

213 The methodological details for a re-analysis of the initial data and new hydrokinetic data are detailed hereafter
214 by type of acquisition.

215



216

217 *Fig. 5 – Typology and spatial distribution of the Cirque des Graves data used for morphostructural (A), hydrological (B) and*
 218 *geophysical (C) assessment. 1. Drillings from 1978 to 2019; 2. chalk compartment delineation (from Lissak, 2012); 3.*
 219 *spring/resurgence; 4. pond; 5. piezometric measurements; 6. drainage network; 7. SP measurement (2018 campaign); 8.*
 220 *ERT profile (2008/2009 survey).*

221

222 3.1. Structural predisposition assessment

223 At first, our analysis consisted of synthesizing and reinterpreting borehole descriptions from multiple
 224 geotechnical analyses carried out since the 1970s. Logs of 62 boreholes drilled between 1978 and 2019 have
 225 been collected and used to assess the characteristics of the chalk (condition and thickness). These boreholes are
 226 mainly distributed in the eastern and upstream part of the site (Fig. 5A), characterizing the areas with stakes. One

227 issue was to dissociate the chalk and non-chalk horizons because saturated and weathered chalk is rarely
228 described as such.

229 The assessment of chalk is crucial because thick and cohesive chalk generates cracks and sharp discontinuities.
230 In the Cirque des Graves, these cracks are filled with heterogeneous and loose formations (Lissak, 2012). It
231 significantly increases the transmissivity and recharge of aquifers (Foster and Milton, 1974). The transmissivity
232 of a rock is defined by its void volume, which in turn depends on its matrix porosity and fissure porosity (Sausse,
233 1998). In the case of chalk, which is a material with low elasticity and subject to karstification, the fissure
234 porosity can represent a significant part of the total flow. Therefore, this material illustrates heterogeneous
235 behaviour (Buscarlet et al., 2011) with variable permeabilities and generally slow flow but also very localized
236 fast flows (Crampon et al., 1993). This particular stratum is therefore critical for assessing the hydrogeological
237 functioning of the slope.

238

239 **3.2. Surface and groundwater study**

240 Two physicochemical water characterization campaigns were carried out. The first one was conducted during a
241 low water period (October 2018) and the second during a high water period (April 2019) to compare the results
242 at hydrological extremes. The objective was to measure descriptive parameters such as water flow, pH index,
243 conductivity and temperature in all ponds, springs and drained waters that were accessible and identified (Fig.
244 5B).

245 In situ measurements were performed using a TetraCon® 325 sensor for conductivity ($\pm 1 \mu\text{S}\cdot\text{cm}^{-1}$) and a
246 Sentix® 41 electrode for pH index (± 0.01) and temperature ($\pm 0.1^\circ\text{C}$), which were both connected to a
247 WTW® 340i measuring device (instruments by WTW® GmbH, Weilheim, Germany). Flow measurements were
248 carried out by collecting water from springs and drains using suitable containers. The containers were filled over
249 a stop-time period, and the quantities were measured in graduated test tubes. The locations of the measurement
250 points were determined using a Trimble Juno 3B lightweight GPS field device.

251 These measurements allow for characterization of all areas where the aquifer is outcropping on the landslide.
252 The locations of the measurement points were coupled to a set of 34 piezometric measurements carried out over
253 the entire site. The currently operational network is composed of 13 piezometers and 7 inclinometers, followed
254 by continuous piezometric sensors or a manual contact gauge. In addition, data from 14 other piezometric

255 measurement locations could be retrieved and integrated into the analysis. These data are partly from old
256 piezometers no longer in use and partly from exploratory drilling enabling spot measurements of the
257 groundwater level. The latter only provide point information on the piezometric levels. Because we know that
258 the variations in the water table are low on a seasonal scale but exhibit strong site variability (Lafenetre, 2010;
259 Lissak, 2012), it seems interesting to include them in the analysis.

260 The objective here is to characterize not the temporal variability but rather the spatial variability of the
261 unconfined aquifer, which is directly related to the lithostructure. This is a first approach to try to highlight the
262 preferential water flow paths over the unstable area.

263 A total of 97 groundwater measurements were used to produce a map of the mean piezometric surface. An
264 ordinary kriging probabilistic interpolation method was applied, which considers the spatial dependency of the
265 data while minimizing prediction errors (Matheron, 1963; Oliver and Webster, 1990; Baillargeon, 2005). This
266 method is assumed to be suited to conventional hydrological problems (Gambolati and Volpi, 1979).

267

268 **3.3. Geophysical imaging methods**

269 *Self-potential surveys*

270 To refine internal flows, we used geophysical imaging tools. Self-potential is a suitable method for highlighting
271 disturbances in electric currents and potential fields associated with water circulation (Gex, 1990; Naudet et al.,
272 2008; Revil and Jardani, 2013). Natural electric fields originate directly from the infiltration of water through the
273 process of electrofiltration. When water flows through a rock matrix, then the positive ions are fixed while the
274 negative ones are carried away by the water. This creates an ion imbalance, which induces an electrical anomaly
275 that can be measured by electrodes on the ground, thus making it possible to explain the orientation and intensity
276 of the flow in said matrix (Naudet, 2004).

277 A self-potential mapping was carried out in April 2018 at the end of the high-water period. Some measurements
278 have been performed using a Fluke175 high-impedance multimeter and two WM "Wolf LTD" ceramic
279 electrodes saturated with Pb-PbCl liquid (lead-lead chloride), which cancels polarization effects. A total of 1,171
280 measurement points was realized with a fixed base configuration (one reference electrode and one moving
281 electrode). The theoretical spacing of the survey was 10 m between each point on upstream-downstream profiles

282 and 25 m between two successive parallel profiles. Some adjustments had to be made according to accessibility
283 conditions, particularly in the western zone of the landslide.

284 To avoid drifts due to the excessive distance between the electrodes, the location of the base station was shifted
285 every five profiles (≈ 125 m). Between each base station, a measurement of the anomalies was made so that all
286 values could be reported at the starting point of the survey. We also carried out numerous check points on the
287 previous profiles each time that the base electrode location changed so that we can interconnect the survey lines
288 and obtain a better monitoring of drifts and errors.



289
290 *Fig. 6 – Self-potential device: (A) Derivation control between fixed and mobile electrode 1. High-impedance multimeter; 2.*
291 *Mobile electrode; 3. Fixed electrode protected by the silver side of a survival blanket (to reflect solar radiation and limit the*
292 *thermal drift). The blanket is then covered with clumps of earth during the survey; 4. Bucket with bentonite slurry. (B)*
293 *Measurements of SP anomaly in the sliding area. 5. Cable reel connecting both electrodes; 6. GNSS acquisition to localize*
294 *SP measurements.*

295 The overall survey reference was located upstream of the park, outside of the unstable slope. We buried the fixed
296 electrode 30 cm under the ground, protected from solar radiation to limit thermal drift or drifts due to ground
297 drying (Fig. 6A).

298 Bentonite mud was systematically used to improve the electrical contact between the electrode and the ground.
299 In addition, the positions of the measurement points were collected by GNSS surveying using a Trimble R8
300 differential GPS (Fig. 6B). 65% of the slide surface was covered, and the remaining 35% were not accessible.

301 A validation survey was conducted in April 2019 to check the reliability of the initial survey. To limit the
302 acquisition time and the associated drifts, we chose to establish two lateral profiles and one longitudinal profile
303 for a total of 225 measurement points.

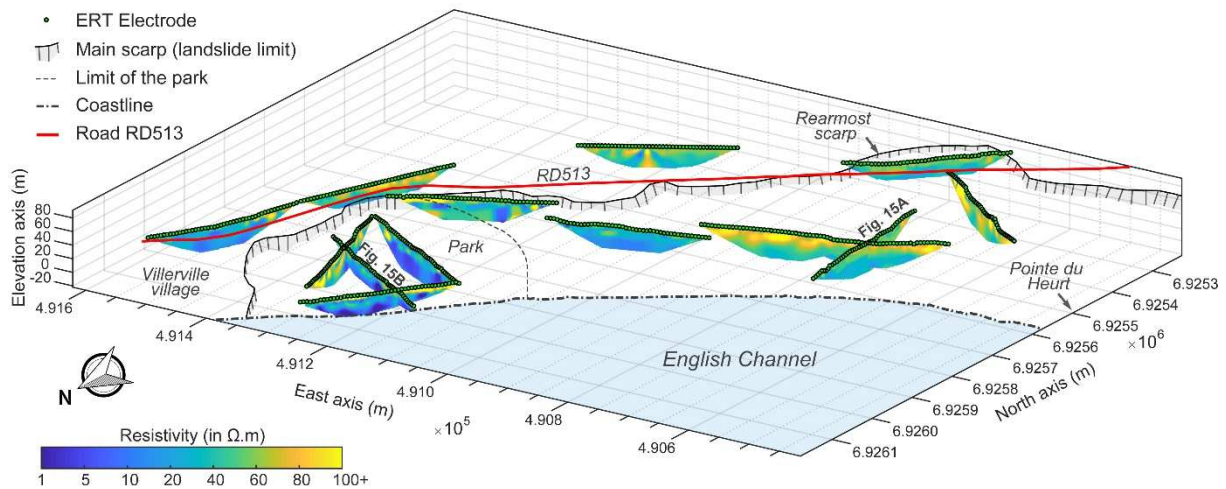
304 All data points were interpolated using kriging, which is also a conventional and suitable method for
305 interpolating geophysical data (Grandjean et al., 2006; Travelletti and Malet, 2012; Jaboyedoff et al., 2020). The
306 regional model was filtered to eliminate the topographic effect and make local anomalies more apparent (Goto et
307 al., 2012). The topographic influence was considered constant and calculated using a linear regression equation
308 over all survey datapoints. This regression equation was applied to the DTM to define the theoretical influence
309 of topography at each point of the raster. Finally, this influence was subtracted from the raster of measured raw
310 anomalies.

311

312 *Cross-analysis with ERT and geological data*

313 In the second phase, the ERT campaign conducted in 2008-2009 in the frame of the SISCA project (Lissak et al.,
314 2014b) was also reused. It aims to compare resistivity information, which depends on saturation and
315 petrophysical characteristics, with hydrokinetic information from self-potential. The local materials of the Pays
316 d'Auge are generally not very resistive. Chalk is above 60 Ω .m (Göktürkler et al., 2008; Naudet et al., 2008),
317 while clays, sands and reworked materials are rather below 30 Ω .m (Perrone et al., 2004; Van Den Eeckhaut et
318 al., 2007; Naudet et al., 2008; Jongmans et al., 2009; Sudha et al., 2009; Travelletti and Malet, 2012). Despite
319 this slight distinction, the range of values makes it possible to discretize the lithofacies encountered.

320 Lissak et al. (2014b) identified a sharp contrast in resistivity with east-west growth explained by the litho-
321 stratigraphy (Fig. 7). This observation is compared with the values and the type of underground flows
322 highlighted by SP monitoring.



323

324 *Fig. 7 – ERT inversion profiles from the previous geophysical campaign conducted in the frame of the SISCA project (2008).*

325 *An east-west ascending gradient of resistivity values is observed.*

326 A cross-interpretation work focused on two representative ERT profiles in the eastern and western parts of the
 327 landslide with an upstream-downstream orientation (located in Fig. 7). More information can be found in Lissak
 328 et al. (2014b) concerning the acquisition parameters and inversion processes of these ERT profiles. Drillings in
 329 the vicinity of these profiles were analysed so that they can contribute to the interpretation of these geophysical
 330 data.

331

332 4. Results

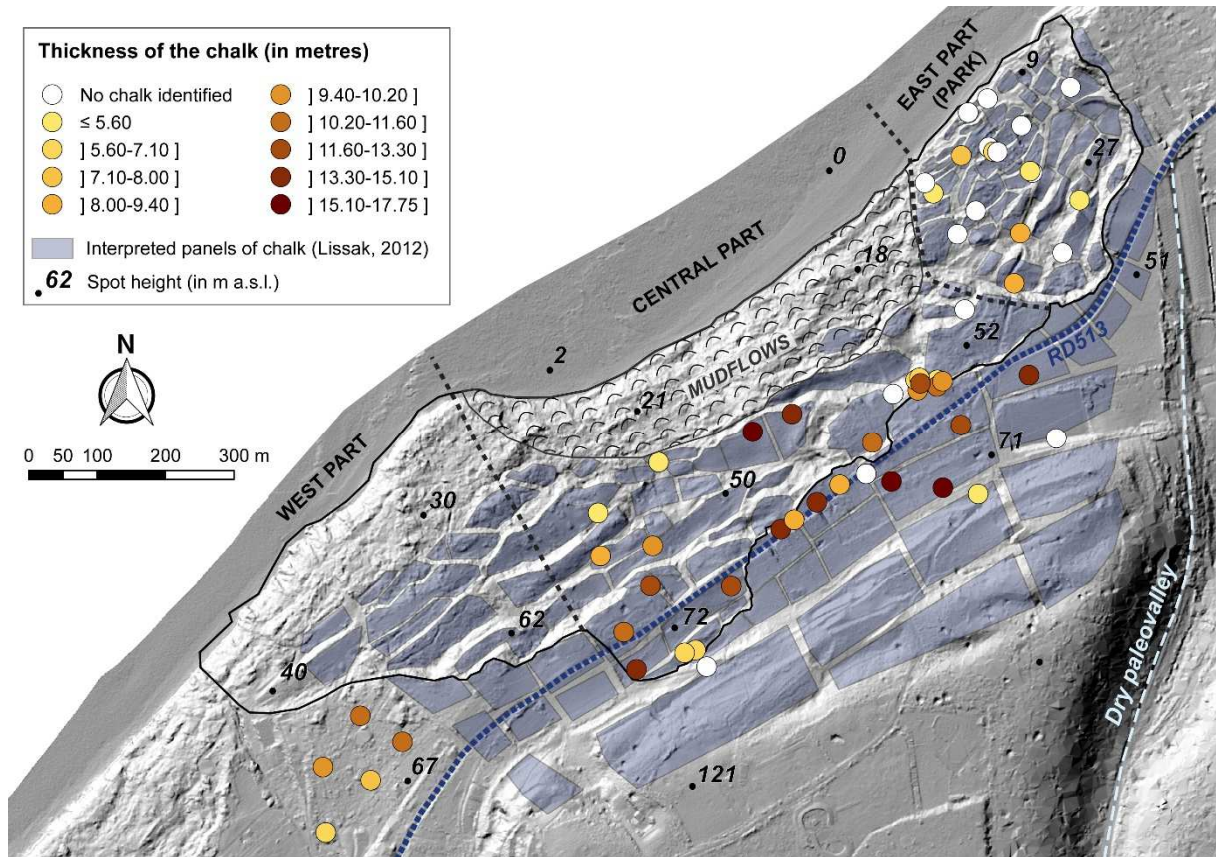
333 4.1. Structural analysis

334 The analysis of the 62 boreholes revealed a chalk thickness ranging from 2.80 m to 17.75 m (Fig. 8). The chalk
 335 layer is significantly thicker in the middle of the cirque, with a progressive decrease towards both extremities
 336 (northeast and southwest). The median zone includes the 14 drillings with the greatest thickness (>11 m).

337 The thinnest chalk layers are located in the northeast (park). In this zone, some drill logs do not clearly indicate
 338 the presence of chalk, even though the geomorphology suggests so. Out of a total of 20 drill holes, seven
 339 describe very significant weathering, most often referred to in descriptions as "yellowish-beige silty or sandy
 340 formations". This has been reinterpreted as "chalk", although the hydraulic conductivity of this altered material is
 341 closest to that of clay. The 13 other boreholes of the park did not identify any chalk, as the horizons were very
 342 heterometric with no clear distinction between the encountered facies.

343 The absence of obvious chalk is also possible elsewhere in the cirque. That is the case when the boreholes are
 344 located between compartments, in heterogeneous fill formations, or in the most reworked areas in the
 345 downstream part of the landslide.

346

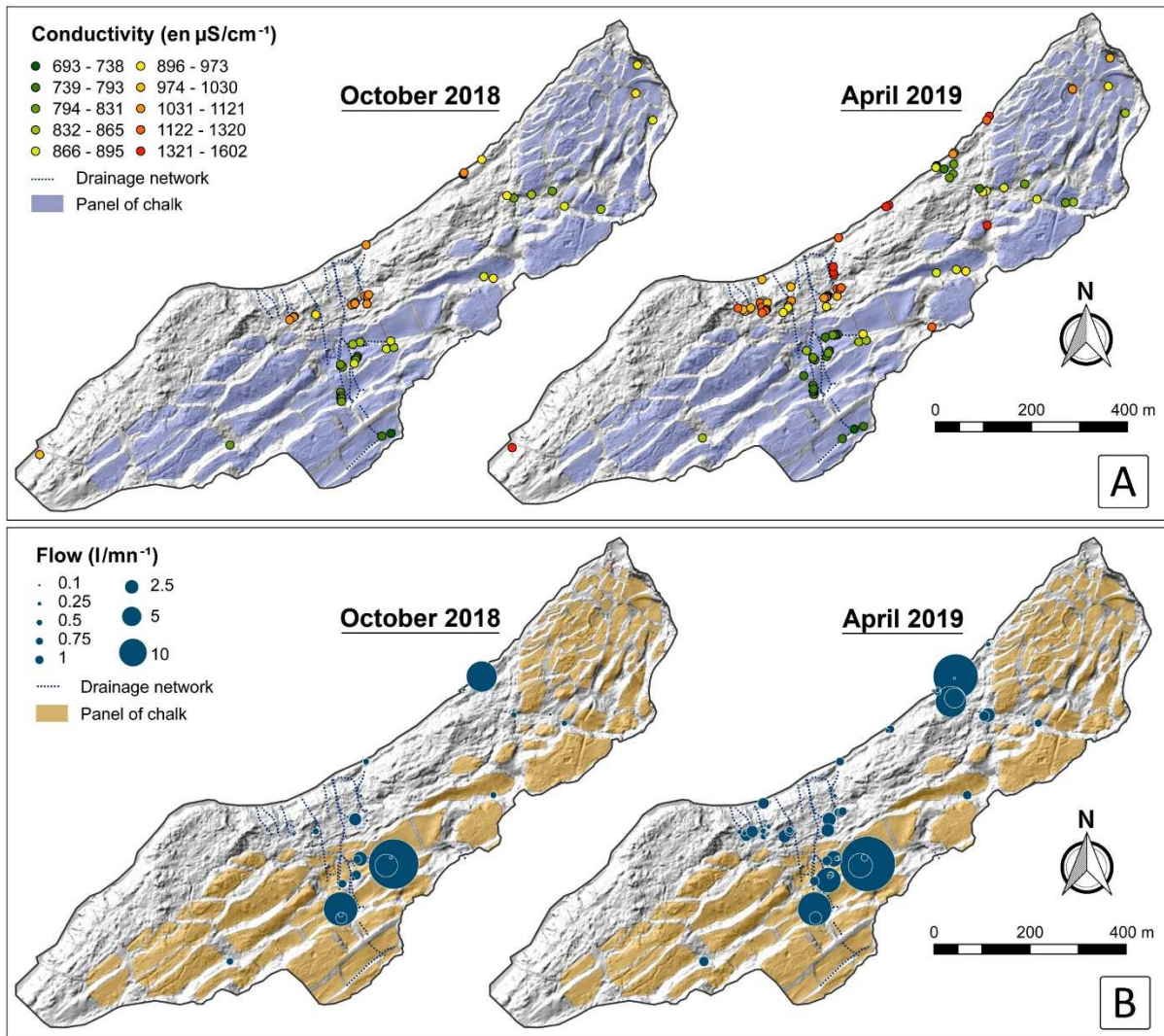


347
 348 Fig. 8 – Variation of the chalk thickness from borehole logs in or nearby the Cirque des Graves and geometry of the
 349 identified slabs by Lissak, 2012

350

351 **4.2. Characteristics of surface and underground waters**

352 The physicochemical measurements reveal a concentration of the measured surface water in three sectors (Fig.
 353 9). First, there is a significant flow at the southwestern border of the park along thalweg feeding ponds connected
 354 from the scarp. A large number of springs also outcrop at the foot of the chalk scarps in the middle of the cirque.
 355 Finally, a large number of springs, mainly in the form of seepage, can be observed at the point of contact
 356 between the Albian sands and clays, upstream of the mudflow area. Some springs outcrop directly on the
 357 foreshore, at the foot of the slope, in the heterogeneous reworked formations and heads, along the park's
 358 coastline. It is partly linked to the presence of flows at the park's boundary, located right on the upstream side.



359

360 Fig. 9 – (A) Conductivity measured in springs, ponds, and drains and (B) water flow measured in natural springs at the
 361 Cirque des Graves during the field campaigns of October 2018 and April 2019

362 The seasonality of the flows shows significant variations between October (low water) and April (high water). 39
 363 additional measurement points were recorded during the high-water period, including 25 springs, 10 stagnant
 364 areas, and 4 drains that had dried up by October 2018. The total flow measured at the springs increased from
 365 $115.0 \text{ l}\cdot\text{min}^{-1}$ to $171.80 \text{ l}\cdot\text{min}^{-1}$ (Table 1). The gap can be explained by an increase of $22.9 \text{ l}\cdot\text{min}^{-1}$ from the
 366 springs already measured in low waters and is related to new resurgences for a total of $33.9 \text{ l}\cdot\text{min}^{-1}$.

367 The spatial distribution highlights the importance of the central area on the measured flow (Fig. 9B), with 74%
 368 of the total flow in October ($85.1 \text{ l}\cdot\text{min}^{-1}$) and 69% in April ($119 \text{ l}\cdot\text{min}^{-1}$). In this central area, a network of
 369 artificial drainage is implemented to evacuate the excess of surface water through trenches, gullies and pipes
 370 (Fig. 10B, C, D).

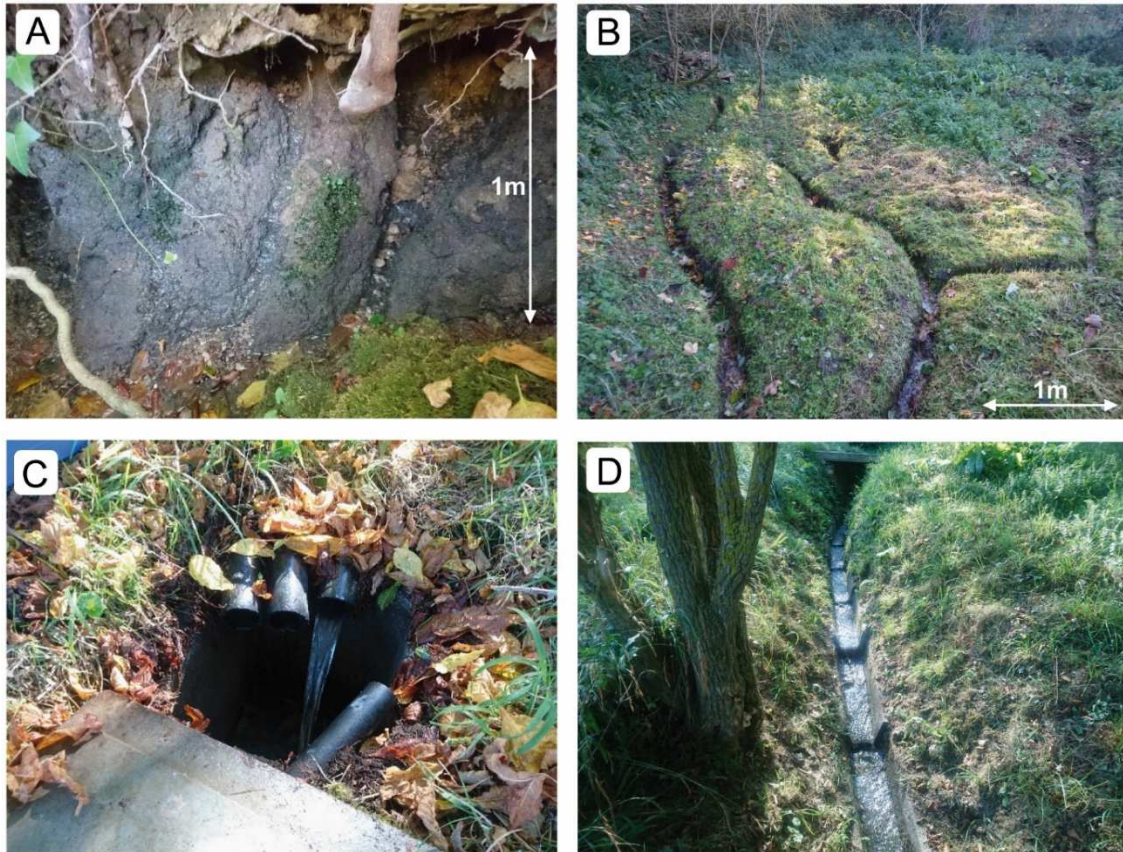
371 Electrical conductivity measurements increase from upstream to downstream and imply an ionic enrichment
 372 (Fig. 9A). In low waters, the difference between the extreme values is 400 $\mu\text{S}\cdot\text{cm}^{-1}$ and rises to more than 700
 373 $\mu\text{S}\cdot\text{cm}^{-1}$ in periods of high water. The nine springs with the highest values during the April 2019 survey are from
 374 1321 to 1602 $\mu\text{S}\cdot\text{cm}^{-1}$ and were all dry in October 2018. They are also all located at the outcropping of sands,
 375 after the last chalk slabs.

Measurements	October 2018	April 2019
Number of flowing springs	47	72
Number of ponds with stagnant water	8	18
Number of flowing drains	7	11
Temperature range (in °C)	12.0-20.3	9.3-17.7
Conductivity range (in $\mu\text{S}\cdot\text{cm}^{-1}$)	711-1103	693-1602
pH index range	6.68-8.55	7.19-8.58
Mean measured flow ($\text{l}\cdot\text{mn}^{-1}$)	2.52	2.59
Total measured flow ($\text{l}\cdot\text{mn}^{-1}$)	115	171.80

376 *Table 1 – Compared results between the two campaigns for surface water analysis*

377 Given the sedimentary context, waters logically tend to be basic. In both surveys, only one measuring point was
 378 below pH 7. The October 2018 campaign reports an average pH of 7.74 compared to 7.82 in April 2019. The
 379 variability is therefore low at the scale of the massif, with the lowest measurements systematically located in the
 380 ponds, often eutrophicated with the presence of duckweed.

381 The variability of the temperature is related to the type of measurement. The lowest temperatures measured are
 382 those of springs with an average of 14.9°C in October 2018 and 12.2°C in April 2019.



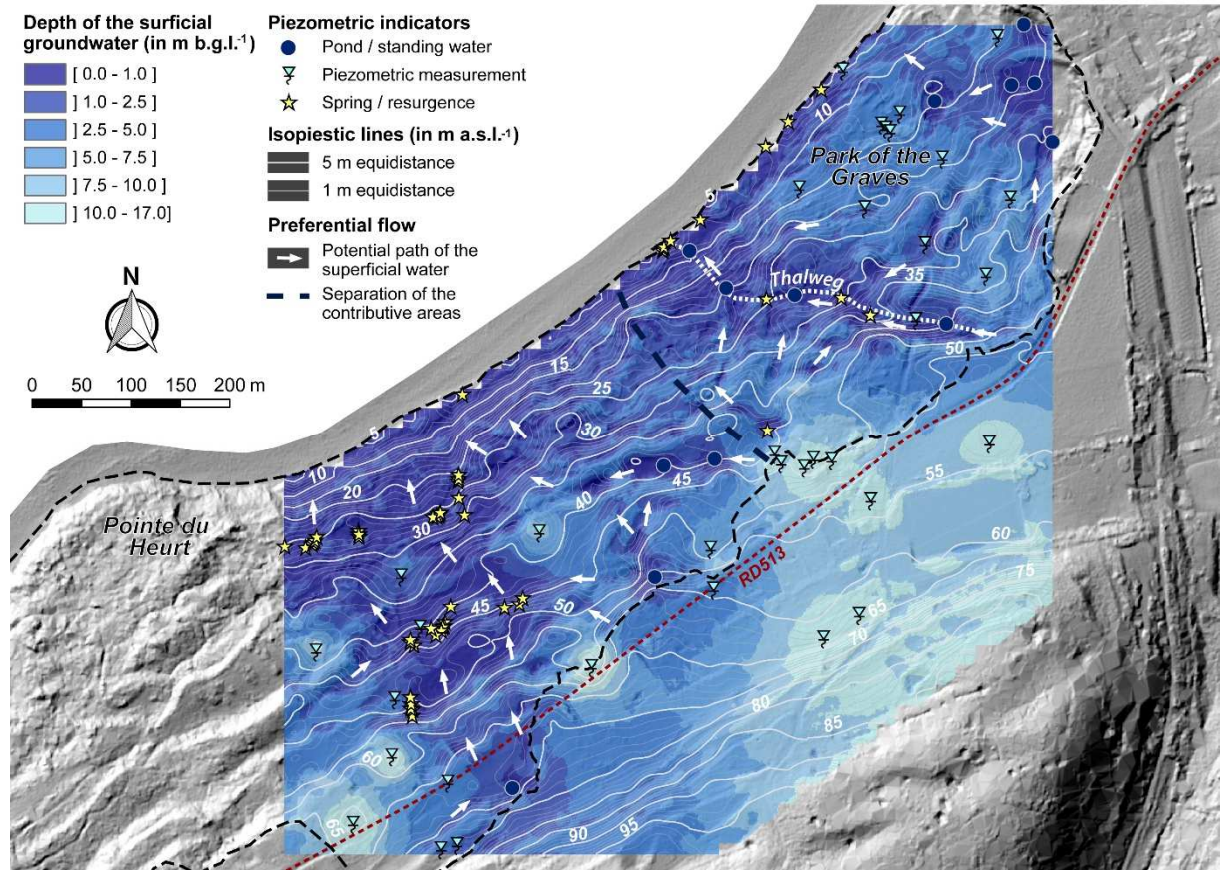
383

384 *Fig. 10 – A: Natural spring from the Cenomanian chalk at the basis of a five-metre-deep scarp; B: Trenches dug in the soil to*
 385 *channel and drain surface waters; C: Water arrivals from four drainage pipes in a concrete manhole for recovery in a larger*
 386 *pipe; D: Open trench built with overlapped precast concrete elements*

387 A piezometric map of the site has been produced by linking these measurement points with the piezometric data
 388 recorded (Fig. 11). The shallowest waters are located in the two areas of concentration characterized by
 389 numerous resurgences, with sub-surface levels of approximately 1 m in depth. In the unstable area, the water
 390 level seldom exceeds 5 to 7 m in depth. This water table depth increases in the upstream direction to reach up to
 391 17 m near the plateau.

392 There is a slight edge effect due to overly large gaps, generating a "cluster effect" when the values are too
 393 different over a short distance. However, the deepest values located along the RD513 and generating these
 394 clusters are indicative of a descending movement of the water in the filling materials in the cracks between chalk
 395 compartments, with values sometimes lower than 15 m b.g.l.⁻¹.

396



397

398 *Fig. 11 - Piezometric level of the unconfined groundwater considering the identified pond, springs and piezometric*
 399 *information from piezometers and drilling descriptions*

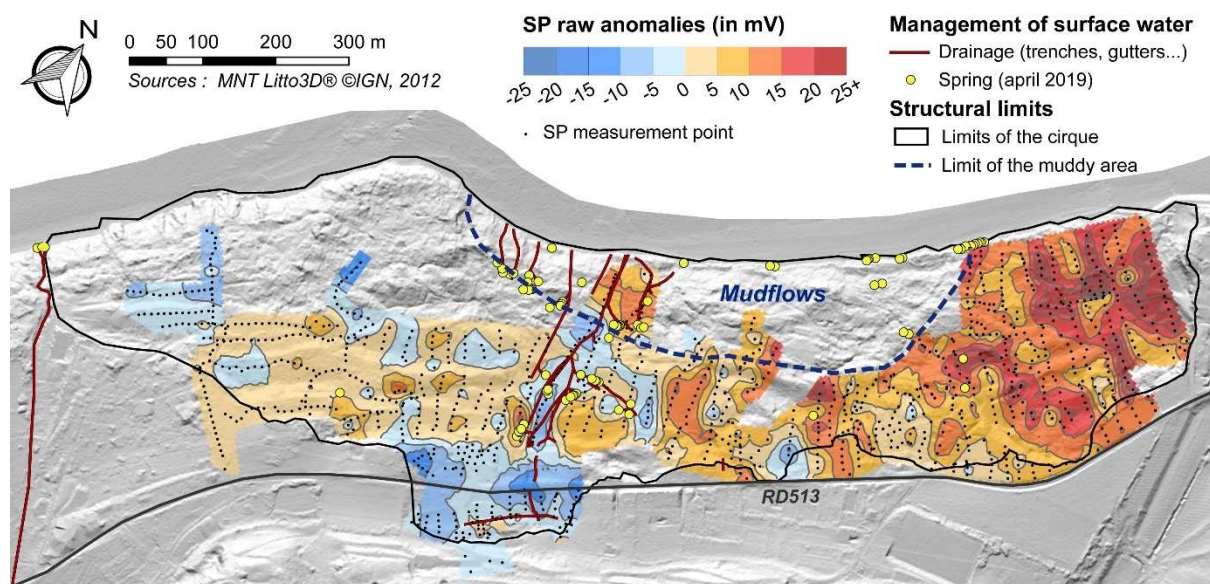
400

401 **4.3. Hydrokinetic assessment based on geophysics**

402 The interpolated SP map shows spatial heterogeneities (Fig. 12), with an increasing gradient of SP values from
 403 west to east. The range of values is quite limited, from -14.1 to +32.6 mV. The highest values are those found in
 404 the park, with values ranging from +3 to +32 mV, and no negative anomalies are found. The clusters having the
 405 highest values correspond to areas with springs and ponds (+20 to +32 mV). The maximum value of the entire
 406 survey was measured in the centre of a shallow pond located in this part of the cirque.

407 In the central section, the anomalies also increase from upstream to downstream, with marked negative values
 408 along the RD513 (-5 to -14 mV) and positive anomalies (+12 to +15 mV) at the contact of the clay-marl flows at
 409 the bottom of the slope. The upstream zone of the RD513 (in the centre of the landslide) corresponds to the
 410 strongest negative anomalies, near the main scarp.

411

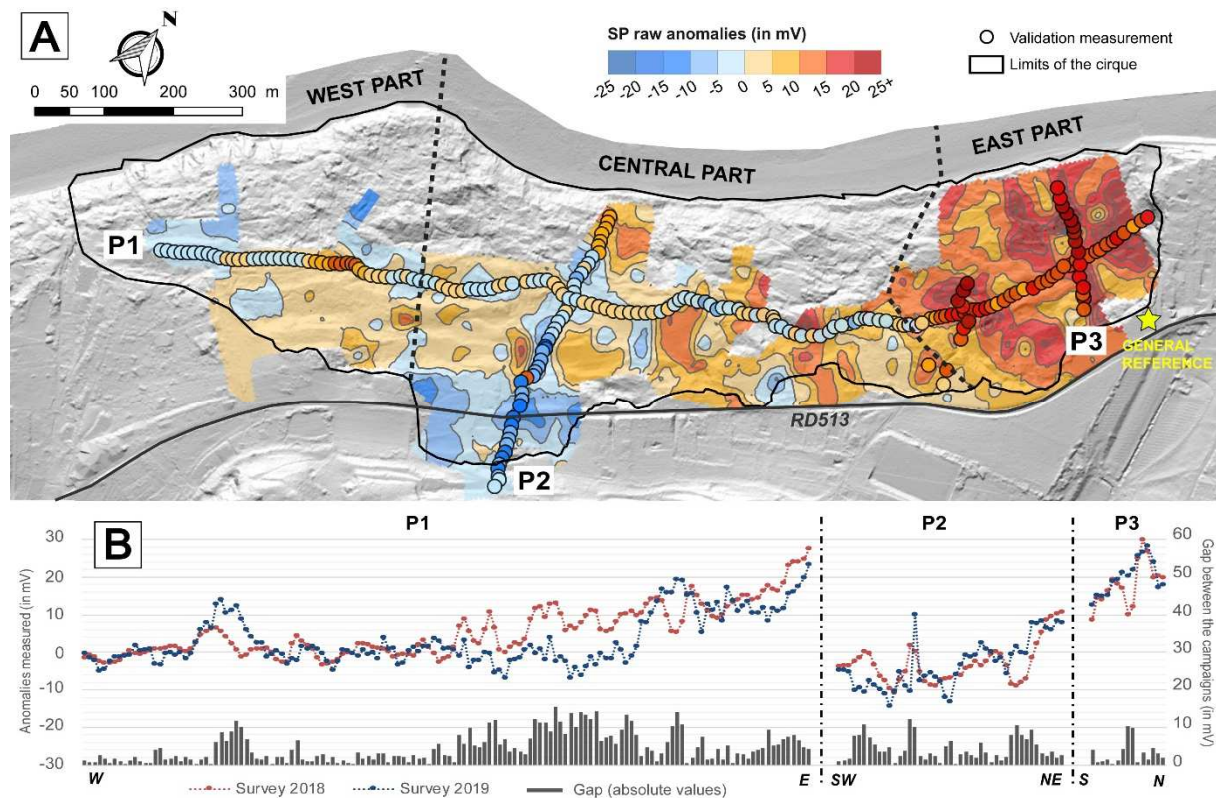


412

413 *Fig. 12 – Interpolated map of the self-potential over the Cirque des Graves and location of major drains in the middle part of*
414 *the landslide*

415 A validation survey was conducted at the end of April 2019, in exactly the same period as the previous year, in
416 an attempt to get as close as possible to the initial acquisition conditions. By relating the values from the second
417 survey to the general reference from the first survey, the range of anomalies seems to be comparable in broader
418 terms, ranging from -13.4 to +32.0 mV (Fig. 13A).

419 There are a few areas with deviations greater than 10 mV between the two surveys, which may be related to local
420 changes in subsurface flows. However, the curves are comparable (Fig. 13B). The overall survey error is
421 acceptable, with a median of 3.16 mV and a RMSE (Root Mean Square Error) of 2.14 mV.



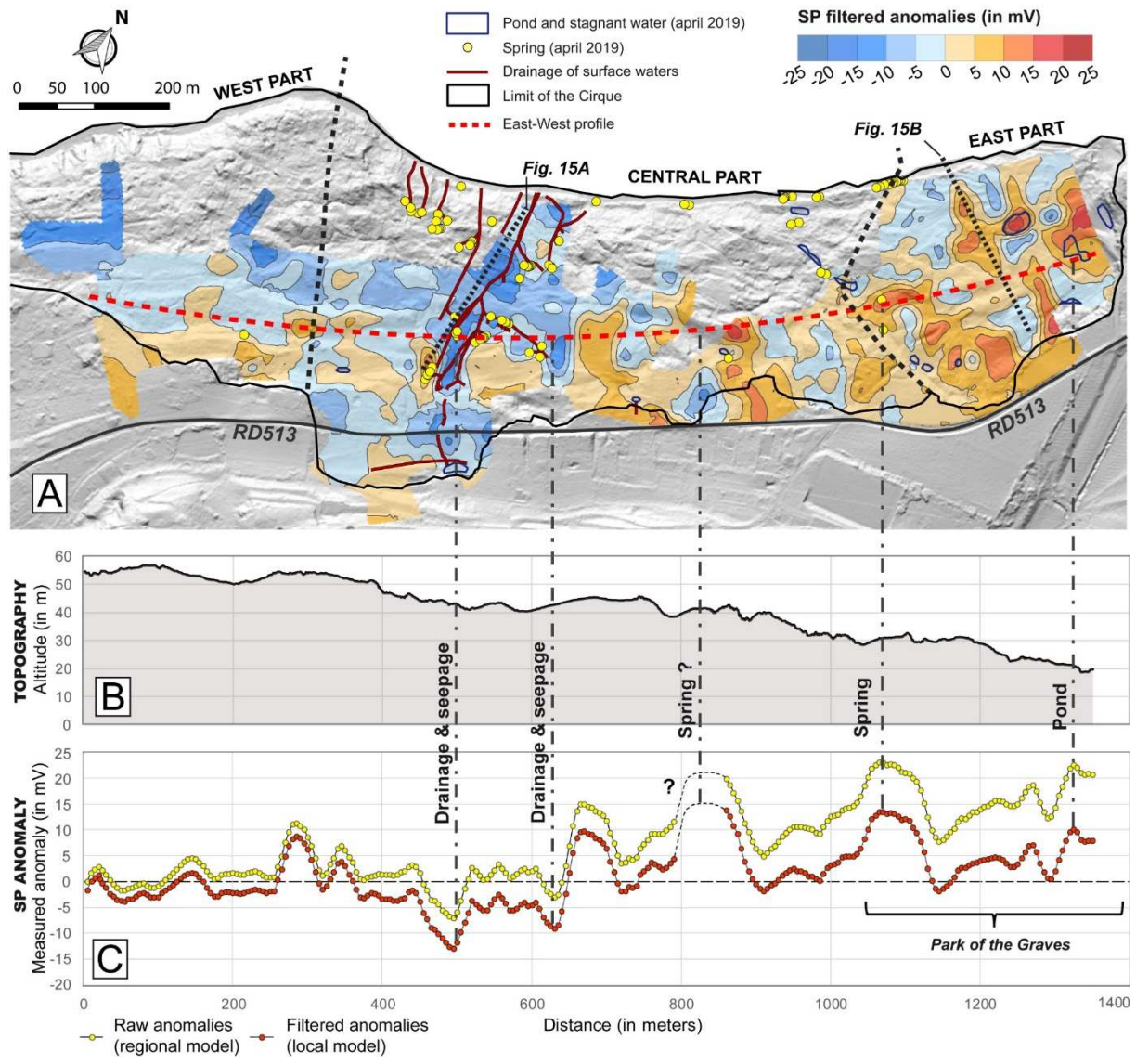
422

423 *Fig. 13 – Comparison between the SP anomalies from the initial survey (2018) and those from the validation survey (2019).*
 424 *(A) Spatial distribution of the points of the three validation profiles superimposed on the initial raw anomaly map (the colour*
 425 *range is identical); (B) “Point to point” comparison between the 2018 raw anomaly map and the 2019 verification*
 426 *measurements for the three profiles. The grey bar graph represents the absolute gap between the campaigns.*

427

428 Fig. 13 clearly shows the increase of anomalies from west to east and from upstream to downstream, thus
 429 highlighting a link with elevation following the same trends. The regional topographic effect was therefore
 430 filtered to more accurately reveal local anomalies (Fig. 14A) and was calculated to be -0.296 mV.m .

431 On the filtered map, the negative anomalies remain pronounced near the RD513 ($\approx -10 \text{ mV}$) and are heightened
 432 around the drainage systems (from -10 to -20 mV). The western side of the Pointe du Heurt is marked by a
 433 progressive upstream-downstream gradient of anomalies, with the lowest values (-15 to -25 mV) in the direction
 434 of the cliff. In the park, slightly negative zones appear with an upstream-downstream decrease. However, this
 435 eastern area remains more characterized by positive anomalies up to $+20 \text{ mV}$, particularly near the ponds and
 436 stagnant areas identified.



437

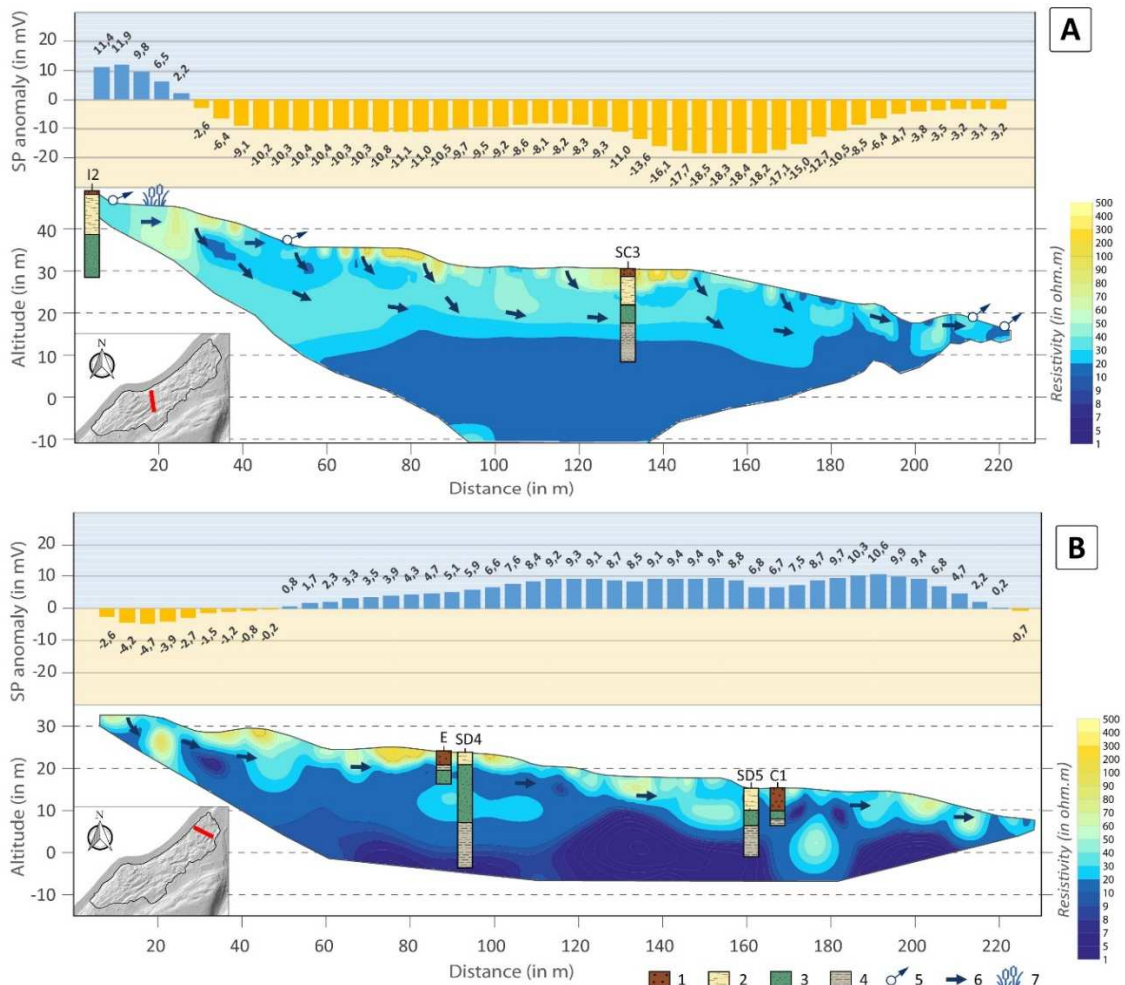
438 *Fig. 14 – Filtering of the topographic effect on SP anomalies for the characterization of local hydrokinetic dynamics*

439 To compare these surface acquisitions with in-depth information, two ERT profiles were selected and compared
 440 with nearby boreholes. These are respectively located in the park and in the central zone, above the mudflows
 441 (Fig. 15).

442

443 On the western profile (A), the SP anomalies decrease at the slope breaks. The upstream break shows a decrease
 444 of approximately 20 mV, and the anomaly reaches -18.5 mV in downstream direction. The flattened area in the
 445 middle of the profile also exhibits slightly less pronounced negative anomalies of up to -8 to -10 mV.
 446 Conversely, a positive anomaly located at the top of the profile reaches almost +12 mV in an area of stagnant
 447 water near springs. ERT and geology confirm these observations. The highest resistivities near the ground
 448 surface (50 to 500 Ω .m) are correlated with a fairly thick layer of chalk of up to ten metres. The contact with the

449 marls, constituting the lower limit of the aquifer formations, is identified at approximately 15 m in depth in
 450 borehole SC3 and coincides with resistivities below 30 Ω.m.



451
 452 *Fig. 15 – Correlation between SP filtered data (2018) and ERT profiles (2007-2008) in (A) the middle-western part of the*
 453 *Cirque and (B) in the eastern part of the Cirque (park). The location is specified in the thumbnail in the bottom left and*
 454 *corresponds to the two profiles indicated in Fig. 7 and Fig. 14. Legend: (1) Heterogeneous slope formation, (2) Chalk +/-*
 455 *weathered, (3) Glauconitic sand, (4) Clay & marl, (5) Spring, (6) Flow direction, (7) Stagnant water and hydrophilic*
 456 *vegetation.*

457 On the eastern profile (B), weak negative anomalies (down to -4.7 mV) are observed upstream of the profile, in a
 458 zone located below the main scarp. The anomalies become positive and reach +10.6 mV in the last third of the
 459 slope. Lithologically, this zone is characterized by the splitting of weathered chalk slabs (boreholes SD4 and
 460 SD5), with a thickness ranging from 2 to 5 m. These slabs' fragments are interspersed with heterometric clay-
 461 loam slope formations (boreholes E and C1). The ERT also indicates fairly conductive layers (< 20 Ω.m) close
 462 to the ground surface. Near-surface resistivities are generally lower compared to the western side and barely
 463 exceed 70 Ω.m on a few isolated clusters, with a thickness that does not exceed 5 m.

464

465 **5. Discussion**

466 **5.1. Towards a new hydrogeological model**

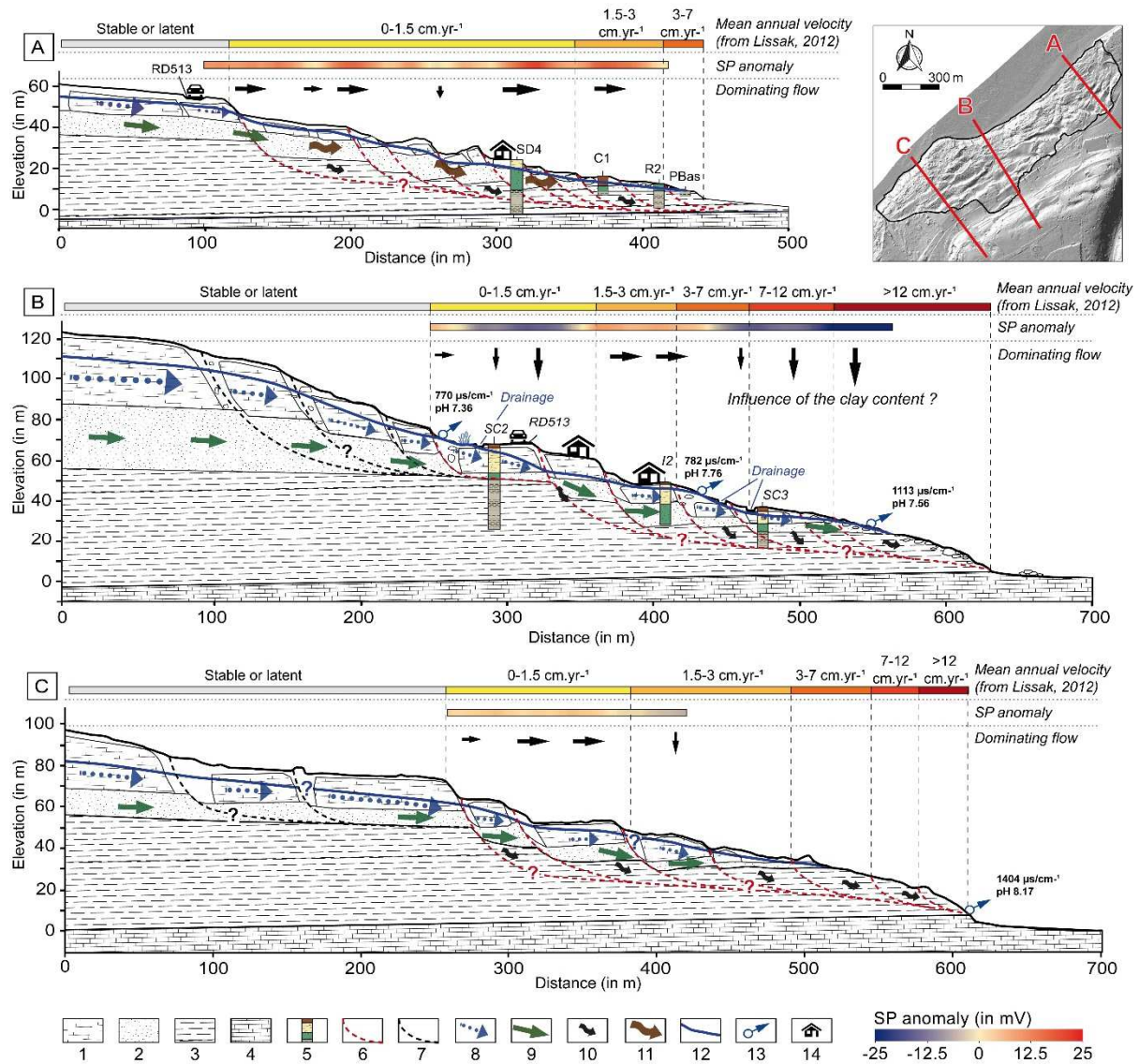
467

468 Thanks to the characterization of the SP anomalies and the physicochemical analyses, we are able to propose
469 hypotheses of spatial distribution, flow direction and intensity regarding the internal circulations, which could
470 not have been done with the initial geometric model.

471 We are in a position to confirm the hypotheses of preferential discharge from the plateau. Alternately, the initial
472 assumption of erratic flow and entanglement of the chalk and sand aquifers in the landslide is not confirmed
473 throughout the unstable zone. The variations in water conductivity between the chalk and sand outcrops may
474 suggest a partial dissociation of the aquifers in part of the slide.

475 The trend of the regional SP map is observed at a local scale, despite some infiltration zones that were not
476 noticeable before the topographic filtering. It tends to show that topography is not solely responsible for the
477 West-East increasing anomalies shown, with other electrokinetic sources being incriminated as the cause of these
478 anomalies.

479 In light of these results, the three approaches (structural, hydrological and geophysical) converge towards the
480 same trend, though with a division concerning the hydrogeological behaviour. Water flow dynamics are not
481 homogeneous at the landslide scale, and it is possible to delineate three different compartments in the eastern,
482 central and western parts of the landslide (Fig. 16). Such compartmentalization has already been performed for
483 other landslides, such as the Super-Sauze earthflow (Malet, 2003; Malet et al., 2005; Montety et al., 2007).
484 However, the use of SP provides essential information that allows the validation of the flow directions assumed
485 from field surveys and laboratory data.



486

487 *Fig. 16 – Cross-sections of the conceptual hydrogeological behaviour identified in each of the three parts of the cirque.*

488 *Legend: (1) +/- weathered chalk, (2) glauconitic sands, (3) marls/clays, (4) gritty limestone, (5) boreholes, (6) potential*
 489 *active sliding surface, (7) potential latent sliding surface, (8) chalk water table flow, (9) sand water table flow, (10)*
 490 *clays/marls wetting, (11) undifferentiated matrix flow, (12) free water table level, (12) spring, (13) building.*

491

492 **Eastern Part**

493 On the east side (park), the high permeability of the heterogeneous matrix, with thin and fragmented chalk slabs,
 494 facilitates the internal flow. These observations are validated by the geological descriptions and by the low
 495 resistivities up to near-surface. It explains the small number of resurgences in this zone, rather characterized by
 496 the presence of ponds, in relation to a near-surface water table level. In this zone, the groundwater seems to be
 497 undifferentiated in the sliding and heterogeneous matrix (Fig. 16 a). The surficial water supply from the Plateau

498 has been quantified with a continuous flow of water throughout the year, although the flow is quite limited. Even
499 if the regional SP map (raw) shows no negative anomalies, the filtering map suggests the presence of some
500 moderately infiltrated areas. Despite the eastward slight regional dip (1°), we detected no evidence of a direct
501 inflow coming from the west of the cirque into the park. The thalweg delineating the park acts as a natural drain
502 and allows the evacuation of a significant amount of water from springs and surface runoff. This area is also
503 characterized by positive anomalies, consistent with water accumulation zones. This draining function is
504 confirmed by the electrical conductivity values, which are steady from upstream to the foot of the slope,
505 reflecting a continuous surface and sub-surface flow with the same origin.

506

507 Central Part

508 This central zone appears to be the main contributor of water inflows, which is confirmed by (1) the
509 retrogressive saturated zone (pond and drains) behind the RD513; (2) the numerous springs found at mid-slope
510 with high flow drained by hydraulic structures; and (3) the flow dynamics affecting the saturated clayey and
511 marly materials at the bottom of the slope. This postulate was made by Lissak (2012), and the new acquisitions
512 have improved this empirical knowledge and validated the assumed behaviour.

513 Field observations and negative SP anomalies express a main discharge from the plateau water table upstream of
514 RD513, followed by a drawdown of the water table at the main scarp downstream of the road (Fig. 16 B). This
515 discharge is probably amplified by the greater thickness of the aquiferous layers in the upstream zone, compared
516 to the Eastern and Western areas. Such infiltration and SP anomalies in the detachment zone were noticed in
517 other places, such as the Varco d'Izzo or Giarossa landslides (Lapenna et al., 2003; Perrone et al., 2004;
518 Colangelo et al., 2006)

519 The flow appears less erratic in this part of the unstable zone, thanks to larger, thicker and generally less
520 weathered chalk slabs, as shown by borehole logs, and higher resistivity values than in the park. Average water
521 conductivities are measured at the top and middle of the slope ($700-900 \mu\text{S}\cdot\text{cm}^{-1}$) at the chalk scarps, while a
522 substantial increase occurs at the outcrop of the Albian sands down the slope. The values range from 900 to 1100
523 $\mu\text{S}\cdot\text{cm}^{-1}$ in low waters and can reach $1600 \mu\text{S}\cdot\text{cm}^{-1}$ in high waters on springs that were initially dry in October.
524 These high conductivities result in a decline in the SP anomaly, as the fluid conductivity decreases the
525 electrokinetic coupling effect (Revil et al., 1999; Naudet et al., 2008).

526 The increase of the conductivity in a short distance could imply an extended residence time of these waters at the
527 outcrop and a possible dissociation between both aquifers. This hypothesis needs to be confirmed by means of
528 geochemical methods (major ions, isotopic or artificial tracing). Marine pollution due to sea spray also cannot be
529 excluded in a coastal context (Moore and Brunnsden, 1996), and could influence these values, despite a dense tree
530 cover at the springs' location. Nevertheless, this influence seems limited, as suggested by the values measured at
531 the bottom of the thalweg delimiting the park, near the foreshore ($\pm 800 \mu\text{S}\cdot\text{cm}^{-1}$).

532 In this area, the sand water table moistens the clays, which behave like an aquitard (i.e., weakly aquiferous but
533 containing an important quantity of water). This moistening, which induces an increase in pore pressure, is a
534 long process. The flow is delayed and dampened compared to a normal aquifer, given the low hydraulic
535 conductivity of clays (Tavenas and Leroueil, 1981). This process, associated with the discharge of the overlying
536 aquifers, explains the mudflow processes identified at the bottom of the slope.

537 Alternately, negative SP anomalies along the drainage systems have to be explained. These values are consistent
538 with the results of Bogoslovsky and Ogilvy (1973), which show a decrease in the SP values and negative
539 anomalies reaching -8 to -12 mV along ditches and channels. In general, the drainage structures increase the flow
540 velocity of groundwater and thus the electric fields of filtration. For Bogoslovsky and Ogilvy (1973), these
541 negative anomalies might be regarded as explaining a concentrated infiltration of water into the rock and are
542 therefore related to the efficiency of drainage or leakage due to the poor structural condition of the
543 developments. On the local network, this could be explained by seepage from trenches dug directly into the
544 ground or permeabilities at the junction of concrete elements forming part of the drainage network (Fig. 10).

545

546 *Western Part*

547 The behaviour of the western part of the cirque (west of Pointe du Heurt) is not completely clear because the
548 available data set is more restricted in this area of the landslide due to the greater difficulty of access.

549 However, it is possible to define behavioural hypotheses from the few available data and field interpretations
550 (Fig. 16 C). The electrofiltration signal is quite heterogeneous from the raw results. However, there is a
551 decreasing gradient of SP anomalies in the direction of the foreshore, more clearly revealed by the filtering of the
552 topographic effect. For the upstream front, there are slightly positive anomalies (<10 mV), which may indicate a
553 weak surface discharge from the stable zone.

554 Thus, a decrease in anomalies towards the bottom of the slope could imply progressive water infiltration. Several
555 other observations mainly suggest deep-water functioning, such as: (1) the low number of resurgences and areas
556 of stagnation (springs and ponds); (2) the high scarps in this area (>15 m); and (3) permanent flows with high
557 conductivity (1003 Ω .m in autumn and 1404 Ω .m in spring) at the contact with the Villerville marls on the cliff
558 edge involving a rather long residence time. The significant decrease in the downstream SP values may also be
559 influenced by the change in substrate, with the clay-marl facies outcropping on the lower half of the slope of this
560 western zone. Indeed, the soil resistivity distribution controls the amplitude of the self-potential anomaly. Clay is
561 characterized by a very low resistivity (visible with ERT) and does not favour the occurrence of strong SP
562 anomalies (Skianis, 2012).

563

564 **5.2. Interpretation limits and errors in the SP survey**

565 Aubert (1997) points out an insufficient knowledge of the generation of SP potentials, which makes the
566 interpretation complicated and requires case-by-case argumentation. Over the years, the improvements of this
567 knowledge made it a powerful and proven method for solving hydrogeological problems (Naudet and Binley,
568 2006). However, many parameters may explain the SP signal, defined by Hämman et al. (1997) as an overlay
569 of elementary sources. It is therefore crucial to consider data processing and interpretations in the environmental
570 context.

571 For example, SP can help to determine the piezometric level by inverting the self-potential values (Darnet et al.,
572 2003; Naudet, 2004; Naudet and Binley, 2006), which is not appropriate in our case due to the topography and
573 very rugged geology of the area. A bivariate analysis between SP anomalies and measured piezometric levels
574 shows a total absence of correlation with $R^2=0.027$. The electrofiltration anomalies measured in our case seem to
575 be mainly influenced by water infiltrations and accumulations. We have therefore limited our approach to the
576 identification and mapping of the main upward or downward preferential flow paths, which brings to light the
577 discharge zones and the deep or surficial behaviour. This geophysical method requires a-priori information on
578 the distribution of cross-coupling and the origin of primary electric field sources (Hämman et al., 1997), which
579 is enabled by the hydromechanical information from in situ surveys.

580 A discussion regarding data quality is necessary to determine the robustness of the survey. Electrical drifts occur
581 between the two electrodes during the survey depending on many local factors, such as thermal variations, soil
582 drying, vegetation suction or local disturbance signals (e.g., power lines, pipes). It is therefore necessary to check

583 the drift regularly during the survey. We carried out 46 control measurements during the campaign by checking
584 the electrical potential difference between the two electrodes at the location of the reference electrode or by
585 monitoring the drift over time at the location of the measurement points.

586 The distribution of the control measures reveals discrepancies that need to be analysed. The range of absolute
587 deviations between initial and control measurements is from 0.0 to 17 mV. The RMSE of the distribution is 2.32
588 mV, and the standard deviation is 5.07 mV. The difference between the RMSE and standard deviation is large,
589 which tends to confirm the influence of the few extreme values, with the RMSE being less sensitive to these
590 values (Bouthevillain and Mathis, 1995). These extremes may be related to local factors that may generate
591 disturbances of the measurable electric field at the surface, as mentioned above. It is also important to note that
592 the drift is relative to the size of the study area (cumulative effect at each base change) and the duration of the
593 monitoring (time-varying effect of soil properties) (Corwin and Ward, 1990). We investigated 47 ha over 12
594 days in dry conditions, with more than a thousand measurement points, which is significant compared to
595 conventional landslide surveys.

596 Previous work has shown similar errors. For example, a standard deviation in the range of ± 10 -15 mV was
597 reported by Naudet (2004); ± 20 mV was reported by Fournier (1989); ± 10 mV was reported by Birch (1993); \pm
598 5 mV was reported by Corwin and Hoover (1979), Perry et al. (1996), Panthulu et al. (2001), and Revil et al.
599 (2004); and ± 3 mV was reported by Hämman et al. (1997). The main difference with the current study is the
600 range of values, which remains quite small, of the order of a few tens of mV, probably influenced by the clayey
601 substrate. Comparatively, the range of values was approximately 600 mV for Naudet (2004) or 900 mV for
602 Fournier (1989) and Revil et al. (2004) but in different environments or with different scientific issues (e.g.,
603 pollutant dispersion).

604 The maximum drift of the survey is 17 mV and corresponds to 48.7 % of the total range of measured values (0-
605 34.9 mV), while the root mean square error is only 6.6 % of this range. The errors in the estimation of
606 geophysical surveys carried out in a natural environment are generally of the order of 20 %, a value that is
607 considered by geophysicists as a tolerance threshold (Abdul Samad, 2017). Therefore, excluding the few extreme
608 ones, the highlighted errors seem to be acceptable.

609

610

611 **6. Conclusion**

612 The objective of this study was to assess the hydrogeologic behaviour of the Cirque des Graves using spatially
613 and temporally heterogeneous datasets to improve the limited initial knowledge. Structural and piezometric
614 datasets acquired over the long term were first used to assess the predisposition of the massif to underground
615 flows. This allows the highlighting of litho-stratigraphic disparities explaining contrasts regarding water table
616 levels and the distribution of runoffs on the surface. The wide heterogeneity of the geological formations
617 encountered (variation in thickness, reworking and discontinuities due to landslide dynamics, varying density
618 and hydraulic conductivity of the materials, etc.) therefore explains the uneven functioning of the slope. Second,
619 geophysical and physicochemical data acquisition aimed to explain this variability by characterizing the
620 hydrokinetics of the slope.

621 We have shown that the slope is divided into three compartments with an east-west division. The eastern part
622 (park) is characterized by a heterogeneous matrix with weathered and fractured chalk slabs favouring the mixing
623 of the two water tables of chalk and sand. These observations are verified by (1) the borehole logs, (2) a low
624 electrical conductivity of the matrix to the near surface, (3) the low number of resurgences despite a high water
625 table, (4) a homogeneous water conductivity from upstream to downstream, and (5) the quite positive SP
626 anomalies despite some small infiltration zones identified after the removal of the regional topographic effect.

627 Conversely, the central part is characterized by a less linear and more marked flow (upstream pond, numerous
628 resurgences and drainage networks). This is due to more pronounced scarps enhancing transmissivity along slip
629 surfaces and a thicker, less weathered chalk, which is confirmed by geological logs and ERT. Regarding the
630 hydrokinetic features, the numerous SP anomalies support the hypotheses of (1) a greater verticality of the flows
631 than in the east, (2) an inflow of the continental water table upstream and then an infiltration in the vicinity of the
632 RD513, and (3) an incomplete dissociation of the chalk and sand aquifers. The influence of the drainage
633 networks on the electrofiltration phenomena is also clearly noticeable.

634 Finally, the west of the Pointe du Heurt, while less characterized, is fairly close to the central zone from a
635 structural point of view but seems to experience infiltration with less intense water inflows than the central area,
636 as shown by the low number of resurgences and decreasing SP anomalies towards the shore.

637 In the end, the implementation of boreholes and piezometers are classical methods of investigations for
638 monitoring unstable slopes. Based on this available empirical knowledge concerning numerous landslides, the
639 results obtained show that hydrogeological knowledge can be substantially improved with light, easily

640 transportable investigation methods and with limited costs, even on large and complex slopes. In non-
641 homogeneous environments, ERT and SP are recognized as the two easiest geophysical methods to implement
642 (Meric et al., 2006) and are interesting to couple. The acquisition of physicochemical parameters is also simple
643 and inexpensive.

644 However, the observed conductivity measurements open up new research prospects. Further research would be
645 required to dissociate the contributions between the two aquifers of chalk and sand in a reliable manner and to
646 determine what comes under the impluvium. Knowledge of residence times would also help with interpretation.
647 To that end, the study of the strontium isotopic ratio $87\text{Sr}/86\text{Sr}$ could be used (Siegel et al., 2000; Legeay, 2013)
648 to determine the level of connection between the two regional aquifers at a local scale (Brenot et al., 2008).

649 Other geophysical methods should also be explored. The international literature shows recent developments in
650 terms of SP data processing, which may allow for more advanced acquisitions and data inversions for future
651 studies but is beyond the main focus of the present paper. For example, continuous SP monitoring could be
652 processed over a given period (from several hours to several months) to characterize the variability of
653 groundwater flows (Meric, 2006). The realization of self-potential tomograms (Patella, 1997) would also be
654 considered and has already proved its worth in landslide studies (Perrone et al., 2004; Colangelo et al., 2006).

655

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