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1 **Inferring the heterogeneity, transmissivity and hydraulic conductivity of** 2 **crystalline aquifers from a detailed water-table map**

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9 **Abstract**

10 Estimating the transmissivity or hydraulic conductivity field to characterize the heterogeneity
11 of a crystalline aquifer is particularly difficult because of the wide variations of the
12 parameters. We developed a new approach based on the analysis of a dense network of water-
13 table data. It is based on the concept that large-scale variations in hydraulic head may give
14 information on large-scale aquifer parameters. The method assumes that flux into the aquifer
15 is mainly sub-horizontal and that the water table is mostly controlled by topography, rather
16 than recharge. It is based on an empirical statistical relationship between field data on
17 transmissivity and the inverse slope values of a topography-reduced water-table map. This
18 relationship is used to compute a transmissivity map that must be validated with field
19 measurements. The proposed approach can provide a general pattern of transmissivity, or
20 hydraulic conductivity, but cannot correctly reproduce strong variations at very local scale
21 (less than 10 metres), and will face of some uncertainties where vertical flows cannot be
22 neglected.

23 The method was tested on a peridotite (ultramafic rock) aquifer of 3.5 km² in area located in
24 New Caledonia. The resulting map shows transmissivity variations over about 5 orders of
25 magnitude (average LogT: -5.2±0.7). Comparison with a map based on measured water-level
26 data (n=475) shows that the comparison between LogT-computed values and LogT data
27 deduced from 28 hydraulic tests is estimated with an error less than 20% in 71% of cases
28 (LogT±0.4), and with an error less than 10% (LogT±0.2 on average) in 39% of cases. From
29 this map a hydraulic-conductivity map has been computed showing values ranging over 8
30 orders of magnitude. The repeatability of the approach was tested on a second data set of
31 hydraulic-head measurements (n=543); the mean deviation between both LogT maps is about
32 11%. These encouraging results show that the method can give valuable parameter estimates,
33 and can characterize aquifer heterogeneity. The computed LogT and LogK maps highlight the
34 spatial distribution of parameters that show a pattern clearly controlled by the fault network of
35 this ultramafic massif. However, the faults are mainly characterized by low-permeability
36 zones; this differs from results on other crystalline aquifers and may be due to the fact that
37 weathering products of peridotite are clay-like materials.

38 The resulting transmissivity map can be used as a starting point for modelling or to direct
39 additional fieldwork.

40 Key words: regionalization of aquifer parameters, transmissivity, fractured aquifer, hard-rock
41 aquifer, crystalline aquifer

42 43 **1. Introduction**

44 Estimating the spatial variations of aquifer parameters is one of the most difficult tasks to
45 ensure a correct calibration of hydrogeological tools and models and, therefore, a correct
46 description of flow and transport into a groundwater system. Though it is relatively easy to
47 evaluate these properties at a local scale, for instance hydraulic conductivity and effective
48 porosity deduced from hydraulic tests, it is more difficult to assess their variability at aquifer-
49 system scale, where spatial variations may occur over several orders of magnitude. This is
50 particularly true in crystalline rocks where hydraulic conductivity can vary over 12 orders of
51 magnitude (Tsang et al., 1996; Hsieh, 1998).

52 Over the past decades, various methods, combining hydrodynamic parameters, geostatistics,
53 geological facies, inverse-modelling techniques, geophysical data, etc., have been proposed
54 for estimating hydraulic conductivity, transmissivity or storativity at the scale of groundwater
55 systems (e.g. Carrera et al., 2005; de Marsily et al., 2005). However, most methods were
56 designed for alluvial and sedimentary aquifers, and require thorough field investigations.

57 In crystalline aquifers, the regionalization of these hydrogeological properties makes the
58 problem much more complex because of their strong natural heterogeneity. Indeed, various
59 degrees in fracturing and connection between fracture networks induce strong variations of
60 properties at all scales (e.g., Paillet, 1998; Maréchal et al., 2004; Le Borgne et al., 2004,
61 2006). Moreover, where exposed to deep weathering, such rocks can develop several
62 stratiform layers parallel to the weathering surface, in which hydrogeological properties are
63 closely related to the degree of weathering (Taylor and Howard, 2000; Dewandel et al., 2006,
64 Lachassagne et al., 2011). This increases the difficulty for regionalizing hydrodynamic
65 parameters. Few works describe the spatial heterogeneity of aquifer parameters in crystalline
66 aquifers, and mainly focus on transmissivity or hydraulic-conductivity mapping based on data
67 from hydraulic tests (Razack and Lasm, 2006; Chandra et al., 2008), on classified
68 transmissivity (i.e. indexed) maps, or on potential aquifer-zone maps (Krásný, 1993, 2000;
69 Lachassagne et al., 2001; Darko and Krásný, 2007; Madrucci et al., 2008; Dhakate et al.,
70 2008; Courtois et al., 2010).

71 More recently other approaches have been proposed, based on the concepts of stratiform
72 layers and that large-scale variations in hydraulic head may characterize large-scale properties
73 (Dewandel et al., 2012, 2017-in press). They base the regionalization of hydraulic
74 conductivity on an empirical statistical relationship between the hydraulic-conductivity
75 distribution from small-scale tests and linear-discharge rates from numerous pumped wells.
76 For effective porosity, in the absence of recharge from rainfall, the method combines—at a
77 cell scale—water-table fluctuation and groundwater-budget techniques, and a coarse-graining
78 (i.e. aggregation) method. First developed for regionalizing the effective porosity for that part
79 of the aquifer where the water table fluctuates, it has been extended in 3-D to the entire
80 aquifer thickness while introducing the geometrical structure of the weathering profile
81 (Dewandel et al., 2017-in press). The methods for regionalization hydraulic conductivity and
82 effective porosity were tested on several unconfined granitic aquifers exposed to deep
83 weathering in southern India (50 to 1000 km²), showing very good estimates when compared
84 with existing data (Maréchal et al., 2004, 2006; Dewandel et al., 2006, 2010).

85 The present study describes a new method for estimating the spatial distribution of
86 transmissivity and hydraulic conductivity in naturally drained crystalline aquifers. Previous
87 methods required that aquifers were significantly influenced by pumping to cause large
88 depressions in the water table, and/or that numerous data were available on linear-discharge
89 rates of exploited wells (Dewandel et al., 2012, 2017-in press). The proposed approach has
90 also advantage over other methods that use inverse modelling techniques where prescribed

91 boundary flux and recharge can introduce large uncertainties on the generated hydrodynamic
92 parameters. The present method is still based on the concept that large-scale variations in
93 hydraulic head can provide information on large-scale hydrogeological properties. It uses
94 detailed water-table maps, hydraulic tests and data on aquifer thickness. Its application is
95 illustrated on a small peridotite aquifer in New Caledonia (Fig. 1a). The main objective of this
96 study is to show that, with a few assumptions on groundwater flux and basic hydrogeological
97 data, the method gives a reliable picture of the spatial aquifer heterogeneity as well as valid
98 estimates of transmissivity and hydraulic conductivity. Information on deep aquifer properties
99 of fault zones in this ultramafic environment is also given. Furthermore, we provide additional
100 information on the functioning of aquifers in these geological environments that are still
101 poorly understood (Boronima et al., 2003; Dewandel et al., 2004, 2005; Join et al., 2005,
102 Nikić et al., 2013).

103

104 **2. Site description and field data**

105 **2.1 Location, geological history and climate**

106 One third of the main island of New Caledonia, Grande Terre, is underlain by a giant
107 ophiolitic nappe (Paris, 1981), represented by a large massif in the south and several isolated
108 massifs (klippe) on the west coast and in the northern part of the island (Fig. 1a). This
109 Peridotite Nappe is highly fractured, resulting from its complex geodynamic history that can
110 be summarized by: (i) Seafloor spreading of the South Loyalty Basin (SLB) during the Late
111 Cretaceous, (ii) Paleocene-Early Eocene convergence and subsequent Eocene subduction, (iii)
112 Obduction of the SLB over the Norfolk Ridge during the Late Eocene, (iv) Post-obduction
113 unroofing, and finally (v) Arrival of the New Caledonia block in Vanuatu's active east-
114 dipping subduction zone (Dubois et al., 1974; Cluzel et al., 2001). Post-obduction extensional
115 tectonics and uplift related to isostatic re-equilibrium during the Oligocene and Miocene took
116 place along major listric normal faults bounding the obducted ultramafic klippe (Lagabrielle
117 and Chauvet, 2008), isolating the massifs at various altitudes (Leguéré, 1976; Sevin et al.,
118 2014). Since the latest Oligocene, the massifs are exposed to deep weathering (e.g. Routhier,
119 1953; Sevin et al., 2012) that have generated thick laterite profiles in which nickel and cobalt
120 ores are concentrated and mined (Trescases, 1975; Llorca, 1993).

121 The study area is located in the north of 'Grande Terre' over a small aquifer, 3.5 km² in area,
122 of the Tiébaghi Massif (Fig. 1a). The massif is prospected and mined by Société Le Nickel
123 (SLN) for nickel ore found in the saprolite layer, and in the past for chrome ore through a 3-
124 km-long tunnel within the bedrock (Moutte, 1982; Fig. 1b).

125 The area is characterized by an oceanic tropical climate with a hot and humid cyclonic season
126 (December to April) followed by a cooler and drier season. Mean annual rainfall is about
127 1500 mm.

128

129 **2.2 Geology of the study area**

130 The Tiébaghi Massif is a 20-km long and 8-km wide tectonic peridotite klippe. A lateritic
131 profile of 20- to 70-m thick envelopes the massif and forms a gently west-dipping plateau (8°)
132 at about 600 m above mean sea level (Robineau et al., 2007; Beauvais et al., 2007; Fig. 1b,
133 Fig. 3a). Figure 2 shows a typical weathering profile of ultramafic rocks in New Caledonia
134 (Trescases, 1975; Ouangrawa et al., 1996; Sevin, 2014). At Tiéghagi, from top to bottom, it

135 consists of a thin ferruginous nodular layer and a ferricrete cap (about 5-m thick), a thin
136 laterite (alloterite), yellow or fine saprolite (about 20-25 m), and a coarse saprolite layer (10-
137 15 m) that overlies highly weathered/fractured bedrock (Join et al., 2005; Sevin et al., 2012).
138 Below, the bedrock is fresh (unweathered) and, from a hydraulic point of view not fractured
139 except near faults where some fractures may conduct low water flux deeper (Join et al., 2005).
140 This weathering profile structure is similar to that observed over granite and schist (Wyns et
141 al., 1999; Dewandel et al., 2006; Lachassagne et al., 2011), though the fissured zone is
142 thinner.

143 More than 30 electrical tomography profiles, each 1-km long, and about 600 exploration drill
144 holes in the studied zone have revealed the structure of the weathering profile as well as a
145 dense fault system with a dominant NW-SE—and minor N-S—direction, creating troughs and
146 ridges (Fig. 3b; Robineau et al., 2007). The troughs correspond to graben or half-graben with
147 a thickening of the weathering profile layers. Fig. 3b shows the bedrock bottom and the fault
148 network (only available for the northern part). On surface, the main faulted structures are
149 outlined in the ferricrete morphology by elongate NW-SE aligned sinkholes.

150

151 **2.3 Hydrogeological setting**

152 Much hydrogeological work has been carried out on the massif for estimating hydrodynamic
153 parameters, investigating groundwater flow, designing observation wells for groundwater
154 monitoring, spring- and stream-discharge measurements, etc. (e.g. AEP, 1996; Guzik, 1996;
155 MICA, 1999; Golder Associates, 2014). On the plateau, the static water level is at a shallow
156 depth, typically less than 8 m and more or less following the topographic surface (Fig. 3c).
157 Many seeps and springs, with flowrates of up to 5 L/s, occur along the edges of the plateau,
158 feeding perennial waterfalls and streams on the bedrock. A conceptual model of the
159 hydrogeological functioning of the aquifer was proposed by Join et al. (2005), highlighting
160 that all weathered layers contribute to groundwater flow, but that each presents a particular
161 behaviour. Ferricrete and a nodular layer form a temporary (perched) aquifer, but laterite
162 (alloterite) and fine saprolite (isalterite) can be considered as a low-permeability layer (10^{-7}
163 m/s) with quite high storativity because of a high clay content. The underlying coarse
164 saprolite forms the main aquifer (about 10^{-6} m/s), feeding the springs and streams of the area.
165 Deeper down where fractures extend, some groundwater flow may be deviated into bedrock,
166 but measurements of deep flow in a 550-m long chrome-mining tunnel (Fig. 1b, not shown on
167 maps) at 300-400 m below ground level, show that flux is very low, less than 6.5 L/s in total.

168

169

170 **2.4 Water table and hydrodynamic data**

171 Based on the hydrogeological database of the mine, a detailed water-table map was drawn
172 from 475 exploration wells for May 2007, resulting in a very high density water-table map
173 (Fig. 3c). These wells are not equipped (no casing, no screen) and drilled down to the bedrock
174 (typically they enter 3 to 10 m into the unfractured bedrock). Consequently, the resulting
175 hydraulic head corresponds to an average value for the main coarse saprolite aquifer.

176 The map was established using standard geostatistical methods (variogram analysis and
177 kriging). The variogram (Fig. 3d) shows that the data are spatially well-structured and that
178 kriging results in a relevant map. A cubic model combined with a small anisotropy ratio was

179 necessary to fit the experimental variogram (sill: 70, length: 1200 m; anisotropy ratio: 1.4,
 180 dir.: N140°). The anisotropy direction, globally parallel to the main geological structures
 181 (Robineau et al., 2007), suggests that hydraulic-head data are at least partially controlled by
 182 the fault structure of the massif. However, the main groundwater-flow direction is S-SW, at
 183 right angles to the anisotropy direction. The mean hydraulic-head value of the map is 508.2 m
 184 above sea level. Additionally, another water-table database (Oct.-Nov.2006) is used to test the
 185 repeatability of the proposed method (§ 5); data were analysed with the same method.

186 Hydrodynamic data for the coarse saprolite layer (Fig. 3c) consisted of short-duration
 187 pumping tests (n=11, flowrate <1.5 m³/h; duration: few tens of minutes to few hours) and one
 188 slug test (mine database), which was completed with 16 additional slug tests (this study).
 189 Pumping tests were interpreted with the standard the Theis curve fitting method (method for
 190 confined aquifer), while slug tests were interpreted with the Bouwer and Rice method (1976;
 191 method for unconfined aquifer). The literature shows that the application of analytical
 192 solutions for a confined aquifer to unconfined aquifer is possible without introducing any
 193 inaccuracy, particularly on transmissivity, when drawdowns do not exceed 25% of the aquifer
 194 thickness, which is the case in our study. All these small-scale hydraulic tests provided
 195 transmissivity data for the main aquifer near the wells. The distribution of logarithm of
 196 transmissivity, LogT (decimal log; logarithm to base 10) has an average of -4.98 and a
 197 standard deviation of 1.11 (Fig. 4).

198

199

200 **3. Method for regionalizing transmissivity**

201 The proposed method for evaluating the transmissivity field is based on the concept that
 202 large-scale variations in hydraulic head may give information on large-scale properties.

203 Where the aquifer is naturally drained (no groundwater abstraction), where vertical flow can
 204 be neglected, and where the water table is in pseudo-steady state and mainly controlled by
 205 topography rather than recharge (Haitjema and Mitchell-Bruker, 2005), it can be assumed that
 206 the gradient of the water-table map depends on both topographic slope and aquifer horizontal
 207 transmissivity (Fig. 5a, c). Otherwise, recharge will control the water table (Fig. 5b), leading
 208 to a gradient depending on aquifer horizontal transmissivity and not topography.

209 For a water table controlled by topography, as is expected for medium- to low-permeable
 210 aquifers, a flat topography, and where groundwater flow is horizontal, the continuity equation
 211 (Darcy's law) for unit aquifer width is:

$$212 \quad Q = T_1 \text{grad}h_1 = T_2 \text{grad}h_2 = \dots = T_n \text{grad}h_n \quad (1)$$

213 Where Q : is the horizontal flux (m³/s/m), T_i = aquifer horizontal transmissivity (m²/s), $\text{grad}h_i$:
 214 hydraulic gradient (dimensionless). Because of the flat topography, the use of Eq.1 along the
 215 same flow line will give high transmissivity values where the hydraulic gradient is low and
 216 low values where the gradient is high. However, as shown on Figure 5c, the hydraulic
 217 gradient is also strongly influenced by topography. Therefore, where the topographic level is
 218 almost stable (left part of Fig. 5c) the variation in hydraulic gradient depends more on
 219 variations in aquifer transmissivity, in fact inversely related to the hydraulic gradient, whereas
 220 in the right part of Fig. 5c it is more controlled by topographic slope because of stronger
 221 variation in elevation.

222 To obtain information on transmissivity where the water table is related to topography, the
 223 problem is thus to eliminate the influence of elevation on water-table data. To verify if the
 224 water table is expected to be largely controlled by topography, the criterion proposed by
 225 Haitjema and Mitchell-Bruker (2005)—based on a Dupuit-Forchheimer model solving the
 226 problems shown on Figures 5a and b—can be used as a bench mark:

$$227 \quad C_{HMB} = \frac{RL^2}{8Td} \quad (2)$$

228 where R (m/s) is the average annual recharge rate, L (m) the average distance between surface
 229 waters, T (m²/s) the horizontal aquifer transmissivity, and d (m) the maximum distance
 230 between the average surface-water levels and terrain elevation. They stated that if $C_{HMB} > 1$,
 231 the water table is largely controlled by topography (Fig. 5a); otherwise it is controlled by
 232 recharge (Fig. 5b).

233 However, most parameters of Eq. 2 could not be defined precisely and we thus assumed the
 234 following ranges for our study area: $100 < R < 400$ mm/y, $500 < L < 1000$ m (corresponding to the
 235 minimum spacing between two streams up to the plateau width; Fig. 3a), $T = 10^{-5}$ m²/s
 236 (average of hydraulic tests data, Fig. 4) and $5 < d < 10$ m. We found that C_{HMB} ranges between 1
 237 and 40, and concluded that the water table is essentially controlled by topography.

238 Figure 6 shows the relationship between hydraulic-head data and elevation for each of the 475
 239 wells, and thus the effect of elevation on well-head data. This very good relationship
 240 ($R^2 = 0.99$) is also consistent with the evaluated Haitjema and Mitchell-Bruker's criterion
 241 (C_{HMB}). Therefore, subtracting this trend from the original water-level data will remove the
 242 overall effect of topography and results in flattening the hydraulic-head data, thus only
 243 leaving those variations that are controlled by transmissivity. A reduced water-table map now
 244 can be drawn (Fig. 7) from which a slope map is computed. It is expected that the low values
 245 represent the highest transmissivity zones, whereas the high values show the low-
 246 transmissivity zones. Finally, the statistical distribution of the slope data—in practice the Log
 247 of the inverse of the slope—is compared to that of the transmissivity data. Once a relationship
 248 between the two distributions is found, a transmissivity map can be computed from inverse-
 249 slope data that has to be validated by the local estimates, the ones that served to establish the
 250 statistical distribution of LogT (Fig. 4).

251

252 **4- Results**

253 The studied aquifer, with only minor vertical groundwater flow compared to horizontal flow,
 254 conforms to the initially established hypotheses and allows testing the proposed methodology.

255 The variogram of reduced hydraulic-head data (Fig. 7a) shows that the data are spatially well-
 256 structured and that kriging results in a relevant map (Fig. 7b). An exponential model
 257 combined with a small anisotropy ratio was also necessary to fit the experimental variogram
 258 (sill: 5.1, length: 190 m; anisotropy ratio: 1.8, dir.: N140°). The anisotropy direction is still
 259 globally parallel to the main geological structures, again suggesting a relationship of this
 260 parameter with the geological structure of the massif. The reduced-water-table map has been
 261 computed on a grid of 20x20 m cells (Fig. 7b), with values ranging from -3.3 to 5.3 m with an
 262 average of 0.57 m and a standard deviation of 1.67 m. From this map, a slope map was
 263 derived (Fig. 7c), showing an alignment of structures approximately in the main NW-SE
 264 structural direction of the massif. Values vary over three orders of magnitude between 0.01

265 and 3 (m/m), with an average and standard deviation of 0.34 ± 0.38 (m/m). The inverse of the
266 slope was computed and shows a near log-normal distribution with an average logarithm of
267 $1/\text{slope}$, $\text{Log}(1/\text{slope})$, of -0.33 and a standard deviation of 0.31 (Fig. 7d).

268 Applying the method described above, the resulting empirical relationship between both
269 distributions of $1/\text{slope}$ and transmissivity shows a successful match between modelled and
270 observed distributions, with a linear regression coefficient (square form), R^2 , of 0.91 (Fig. 8).
271 The relation is:

$$272 \quad T = 1.1 \times 10^{-5} (1/\text{slope}) - 1.6 \times 10^{-5} \quad (3)$$

273 Eq. 3 indicates a linear relationship that partially verifies Darcy's Law in its left—
274 proportionality—part. However, there is a negative constant that suggests that the slope map
275 does not exactly reflect the aquifer transmissivity everywhere. The negative value
276 corresponds to the highest slope values (>0.7) and probably was introduced by a bias because
277 of the average relationship used for removing the impact of topography on hydraulic-head
278 data. Even if the use of this relationship is appropriate at the aquifer scale, this is probably not
279 true for some small perimeters of the area where the trend defined on Fig. 6 may locally
280 differ. These high-slope areas, correspond to the darkest zones of Fig. 7c, and cover about
281 30% of the data. Nonetheless, such zones may correspond to low-transmissivity values, but
282 with an exaggerated slope. Further explanations for this negative constant are also given in the
283 discussion section.

284 The computed-transmissivity map (on a logarithmic scale and a grid with 20x20 m cells;
285 Fig. 9a) is based on a geostatistical approach that uses Eq. 3 for slope values less than 0.7
286 ($n=5851$). The variogram of the computed $\text{Log}T$ (Fig. 9b) shows that the data are spatially
287 structured. An exponential model with an anisotropy ratio has been used to fit the
288 experimental variogram (sill: 0.24, length: 115 m; anisotropy ratio: 1.4, dir.: N140°). The
289 average and standard deviation of the computed- $\text{Log}T$ are -5.2 ± 0.7 .

290 In order to verify if the produced map conforms to field data, local $\text{Log}T$ -computed values
291 from the map were compared to $\text{Log}T$ data deduced from the 28 hydraulic tests (Fig. 9c).
292 Note that no local transmissivity values from hydraulic tests were used for drawing the map.
293 The comparison shows that in 71% of cases, the computed $\text{Log}T$ is estimated with an error
294 less than 20%, which corresponds to $\text{Log}T \pm 0.4$ on average. And in 39% of cases, the
295 computed $\text{Log}T$ is estimated with an error less than 10%, here corresponding to $\text{Log}T \pm 0.2$ on
296 average. Therefore, the method produces reasonable estimates of transmissivity considering
297 the strong heterogeneity of this fractured system with transmissivity variations of almost 5
298 orders of magnitude. In some cases, however—29% of the dataset corresponding to 60% of
299 the dataset with $\text{Log}T > -4.5$ —, high-transmissivity areas are not correctly estimated by the
300 map. This is due to the intrinsic characteristic of fractured systems, where high- and low-
301 permeability zones coexist at a local scale. Therefore, these values are probably very local and
302 not representative of average aquifer properties at the scale of some tens of metres. This point
303 was already mentioned by Dewandel et al. (2012).

304 The hydraulic-conductivity map, on a logarithmic scale ($\text{Log}K$; grid with 20x20 m cells)
305 (Fig. 9d), was based on the transmissivity map and on the thickness of the saturated aquifer
306 deduced from Fig. 3b, c. The $\text{Log}K$ variogram (Fig. 9e) is similar to that of $\text{Log}T$ and shows a
307 strong spatial dependency. The average and standard deviation of the computed $\text{Log}K$ are -
308 6.7 ± 0.8 , and values range over 8 orders of magnitude. Hydraulic-conductivity values are
309 consistent with the ones mentioned by Join et al. (2005), but also with those of the fissured
310 zone of Oman peridotite (Dewandel et al., 2004, 2005).

311

312 **5. Repeatability and validation of the method**

313 To verify if the method is repeatable and to validate it, we tested it on a second set of
314 hydraulic-head data. This new data set provides more water-table data ($n=543$), but they are
315 less precise because they were measured over a period of two months, between October and
316 November 2006, which may have introduced discrepancies because of potential small
317 recharge events during the measuring period. The mean hydraulic-head value for this map is
318 507.0 metres above sea level, a value slightly lower than the one established in May 2007
319 (508.2 m.a.s.l.). A new relationship of elevation *vs.* hydraulic head ($y = 0.9837x + 1.5371$; R^2
320 $= 0.99$) and a new Eq. 3 [$T = 1.4 \times 10^{-5} (1/\text{slope}) - 0.9 \times 10^{-5}$; $R^2: 0.88$], etc., were established from
321 this data set to compute a new LogT map (Fig. 10a). This map shows a similar pattern to the
322 one established with May 2007 data (Fig. 9a). The data are still spatially structured (Fig. 10b),
323 with a variogram model close to the previous one (model: exponential, sill: 0.32, length:
324 85 m; anisotropy ratio: 1.4, dir.: N140°). The average and standard deviation of the computed
325 LogT for 2006 is close to the earlier one, -5.2 ± 0.7 .

326 The mean deviation between both LogT-2006 and LogT-2007 maps has been computed and
327 Figure 11a presents the results. The average mean deviation is about 11% with a standard
328 deviation of 9% (Fig. 11b), and without dependency according to LogT ranges (random
329 distribution of the deviation; Fig. 11c). Note that the highest deviation is obtained in the
330 south-south-eastern part of the area where information on water level data is available for the
331 Oct.-Nov. 2006 data set only (Figs. 3c and 10a). These results confirm the May 2007 map,
332 and thus the proposed approach.

333

334 **6- Discussion**

335 We present an innovative method for estimating the spatial distribution of aquifer
336 transmissivity and hydraulic conductivity in crystalline aquifers. The approach builds upon
337 the author's earlier work (Dewandel et al., 2012) and considers here the case where there is no
338 pumped well in the area, but sufficient well-head data exist to produce a detailed water-table
339 map. The method's assumptions only consider that flux into the aquifer is mainly sub-
340 horizontal and that the water-table is mostly controlled by topography. Where the water table
341 is recharge controlled (not the case here), the water-table map gradient can directly be used
342 for estimating the transmissivity field. Our approach has a significant advantage over other
343 methods, for instance those that use inverse modelling techniques where boundary flux and
344 recharge can introduce large uncertainties on the generated hydrodynamic parameters. In
345 heterogeneous media, like in the studied case, such hydrological conditions are often
346 unknown and particularly difficult to assess.

347 The relationship between the distributions of inverse-slope ($1/\text{slope}$) and field-transmissivity
348 data does not fully satisfy Darcy's Law because of the constant in Eq. 3 (if it fully satisfies
349 this law, the constant should have been nil). The negative constant is believed to be
350 introduced by the aquifer-scale relationship used for removing the effect of topography on
351 hydraulic-head data, which may differ on a local scale. Another reason that may also explain
352 this bias is our basic assumption that vertical groundwater flux can be neglected compared to
353 the horizontal one. Though this hypothesis, at aquifer scale, is plausible because of minor
354 deep-groundwater flux as confirmed by measurements within the chrome-mining tunnel
355 (<7 l/s), it is possible that locally—within, or near, vertical faults—vertical flux cannot be

356 neglected. Consequently, some high slope values on Fig. 7c may reflect such flux that the
357 proposed method erroneously assimilates to low horizontal-transmissivity zones. In these
358 cases, there is probably a vertical transmissivity component that the method cannot evaluate.
359 The method also assumes that the water table is topography-controlled, which is the case at
360 aquifer scale, but locally it may be recharge controlled, thus introducing an additional bias.

361 To test the validity of our approach, we first verified that the produced LogT map conforms to
362 field data, which gave encouraging results (Fig. 9c). Then, to verify if the method is
363 repeatable, we tested it on a second set of hydraulic-head data (Fig.10a) that confirms the
364 previously established LogT map (Figs.11) and thus the proposed approach.

365 The computed LogT and LogK maps (Fig. 9a, d; Fig. 10a) show a strong spatial dependency
366 with zones of equivalent properties elongated in a roughly N140° direction. This direction
367 corresponds to the main fault directions of the Tiébaghi Massif (Robineau et al., 2007), which
368 shows that fault presence strongly influences the hydrodynamic properties, and thus
369 groundwater flux in the rock. Low-permeability zones (LogK<-7) are quasi-systematically
370 associated with the main fractures. This is explained by the fact that the weathering of
371 peridotite produces mostly low-permeable (clay-bearing, i.e. alloterite) materials that mostly
372 act as barrier to groundwater flow. This behaviour is very different to what can be observed in
373 granite and gneiss aquifers, where deepening of the weathering front along faults, contact
374 zones, veins, etc., generally enhances hydrodynamic aquifer properties (Dewandel et al.,
375 2011; Lachassagne et al., 2011). However, deeper in the bedrock some of these faults are still
376 slightly productive as shown by the flux measurements in the chrome-mining tunnel at 300 to
377 400 m depth.

378 An original feature of the peridotite aquifer in New Caledonia is the presence of subsurface
379 pseudo-karst structures (e.g., Genna et al., 2005) that are unknown from other studied
380 peridotite aquifers, probably because the weathering profile, there, is less developed (Cyprus:
381 Boronima et al., 2003; Oman: Dewandel et al., 2005; Serbia: Nikić et al., 2013). In the
382 Tiébaghi Massif, sinkholes affect the morphology of the ferricrete surface. Aligned and
383 elongated along the main NW-SE fault system, they mainly develop above horst border faults
384 (Fig. 1b), which may reflect underground flow paths that can evacuate the finest weathering
385 products (Genna et al., 2005). However, such sinkholes are now considered as poorly active
386 to inactive because they are commonly filled with water and characterized by low-water level
387 decline. One reason that may explain why the Tiébaghi sinkholes are almost inactive is the
388 almost right angle between the main fault directions (NW-SE and minor N-S) and the SSW-
389 directed groundwater flux, which does not favour the underground evacuation of (fine-
390 grained) weathering products and thus the functioning of these structures. This is not the case
391 in southern New Caledonia, where a large number of sinkholes affects the ferricrete surface
392 (around 60 sinkholes/km²), and where some of the largest are active (Jeanpert et al., 2016).
393 They are elongated and distributed over 50-m-wide zones oriented N140° inherited from
394 tectonic fracturing. However, the role of groundwater-flux directions and main fault directions
395 to explain the development of sinkholes and/or their functioning needs to be confirmed by
396 further works.

397

398

399

400 **7. Conclusions**

401 We show that basic analyses of spatially detailed information on hydraulic-head data can be
402 used for characterizing the heterogeneity of an aquifer and evaluating its transmissivity or
403 hydraulic conductivity fields. This is very useful in fractured media where *in-situ*
404 measurements of hydrodynamic parameters are generally not sufficiently dense to allow a
405 relevant mapping. The method requires detailed information on the spatial variations of
406 hydraulic head in the aquifer and on the statistical distribution of aquifer transmissivity, and
407 the establishment of an empirical statistical relationship between field data on transmissivity
408 and the inverse slope values of a topography-reduced water-table map. Here, hydrodynamic
409 parameters on 28 locations and a density of up to 155 water level observations per km² were
410 used to evaluate the transmissivity field of an aquifer from which transmissivity varies on
411 about 5 orders of magnitude. We assume that the water table is mainly controlled by
412 topography (generally the case for medium- to low-permeable aquifers) and that groundwater
413 flux is mostly sub-horizontal. Consequently, where vertical flux is significant, transmissivity
414 or hydraulic conductivity would be under-estimated.

415 Its application to the peridotite Tiébaghi Massif (New Caledonia) has given very encouraging
416 results with only 11% deviation between two LogT maps computed with different hydraulic-
417 head data sets (Oct.-Nov. 2006 and May 2007). In addition, there is quite good consistency
418 between measured and computed LogT as, in 71% of cases, the computed-LogT is estimated
419 with an error of <20%. The proposed approach will thus give the general pattern of
420 transmissivity or hydraulic conductivity, but cannot correctly reproduce strong variations at
421 very local scale (<10 m). Maps highlight the spatial distribution of the hydrodynamic
422 parameters and show a pattern clearly controlled by the fault network of the massif. However,
423 compared to other crystalline aquifers, the intense weathering along faults seems to seal the
424 structures.

425 The method can be very useful for siting bore wells (e.g. for water supply), but also for
426 identifying potential draining zones. For mining, it may provide valuable information for
427 mine design, such as managing the risk of groundwater inflow, designing the pumping
428 capacity for open pits, or safeguarding ore-storage and tailings areas.

429 In terms of numerical modelling, the resulting hydrodynamic parameters, even if given with a
430 certain amount of uncertainty, can help in identifying the general pattern of parameters, thus
431 providing valuable information on aquifer heterogeneity.

432 Further works should test the robustness of the evaluated parameters with a numerical
433 approach, applying the method to other aquifers including the ones where the water table is
434 mainly recharge controlled, and adapting it to larger aquifers with scarcer data sets. Another
435 important issue would be to define the optimal number of measurements required (hydraulic
436 head, and hydraulic tests) to use this method. A priori, it should depend on the heterogeneity
437 of the aquifer, highly heterogeneous aquifers, as the one presented here, requiring a very
438 dense observation network. Geostatistical tools are foreseen to achieve this issue.

439

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447

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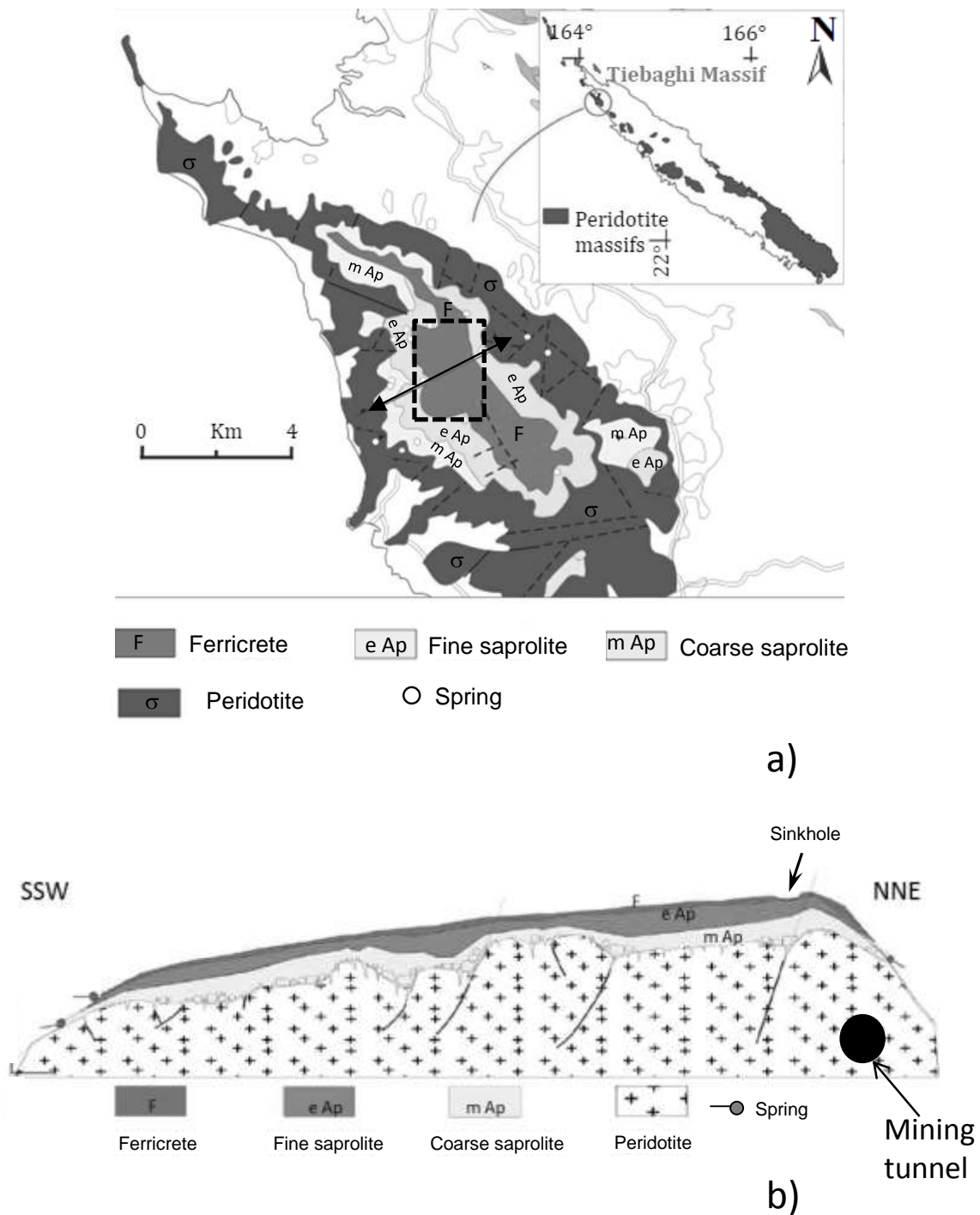
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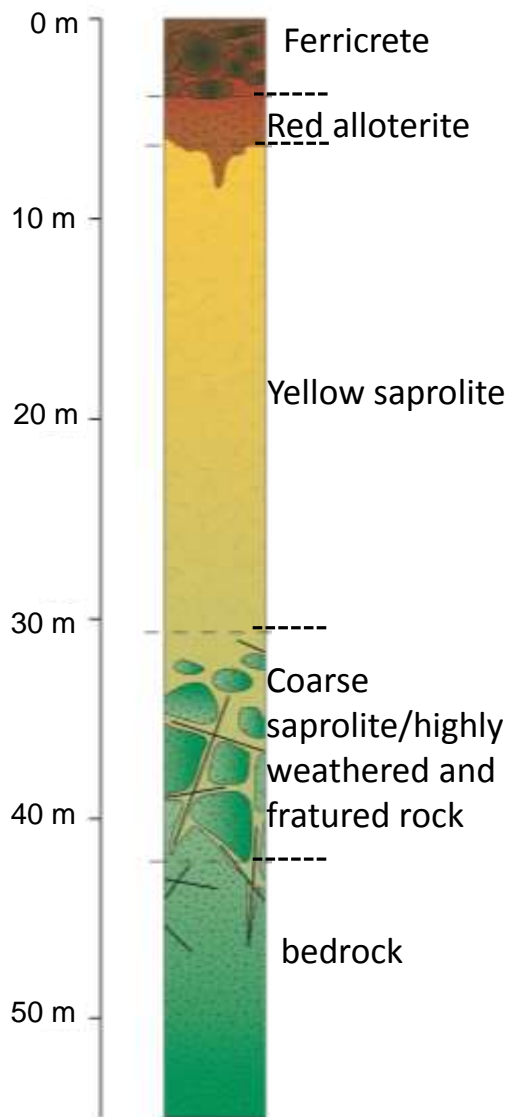
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598 **Figures**



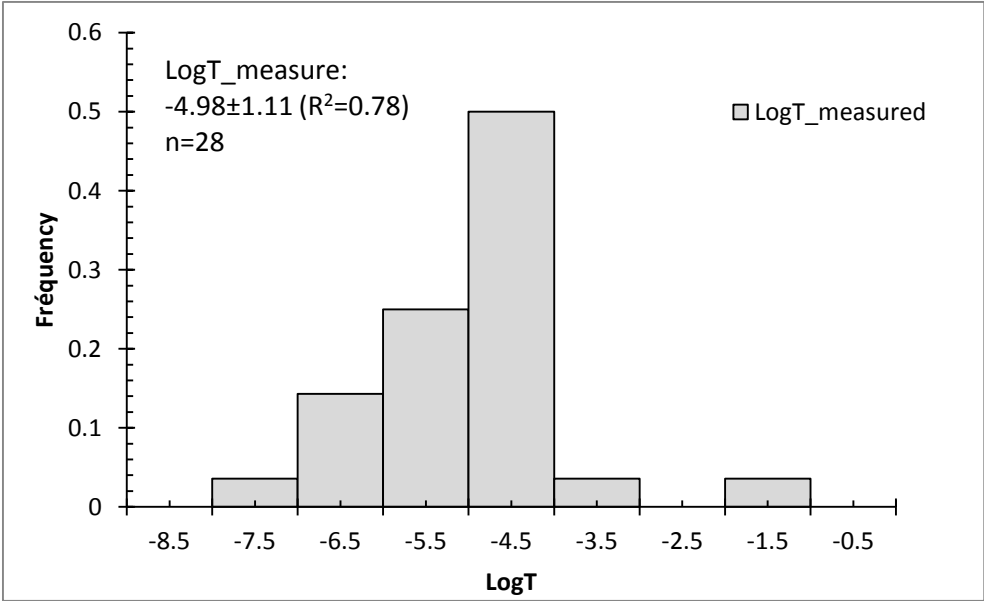
599

600 **Figure 1** a) Simplified geological map of the study area (3.5 km², Tiébaghi Massif, New
 601 Caledonia). b) Geological cross section (arrow on Fig. 1a) showing the stratiform structure of
 602 the weathering profile (modified from Join et al., 2005) and approximate location of the
 603 chrome-mining tunnel.

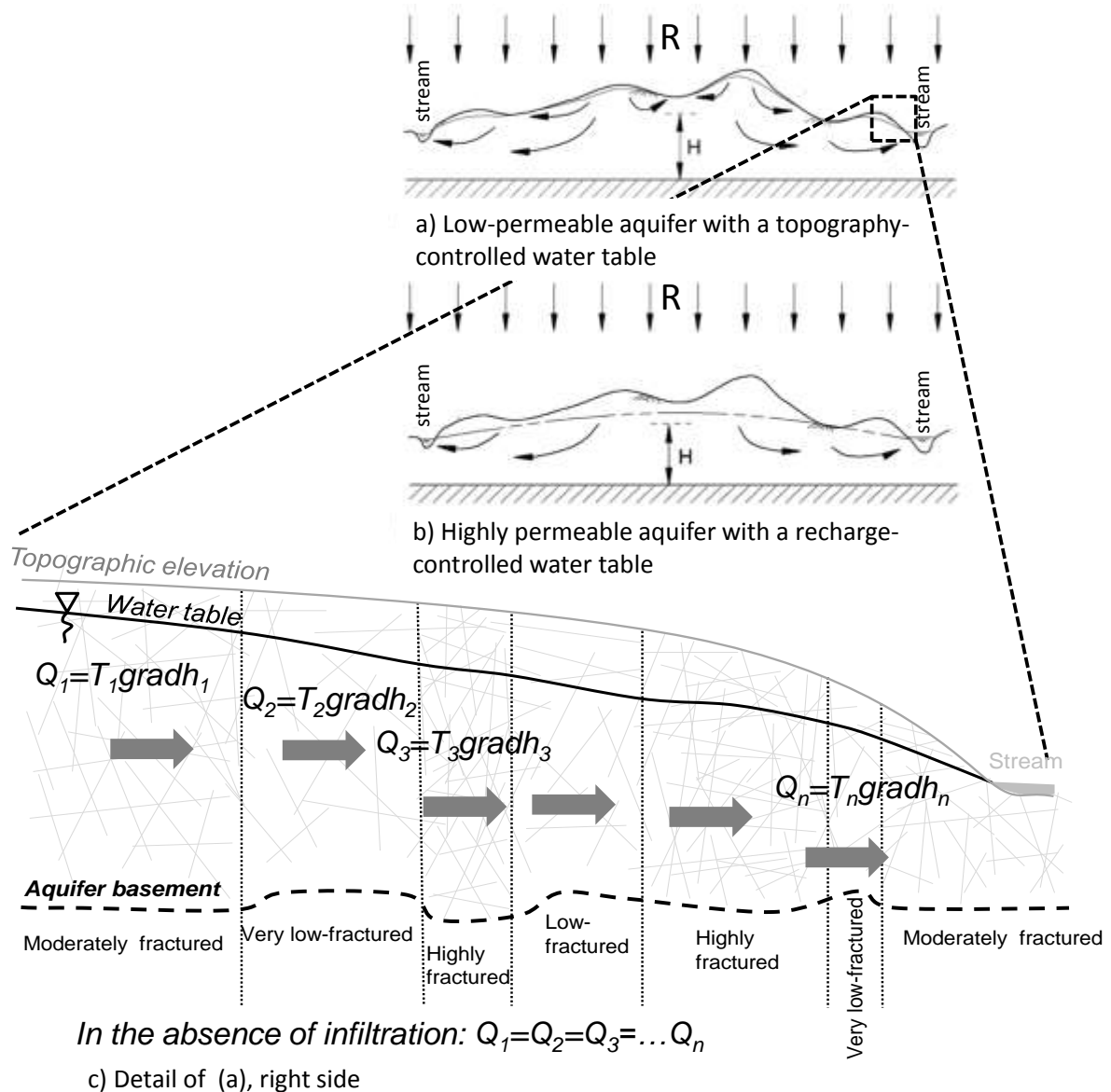


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605 **Figure 2** Typical weathering profile of ultramafic rocks in New Caledonia (Trescases, 1975;
606 Ouangrawa et al., 1996; Sevin et al., 2012).

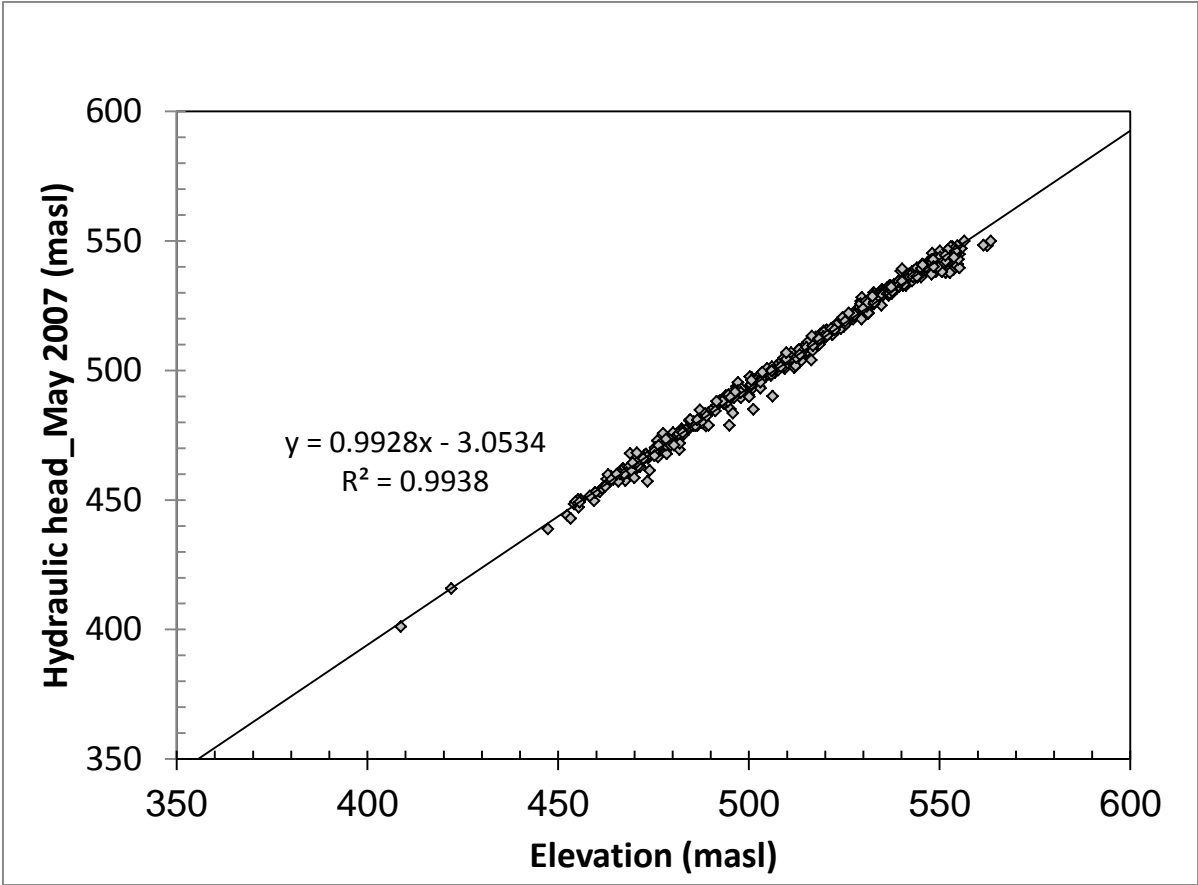


615
616 **Figure 4** Distribution on a logarithmic scale of transmissivity data (n=28). R: linear
617 regression coefficient, ±: standard deviation.

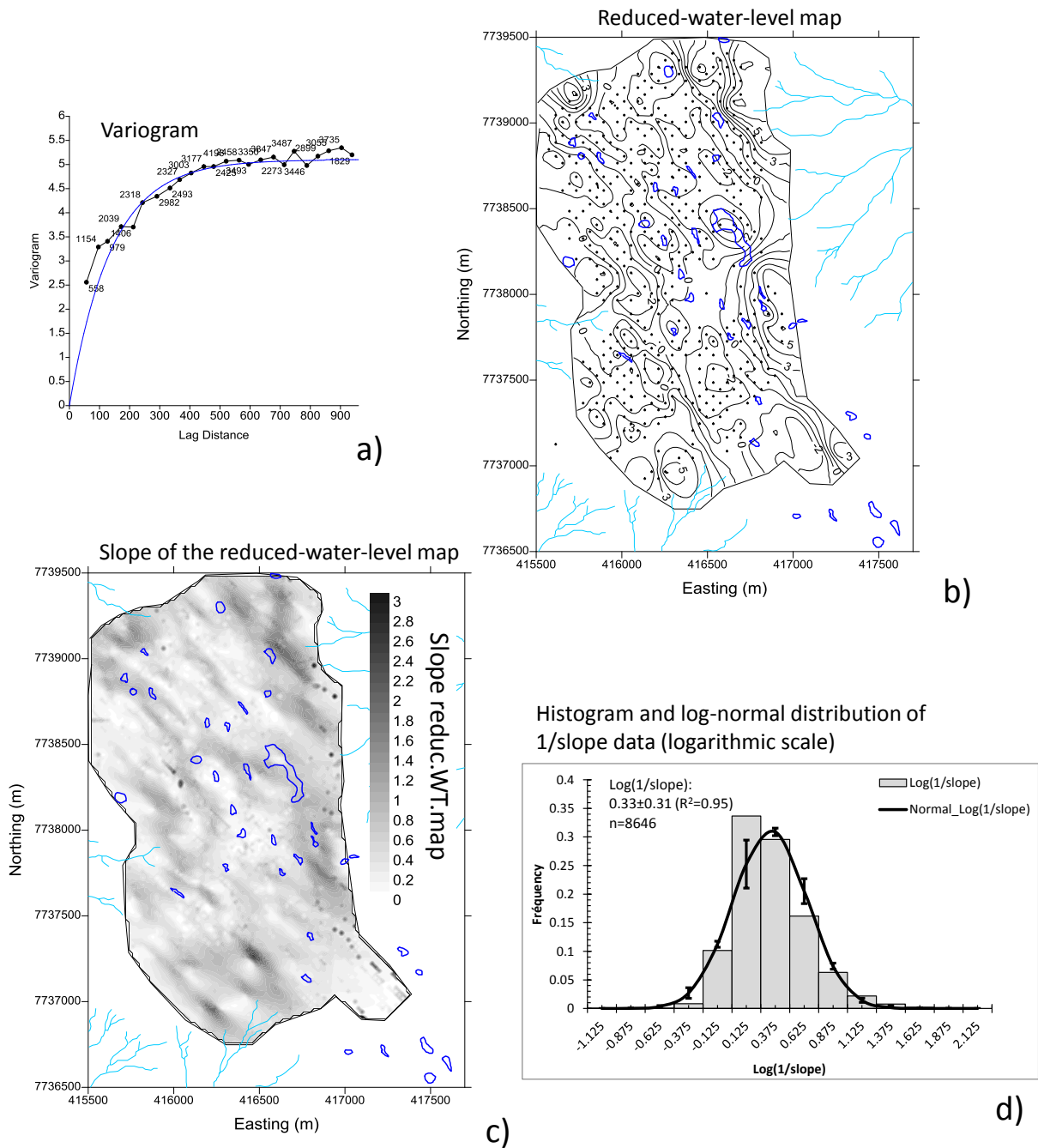


618

619 **Figure 5** Sketches showing the water table controlled by: a) Topography (low-permeable
 620 aquifers), and b) Recharge (highly permeable aquifers); modified from Haitjema and
 621 Mitchell-Bruker (2005). c) Detail on topography controlled water-table with expected
 622 importance of the topographic slope and of the aquifer transmissivity on the hydraulic
 623 gradient. Q : horizontal groundwater flow; T_1, T_2, \dots, T_n : horizontal aquifer transmissivity and
 624 $\text{grad}h_1, \text{grad}h_2, \dots, \text{grad}h_n$: hydraulic head gradient.

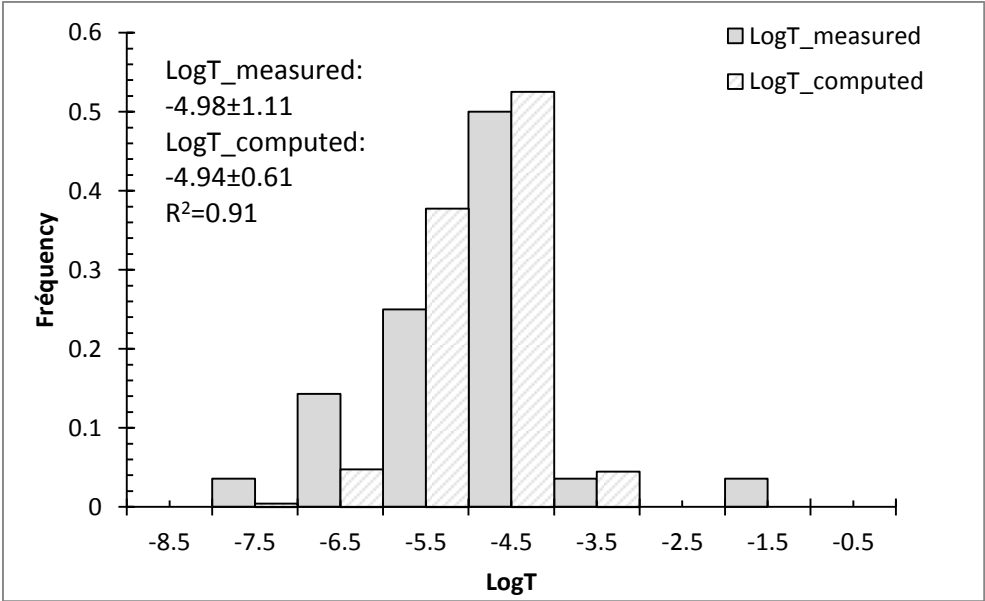


625
626 **Figure 6** Plot of hydraulic head measurements (May 2007) vs. elevation (in metres above sea
627 level; masl), n=475. R: linear regression coefficient.

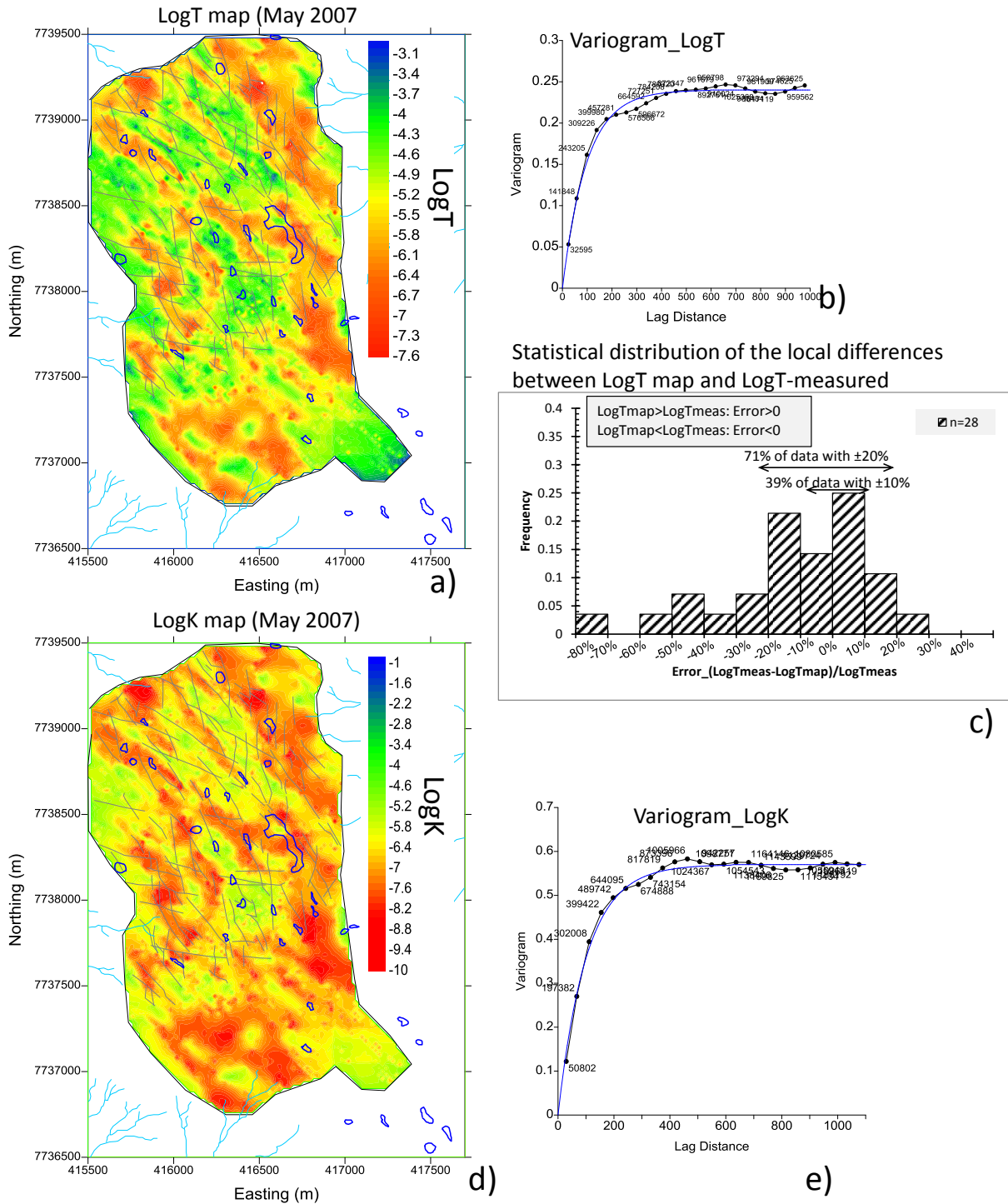


628

629 **Figure 7** Hydraulic-head data reduced from topographic influence (reduced-water level).
 630 a) Variogram model used for establishing Fig. 7b (exponential model, sill: 5.1, length: 190 m;
 631 anisotropy ratio: 1.8, dir.: N140°); number near data points represents data pairs
 632 measurements. b) Reduced-water-level map over 20x20 m cells; dots: hydraulic head data.
 633 c) Slope map established from Fig. 7b. d) Histogram and log-normal distribution on a
 634 logarithmic scale of 1/slope data (n=8646). R: linear regression coefficient, \pm : standard
 635 deviation.

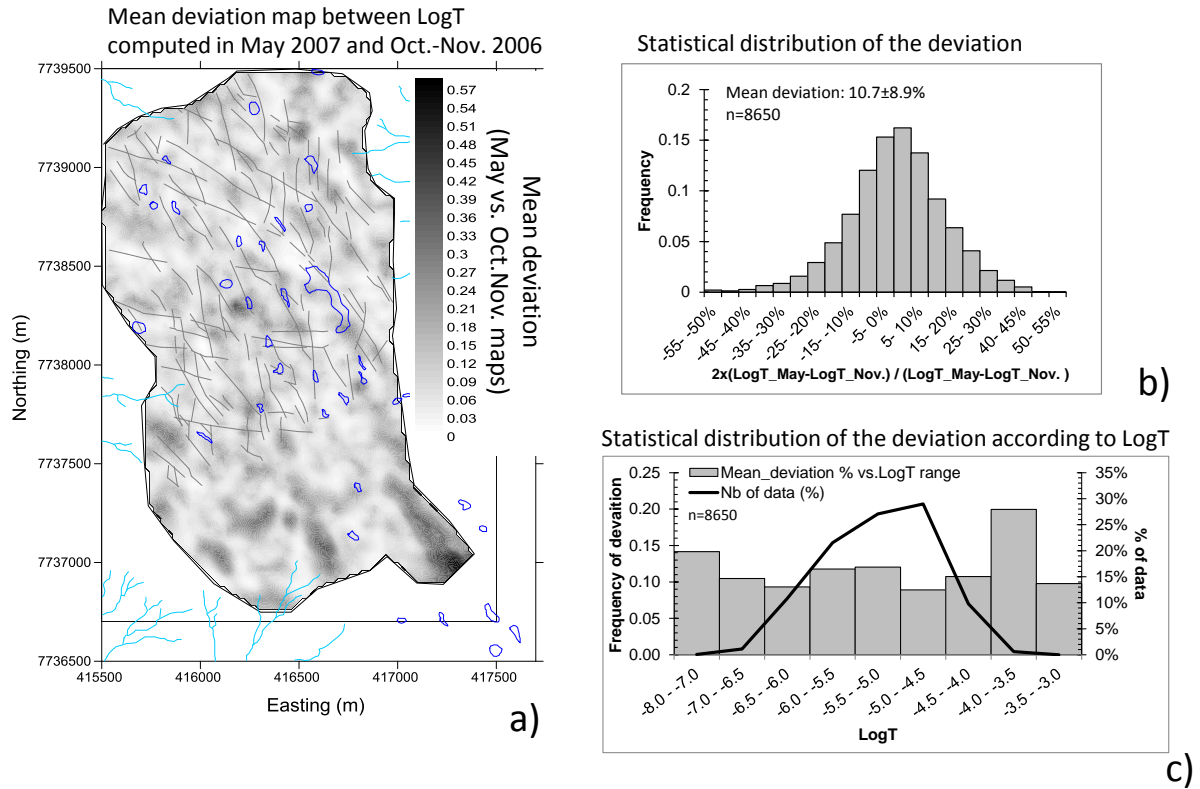


636
637 **Figure 8** Comparison of the distribution on a logarithmic scale of the transmissivity data
638 modelled with 1/slope data, with those from hydraulic tests. R: linear regression coefficient
639 between the two distributions, ±: standard deviation.



640

641 **Figure 9** a) Transmissivity map based on 1/slope data, LogT (20x20 m cells). b) Variogram
 642 model of LogT (exponential model, sill: 0.24, length: 115 m; anisotropy ratio: 1.4, dir.:
 643 N140°). c) Statistical distribution of the local differences between LogT based on 1/slope data
 644 and LogT measured from hydraulic tests. d) Hydraulic conductivity map LogK (20x20 m
 645 cells) established from LogT map, Fig. 3b and Fig. 3c. e) Variogram model of LogK data
 646 (exponential model, sill: 0.57, length: 120 m; anisotropy ratio: 1.4, dir.: N140°). On
 647 variograms, number near data points represents data pairs measurements.



653

654 **Figure 11** Mean deviation between LogT maps computed in May 2007 and Oct.-Nov. 2006.
655 a) Map. b) Statistical distribution. c) The same as b), but according to LogT ranges.