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

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9 Abstract

Estimating the transmissivity or hydraulic conductivity field to characterize the heterogeneity 10 of a crystalline aquifer is particularly difficult because of the wide variations of the 11 parameters. We developed a new approach based on the analysis of a dense network of water-12 table data. It is based on the concept that large-scale variations in hydraulic head may give 13 information on large-scale aquifer parameters. The method assumes that flux into the aquifer 14 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 measurements. The proposed approach can provide a general pattern of transmissivity, or 19 20 hydraulic conductivity, but cannot correctly reproduce strong variations at very local scale (less than10 metres), and will face of some uncertainties where vertical flows cannot be 21 22 neglected.

The method was tested on a peridotite (ultramafic rock) aquifer of 3.5 km² in area located in 23 New Caledonia. The resulting map shows transmissivity variations over about 5 orders of 24 magnitude (average LogT: -5.2 ± 0.7). Comparison with a map based on measured water-level 25 data (n=475) shows that the comparison between LogT-computed values and LogT data 26 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 this map a hydraulic-conductivity map has been computed showing values ranging over 8 29 30 orders of magnitude. The repeatability of the approach was tested on a second data set of hydraulic-head measurements (n=543); the mean deviation between both LogT maps is about 31 11%. These encouraging results show that the method can give valuable parameter estimates, 32 and can characterize aquifer heterogeneity. The computed LogT and LogK maps highlight the 33 spatial distribution of parameters that show a pattern clearly controlled by the fault network of 34 this ultramafic massif. However, the faults are mainly characterized by low-permeability 35 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 additional fieldwork. 39

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

42

1. Introduction 43

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

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.

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

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

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

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

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

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

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

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

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

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129 **2.2 Geology of the study area**

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

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

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

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151 **2.3 Hydrogeological setting**

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

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- 169

170 2.4 Water table and hydrodynamic data

Based on the hydrogeological database of the mine, a detailed water-table map was drawn from 475 exploration wells for May 2007, resulting in a very high density water-table map (Fig. 3c). These wells are not equipped (no casing, no screen) and drilled down to the bedrock (typically they enter 3 to10 m into the unfractured bedrock). Consequently, the resulting hydraulic head corresponds to an average value for the main coarse saprolite aquifer.

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

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

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

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3. Method for regionalizing transmissivity

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

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

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

212
$$Q=T_1gradh_1=T_2gradh_2=\ldots=T_ngradh_n$$

Where Q: is the horizontal flux $(m^3/s/m)$, Ti= aquifer horizontal transmissivity (m^2/s) , gradh_i: 213 hydraulic gradient (dimensionless). Because of the flat topography, the use of Eq.1 along the 214 same flow line will give high transmissivity values where the hydraulic gradient is low and 215 low values where the gradient is high. However, as shown on Figure 5c, the hydraulic 216 217 gradient is also strongly influenced by topography. Therefore, where the topographic level is almost stable (left part of Fig. 5c) the variation in hydraulic gradient depends more on 218 variations in aquifer transmissivity, in fact inversely related to the hydraulic gradient, whereas 219 in the right part of Fig. 5c it is more controlled by topographic slope because of stronger 220 221 variation in elevation.

(1)

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

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

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

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

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

251

252 **4- Results**

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

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

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

Applying the method described above, the resulting empirical relationship between both distributions of 1/slope and transmissivity shows a successful match between modelled and observed distributions, with a linear regression coefficient (square form), R², of 0.91 (Fig. 8). The relation is:

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

(3)

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

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

290 In order to verify if the produced map conforms to field data, local LogT-computed values from the map were compared to LogT data deduced from the 28 hydraulic tests (Fig. 9c). 291 292 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 LogT is estimated with an error less than 20%, which corresponds to LogT±0.4 on average. And in 39% of cases, the 294 computed LogT is estimated with an error less than 10%, here corresponding to LogT±0.2 on 295 average. Therefore, the method produces reasonable estimates of transmissivity considering 296 the strong heterogeneity of this fractured system with transmissivity variations of almost 5 297 orders of magnitude. In some cases, however-29% of the dataset corresponding to 60% of 298 299 the dataset with LogT>-4.5—, high-transmissivity areas are not correctly estimated by the map. This is due to the intrinsic characteristic of fractured systems, where high- and low-300 permeability zones coexist at a local scale. Therefore, these values are probably very local and 301 not representative of average aquifer properties at the scale of some tens of metres. This point 302 was already mentioned by Dewandel et al. (2012). 303

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

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

To verify if the method is repeatable and to validate it, we tested it on a second set of 313 hydraulic-head data. This new data set provides more water-table data (n=543), but they are 314 less precise because they were measured over a period of two months, between October and 315 316 November 2006, which may have introduced discrepancies because of potential small recharge events during the measuring period. The mean hydraulic-head value for this map is 317 507.0 metres above sea level, a value slightly lower than the one established in May 2007 318 (508.2 m.a.s.l.). A new relationship of elevation vs. hydraulic head (y = 0.9837x + 1.5371; R² 319 = 0.99) and a new Eq. 3 [T= 1.4x10⁻⁵ (1/slope) -0.9x10⁻⁵; R²: 0.88], etc., were established from 320 this data set to compute a new LogT map (Fig. 10a). This map shows a similar pattern to the 321 322 one established with May 2007 data (Fig. 9a). The data are still spatially structured (Fig. 10b), with a variogram model close to the previous one (model: exponential, sill: 0.32, length: 323 85 m; anisotropy ratio: 1.4, dir.: N140°). The average and standard deviation of the computed 324 LogT for 2006 is close to the earlier one, -5.2 ± 0.7 . 325

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

333

334 6- Discussion

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

The relationship between the distributions of inverse-slope (1/slope) and field-transmissivity 347 data does not fully satisfy Darcy's Law because of the constant in Eq. 3 (if it fully satisfies 348 349 this law, the constant should have been nil). The negative constant is believed to be introduced by the aquifer-scale relationship used for removing the effect of topography on 350 351 hydraulic-head data, which may differ on a local scale. Another reason that may also explain this bias is our basic assumption that vertical groundwater flux can be neglected compared to 352 the horizontal one. Though this hypothesis, at aquifer scale, is plausible because of minor 353 deep-groundwater flux as confirmed by measurements within the chrome-mining tunnel 354 355 (<7 l/s), it is possible that locally—within, or near, vertical faults—vertical flux cannot be neglected. Consequently, some high slope values on Fig. 7c may reflect such flux that the proposed method erroneously assimilates to low horizontal-transmissivity zones. In these cases, there is probably a vertical transmissivity component that the method cannot evaluate. The method also assumes that the water table is topography-controlled, which is the case at aquifer scale, but locally it may be recharge controlled, thus introducing an additional bias.

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

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

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

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- 400 **7. Conclusions**

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

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

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

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

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

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- 597

598 Figures



Figure 1 a) Simplified geological map of the study area $(3.5 \text{ km}^2, \text{Tiébaghi Massif, New Caledonia})$. b) Geological cross section (arrow on Fig. 1a) showing the stratiform structure of the weathering profile (modified from Join et al., 2005) and approximate location of the chrome-mining tunnel.



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



Figure 3 Tiébaghi Massif. a) Curves: Digital Elevation Model (20x20 m cells); Blue circles: springs; Blue shapes: sinkholes. b) Curves: elevation of the bottom of the coarse-saprolite layer; Gray lines: faults; Crosses: exploration boreholes (n=600). c) Curves: elevation of water table in May 2007; Crosses: location of hydraulic head data (n=475); Triangles: location of hydraulic tests (n=28); d) Variogram model used for establishing Fig. 3c (cubic model, sill: 70, length: 1200 m; anisotropy ratio: 1.4, dir.: N140°); number near data points represents data pairs measurements.



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



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Figure 5 Sketches showing the water table controlled by: a) Topography (low-permeable aquifers), and b) Recharge (highly permeable aquifers); modified from Haitjema and Mitchell-Bruker (2005). c) Detail on topography controlled water-table with expected importance of the topographic slope and of the aquifer transmissivity on the hydraulic gradient. Q: horizontal groundwater flow; T₁, T₂,.., T_n: horizontal aquifer transmissivity and gradh₁, gradh₂,..., gradh_n: hydraulic head gradient.



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



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



Figure 8 Comparison of the distribution on a logarithmic scale of the transmissivity data
 modelled with 1/slope data, with those from hydraulic tests. R: linear regression coefficient

639 between the two distributions, \pm : standard deviation.





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



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Figure 10 a) Transmissivity map based on Oct.-Nov. 2006 hydraulic-head data (n=543),
LogT (20x20 m cells). b) Variogram model of LogT (exponential model, sill: 0.32, length: 85
m; anisotropy ratio: 1.4, dir.: N140°); number near data points represents data pairs
measurements.



Figure 11 Mean deviation between LogT maps computed in May 2007 and Oct.-Nov. 2006.

a) Map. b) Statistical distribution. c) The same as b), but according to LogT ranges.