

Catchment-scale heterogeneity of flow and storage properties in a weathered/fractured hard rock aquifer from resistivity and magnetic resonance surveys: implications for groundwater flow paths and the distribution of residence times

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- 1 Catchment-scale heterogeneity of flow and storage properties in a
- 2 weathered/fractured hard rock aquifer from resistivity and magnetic
- 3 resonance surveys: Implications for groundwater flow paths and residence
- 4 times distributions
- 5

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- 14

15 Running Title

- 16 Flow and storage properties of fractured rocks
- 17

18 Abstract

- 19 Groundwater pathways and residence times are controlled by aquifer flow and storage properties,
- 20 which are characterised by high spatial heterogeneity in weathered/fractured hard rock aquifers.
- 21 Building on earlier work in a metamorphic aquifer in NW Ireland, new clay mineralogy and
- 22 geophysical data analyses provided high spatial resolution constraints on the variations of aquifer
- 23 properties. Groundwater storage values derived from magnetic resonance sounding and electrical
- resistivity tomography were found to largely vary laterally and with depth, by orders of magnitude.
- 25 Subsequent implementation of numerical, 2D-hillslope groundwater models showed that
- 26 incorporating heterogeneity from geophysical data in model parametrisation led to best fit to
- 27 observations as compared to a reference model based on borehole data only. Model simulations
- 28 further revealed that; 1/strong spatial heterogeneity produces deeper, longer groundwater flow
- 29 paths and higher age mixing in agreement with the mixed sub-modern/modern ages (mostly <50
- 30 years) provided by independent tritium data; 2/areas with extensive weathering/fracturing are
- 31 correlated with seepage zones of older groundwater, due to changes in the flow directions, and are
- 32 likely to act as drainage structures for younger groundwater on a catchment or regional scale.
- 33 Implications for groundwater resilience to climate extremes and surface pollution are discussed along
- 34 with recommendations for further research.

35 Weathered/fractured hard rock aquifers underlie over 20% of the global land surface (Sharp 36 2014) and are characterised by a high degree of structural heterogeneity and overall low 37 productivity. In recent years, water managers and policy-makers have moved to adopt a catchment 38 scale approach to the integrated management of surface and subsurface water resources (EU Water 39 Framework Directive 2000/60/EC), including the UK (UKTAG 2011) and Ireland (Daly et al. 2016). 40 Understanding spatial variations of aquifer hydraulic properties at the catchment-scale, which 41 dictates groundwater flow pathways and residence times, is crucial to inform catchment 42 management plans. Yet, resolving such spatial heterogeneity in fractured bedrock remains very 43 challenging due to the typically scattered nature of direct observations points (boreholes and 44 outcrops) that usually do not have sufficient spatial coverage to capture the scale of heterogeneity 45 (De Marsily et al. 2005; Neuman 2005). 46 To address the lack of spatial resolution in heterogeneous fractured rock catchments, traditional 47 direct testing techniques in boreholes, such as hydraulic testing and geophysical logging, are 48 increasingly combined with indirect and more spatially integrative investigation methods, including 49 tracer testing (e.g. Klepikova 2016), geophysical imaging (ground- or airborne-based) (Comte et al. 50 2012; Shakas et al. 2016; Day-Lewis et al. 2017) and remote sensing (Cassidy et al. 2014; Frances et

al. 2014). The use of geophysics has long proven effective in resolving, at catchment scale, the
heterogeneity of fractured rock aquifers (Holbrook *et al.* 2014; Robinson *et al.* 2016). The Electrical
Resistivity Tomography (ERT) method is known to be efficient at imaging spatial variability in
weathering, geological heterogeneity and fracture patterns (e.g. Chandra *et al.* 2010; Rainer *et al.*2007). All ERT studies, however, stress the importance of *a priori* information, especially borehole
data and outcrop mapping to support its hydrogeological interpretation (Skinner & Heinson 2004;
Comte *et al.* 2012).

58 A number of studies have demonstrated the benefits of using ERT in combination with other 59 geophysical methods; in particularly with the magnetic resonance sounding (MRS) that complements 60 imaging of heterogeneity by ERT with lower resolution but more quantitative information on water 61 storage. Both methods have, for instance, been used to map groundwater occurrence and develop 62 hydrogeological conceptual and numerical models in weathered basement aquifers (Frances et al. 63 2014; Baltassat et al. 2005) and monitor groundwater recharge (Descloitres et al. 2008). As yet, 64 however, most of these studies have focused on low latitude regions with deep and relatively water-65 productive weathering horizons (saprolite) that produce strong MRS and ERT responses for relatively 66 simple, layered aquifer geometries. There are fewer examples in higher latitude catchments with a 67 glacial legacy, such as in Ireland, where most of the saprolite (relatively high storage, porous layer) is 68 absent, exposing only the fractured (low storage) and structurally complex bedrock. In addition, 69 these geophysical approaches still remain either (i) not systematically applied in catchment 70 groundwater studies or (ii) applied qualitatively, i.e. used to inform aquifer heterogeneity conceptual 71 models rather than to quantify spatial variations in aquifer properties (permeability and porosity).

This needs further consideration in highly heterogeneous basement rock catchments.
 In Ireland, the fractured rock aquifers provide a good analogue of temperate region bedrock
 aquifers with a glacial legacy. The island of Ireland is underlain by over 60% of hydrogeologically

74 aquifers with a glacial legacy. The island of Ireland is underlain by over 60% of hydrogeologically 75 poorly-productive fractured bedrock (Moe et al. 2010), either cropping out directly or covered by 76 superficial glacio-fluvial and/or alluvial sediments. Most of this poorly-productive bedrock is 77 composed of various grades of metamorphic (basement) rocks; from low-grade metasediments to 78 high-grade gneiss-migmatites and granitoids. Groundwater in these rocks, despite their overall low 79 productivity, is nonetheless crucial for maintaining river base flow during dry periods and supporting 80 aquatic ecosystems and small-scale rural water supply (DCCAE 2017). Over the last decade, the 81 extension of the Irish national groundwater monitoring network to poorly productive basement 82 aquifers as part of implementing the European Union's Water Framework Directive (EPA 2006; Moe 83 et al. 2010) has stimulated hydrogeological research through Irish Government-funded projects. 84 Among them, the Griffith Poorly-productive Aquifers Project (2007-2014), on which this work is 85 based, aimed to improve the understanding of groundwater flow regimes in fractured rock aquifers

86 and the contribution of groundwater to catchment water balance (Comte et al. 2012; Cassidy et al. 87 2014; Caulfield et al. 2014; Cai & Ofterdinger 2016).

88 This paper presents an overview of the latest research in a micaschist catchment in Co. Donegal, 89 Republic of Ireland, with the aim of resolving catchment-scale spatial variations of aquifer 90 properties. It synthesises previously published research on aquifer typology and well-scale hydraulics 91 (Comte et al. 2012; Cassidy et al. 2014), bedrock weathering (Caulfield et al. 2014) and aquifer 92 storage properties (Legchenko et al. 2017), augmented with the latest results from quantitative 93 interpretation of geophysical data (ERT and MRS) to provide a robust spatial understanding of flow 94 and storage property fields. This work further explores the integration of the existing knowledge on 95 the heterogeneity of aquifer properties using numerical models to assess groundwater flow paths 96 and residence time distributions. Finally, we use the results to discuss the impact of climate change 97 and contaminant transport on groundwater resources and catchment management, as well as 98 further recommendations for improved groundwater modelling. 99

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101 Hydrogeological setting

103 Geology

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The Gortinlieve catchment (5 km²), Co. Donegal, NW Ireland (Figure 1) is underlain by Late 105 106 Precambrian micaschists and psammites of intermediate metamorphic grade (low amphibolite 107 facies). These belong to the Grampian terrane in Co. Donegal as part of the Southern Highland 108 Group, spanning through Ireland and Scotland. They originate from turbidite sequences deposited c. 109 550 Ma BP (McConnell & Long 1997; Caulfield et al. 2014) and subsequently subjected to poly-phase deformation and metamorphism during the Caledonian orogeny. The Caledonian tectonic regime 110 111 associated with the closure of the lapetus Ocean (Grampian phase, early Ordovician) generated the 112 current regional NE–SW oriented structures (Chew 2009) characterised by fault scarps and 113 cartographic/topographic lineaments visible in the upper catchment (Fig. 1). The later Taconic phase 114 (late Ordovician) generated the current WNW-ESE orientation deformational structures and 115 imprinted the retrograde amphibolite metamorphic facies (McConnell & Long 1997). The current, 116 Alpine, strike-slip tectonic regime, is reflected by reactivation of NE-SW fracture orientations and 117 further creation of a general NW-SE trend (Worthington & Walsh 2011; Cooper et al. 2012). Localised kaolinite rich Tertiary lateritic horizons preserved on Grampian rocks of western Ireland 118 119 (Legg et al. 1985) provide evidence that the basement was during this time at least partially 120 exhumed and undergoing tropical weathering. Subsequent Quaternary glaciations have resulted in 121 further erosion, including the removal of the upper levels of the weathering profiles, most 122 pronounced in high topographic areas, and in the deposition of heterogeneous glacio-fluvial material 123 (clay-till and sand and gravel) in low-lying areas and valley bottoms.

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125 Hydrogeology and borehole instrumentation

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127 Annual rainfall in the area ranges from 1000 to 1200 mm and mean temperature annually ranges 128 from 6 to 14 °C (Met Eireann 2017). The catchment comprises a headwater stream network of a 129 Carrigans River, a tributary of the River Foyle that discharges into the Atlantic Ocean, approx. 40 km 130 NE of the catchment (Caulfield et al. 2014). The Gortinlieve catchment was instrumented with monitoring boreholes by the Irish Environmental Protection Agency (EPA) in 2006 as part of the 131 132 national groundwater monitoring programme. Three borehole clusters were sited in a linear 133 hillslope transect at high (GO1, 174 m above mean sea level; hereafter noted m amsl), intermediate 134 (GO2, 88 m amsl) and low (GO3, 33 m amsl) elevations within the catchment. Individual clusters 135 contain up to 4 boreholes, each isolated and screened across different depth-distinctive zones 136 commonly encountered in Irish bedrock aguifers. The initial classification of Moe et al. (2010)

137 conceptually described these respective intervals as; subsoil SS (average interval 1–3 m below

- 138 ground surface (bgs); only present in Gortinlieve at the valley bottom), transition zone T (average
- 139 interval 4–5 m bgs), shallow bedrock S (average interval 8–19 m bgs) and deep bedrock D (average
- 140 interval 30–67 m bgs). The conceptual model was further refined by Comte *et al.* (2012) in order to
- help reconcile geological features with geophysical constraints (Fig. 2): overburden deposits (cf.
- subsoil); broken bedrock (cf. transition zone); fissured bedrock (cf. shallow bedrock); massive
 bedrock (cf. deep bedrock). The Irish EPA monitors water levels at 15-minute intervals in each
- borehole using automatic data loggers. The site was also equipped with an automated tipping
- 145 bucket rain gauge (AEG 100) since October 2010 for the duration of the project.
- 146 147

148 Aquifer characterisation methods

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150 Structural, mineralogical and hydraulic investigations

Detailed analysis of fracture patterns and clay occurrence in the bedrock was carried out in order to establish the micro- to meso-scale structural controls on groundwater flow, and to provide constraints on larger (meso- to macro) scale hydraulic and geophysical interpretations and numerical modelling.

156 Fracture orientations were measured on maps, outcrops, boreholes and guarry exposures. 157 Regional structural trends were determined from interpretation of the geological map (Smith 1991; 158 Long et al. 1992; McConnell & Long 1997) and the 20-m resolution digital elevation model (Ordnance 159 Survey of Ireland). During field mapping, the main fracture parameters measured were strike, dip 160 magnitude, dip azimuth with a minimum of 30 fracture measurements for each sample site, 161 subsequently plotted as rose diagram and Schmidt net pole density distribution. The acquisition and 162 plotting methodology are described in details in Comte et al. (2012) and Nitsche (2014). Borehole 163 fractures were analysed using acoustic televiewer logs. Among these, hydraulically active fractures 164 were identified through cross-correlation with electrical conductivity logs whereby marked changes 165 in water electrical conductivity at the depth of observed fractures was interpreted as fracture flow 166 (Nitsche 2014).

Mineralogical and petrographic data are summarised by Caulfield et al. (2014). Representative 167 field outcrop and recovered borehole core samples were further characterised in this study to 168 quantify the relative proportions of identified phyllosilicate minerals to assess their influence on 169 170 geoelectrical properties. The mineral modes of the dominant basement lithologies (psammitic 171 schists uphill vs. micaschists downhill) were quantified by petrographic point-counting (500 points 172 per slide). These compositions were taken to represent fresh, primary (unweathered) bedrock. Free 173 phyllosilicate minerals (mineral grains not bound in the competent fresh bedrock structure) were 174 separated and collected from weathered samples by repeated washing with distilled water and dried 175 at 50°C. Clay sized (<2 µm) fractions were obtained following the method outlined in Caulfield *et al.* 176 (2014), and references therein. The relative proportions of primary (muscovite, chlorite) and 177 secondary (illite, montmorillonite and illite/montmorillonite admixtures) were determined via 178 thermogravimetric analysis (TGA) using a Netzsch libra thermogravimetric analyser, in conjunction 179 with XRD and FTIR results from Caulfield et al. (2014). Samples were heated from 200-880°C to 180 determine sequential dehydroxylation water loss from the clay mineral fraction. Mineral proportions 181 (volume %) were converted to weight %. Using the method of Revil et al. (1998), the total cation 182 exchange capacity (CEC) of the different geological units (transition, shallow, deep bedrock) at each of the three sites (GO1-3) were calculated from (i) the relative mass proportion of clay minerals, (ii) 183 184 their individual CEC, obtained from literature (Swineford 1955; Carroll 1959; Wiklander 1964; 185 Thomas 1976; Ridge 1983; Revil et al. 1998; Crain 2000; Gillespie et al. 2001; Ellis & Singer 2007; 186 Henn et al. 2007) and (iii) the total clay content derived from natural gamma ray logs from each 187 borehole.

188 Aquifer pumping and recovery tests were implemented in every borehole to provide local values 189 of equivalent hydraulic conductivity and storativity for each borehole. The hydraulic testing 190 methodology is described in details in Comte et al. (2012). Pumping test were conducted at a 191 constant rate (2 to 30 L/min depending on the borehole) and both pumping and recovery curves 192 were jointly interpreted using AQTESOLV Pro v4.5 and applying a range of adapted analytical 193 solutions (with regards to the known aquifer structure and the borehole technical characteristics) 194 including single porosity/permeability models. Hydraulic conductivities (K) were calculated from 195 transmissivity values using observed unit aquifer thicknesses from borehole data, ERT data and 196 geophysical logging data.

197 K in fractured rocks is typically anisotropic, i.e. directionally dependent implying different K 198 values in different directions, with the maximum hydraulic conductivity Kmax following the direction 199 of the hydraulic active fractures and the minimum hydraulic conductivity Kmin being orthogonal to 200 the hydraulic active fractures. This further implies that the anisotropy angle in fractured rocks, 201 because of the presence of steep fractures, has a higher vertical component than in most 202 sedimentary rocks where the direction of Kmax is often sub-horizontal, parallel to the sedimentary 203 bedding (in this case Kmax is usually denoted Kh) and the direction Kmin is subvertical (and in this 204 case denoted Kv). In this work, values of hydraulic conductivities provided by hydraulic tests were 205 further analysed in terms of anisotropy in 2D (along the studied transect). As hydraulic test values 206 were assumed equivalent to isotropic hydraulic conductivities, the integration of fracture analysis 207 data allowed for the definition of Kmax Kmin and anisotropy angles (by definition the angle of Kmax 208 to the horizontal) based on the measured orientation of the hydraulically active fractures. The 209 anisotropy ratio Kmin/Kmax (unitless) was derived by considering it to decrease with depth from 0.5 210 to 0.1. At depth, where the steep fracture sets dominated (see section Aquifer characterisation 211 results) and produced a stronger anisotropy, the ratio was set to 0.1, whereas at shallow depths, 212 where other fracture sets and pronounced weathering contribute to flow, particularly in the broken 213 bedrock, anisotropy was expected to weaken and was therefore set to 0.5. The anisotropy angle was 214 derived as an average of the individual angles of the dominant fracture sets for each bedrock 215 conceptual unit (broken, figured, massive).

- 217 *Geophysical investigations*
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19 Electrical resistivity tomography and petrophysical models for porosity estimation

221 A geophysical survey using electrical resistivity tomography ERT was carried out to provide a 222 catchment-scale conceptual understanding of the bedrock heterogeneity and its hydrogeological 223 significance. ERT was applied along the hillslope transect of the three well clusters using a Syscal Pro 224 72 resistivity meter. As described in Comte et al. (2012), the acquisition was carried out with 60 225 electrodes at 5-m unit electrode spacing, subsequently expanded through roll-along. The dipole-226 dipole (DD) and multi-gradient (mGD) quadripole configurations ran sequentially, ultimately allowing 227 a depth of investigation of 50–60 m (Edwards 1977). After noise removal, DD and mGD apparent 228 resistivity data were jointly inverted using RES2DINV v3.58. The inversion provided a 2D model of 229 specific resistivities in which model regions that are insensitive to the input data were removed 230 using the depth of investigation (DOI) method of Oldenburg & Li (1999). The best model retained for hydrogeological interpretation was obtained for the 5th iteration, giving an absolute error between 231 232 observed and calculated apparent resistivities of 9.6%.

In addition to the structural information provided by the interpretation of spatial resistivity
variation, a petrophysical model was applied to the resistivity model in order to derive bedrock
porosity. Due to the substantial amount of clay minerals (both primary and secondary) in the
psammites and the micaschists, Archie's model (Archie 1942) was not applicable and instead, the
Waxman and Smits (1976) model was used. The Waxman and Smits model relates the bulk (specific)
resistivity obtained by inversion to the aquifer total porosity (which is the unknown to resolve), the

pore water saturation (equal to one below the water table), the rock matrix cementation factor, the groundwater electrical conductivity and temperature, and in addition to Archie's Model, the total

groundwater electrical conductivity and temperature, and in addition to Archie's Model, the total
rock CEC, which was obtained from mineralogical investigations and natural gamma logs (see section

Aquifer characterisation results). The model is described in Revil et al. (1998) and was applied to
 estimate the 2D (hillslope) distribution of total porosity along the transect imaged by ERT. Table 1

- estimate the 2D (hillslope) distribution of total porosity along the transect imaged by ERT. Table 1summarises the values of the parameters used for the different hydrogeological units.
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246 Magnetic Resonance Sounding

247

248 The magnetic resonance sounding (MRS) method (Legchenko 2013) was applied at 11 points 249 along the ERT transect and close to the borehole clusters (Fig. 1). The MRS surveys provided 1D 250 (vertical) profiles of water content. MRS measurements were performed in two surveys, in 2010 and 251 2016 using the NUMISPLUS and NUMISPOLY instruments, respectively, both developed by IRIS 252 Instruments (France). As described in more detail in Legchenko et al. (2017), figure-eight square 253 loops were applied in three configurations: two cable turns of 25-m-side, one turn of 37.5-m-side, 254 and one turn of 50-m-side, with a reference loop in 2016 to improve the signal to noise ratio, which 255 is typically low in fractured rock environments, due to a low water content. Eight out of the 11 MRS 256 soundings had acceptable signal-to-noise ratios; the remaining three are not considered in this work. 257 1D inversion of MRS data was carried out using the SAMOVAR software package. It used the 258 Tikhonov regularization method (Legchenko & Shushakov 1998) and the uncertainty in the inversion 259 results was examined applying Monte-Carlo simulations (Legchenko et al. 2017). Geological 260 interpretation of MRS results was performed taking into account the thicknesses of the aquifer units 261 (broken and fissured zone) identified with ERT as structural constraints.

262

263 Estimation of aquifer storage parametres from ERT porosity and MRS water content 264

265 ERT porosity and MRS water content were used to derive aquifer specific yield *Sy* and effective 266 porosity *n_e* subsequently used for recharge estimation, and numerical groundwater modelling, 267 respectively. By definition, *n_e* is the interconnected porosity controlling groundwater mass transport 268 and residence times. Also by definition, *Sy*, also called drainage porosity, is the mobile pore water 269 not retained by capillarity forces and controls water table fluctuations in unconfined aquifers.

270 In fractured rocks, where porosity is mostly the result of the opening of fractures and foliation 271 planes by (groundwater-aided) weathering, the occurrence of closed (unconnected) pore space is 272 limited (Singhal & Gupta 2010). This implies that total porosity, obtained from ERT, can be 273 considered a close estimate of the effective porosity n_e , which was our assumption for subsequent 274 modelling.

The MRS water content is commonly reported (Lubczynski & Roy 2004) as being correlated with, and representing an intermediary between, the drainage porosity *Sy* (mobile pore water non retained by capillarity forces) and the effective porosity *n_e* (immobile and mobile interconnected pore water). Recent studies carried out in low porosity fractured environments (Vouillamoz *et al.* 2012, 2014) have reported MRS water content values linearly correlated to, although higher than *Sy* (within a factor 2) estimated through pumping tests. Consequently, for subsequent recharge calculations we have taken *Sy* at half (0.5 times) the measured MRS water content.

- 282
- 283 Recharge estimations

Recharge values were calculated with the water table fluctuation method applicable to recharge
estimation in unconfined aquifers (Healy & Cook 2002). The method requires water table time-series
as well as values of the specific yield at depth where the water table fluctuates. We used the
piezometric fluctuations in the shallowest boreholes (GO1T, GO2T and GO3SS) as reported in Cai &
Ofterdinger (2016) for the period 2010-2012. For the specific yield, we used the MRS derived values

- obtained at the approximated depth of the water table (see previous Section) from the MRS
 soundings nearest to the boreholes; MRS1 and MRS2 for GO1T, MRS4 and MRS5 for GO2T, and
 MRS8 for GO3SS.
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294 Aquifer characterisation results

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296 Aquifer structure and weathering

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298 Fracture analysis

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300 Measurements of hydraulically active fracture from both outcrops and boreholes at the site 301 showed dominant meso-scale (millimetre to metre) fracture occurrences and orientations which 302 correlate with established past tectonic regimes in Ireland (Fig. 3). Two dominant hydraulically active 303 fracture sets were identified: a dominant set with fractures oriented E-W to NE-SW with a dip angle 304 of about 60-80° S to SE; and a less frequent set with fractures also oriented E-W to NE-SW but with a 305 dip angle of about 60-90° N to NW. The dominant fracture set may be attributed to the strike-slip 306 regional fracturing associated with Alpine compression during the Palaeocene, whereas the second 307 set may be attributed to the compressive fabric of the Neo-Proterozoic to late Carboniferous 308 orogenies (Grampian, Caledonian and Variscan), which also characterise the regional-scale 309 lineaments (Fig. 1) and Lower Carboniferous normal faults (Worthington & Walsh 2011), reactivated by the Alpine strike-slip regime (Cooper et al. 2012). The most frequently recorded fracture set is 310 311 oriented WNW–ESE with a low dip angle of 35 ° NNE corresponding to the highly weathered 312 Dalradian schistosity plane (see below). However, this set does not appear to be clearly associated 313 with hydraulic activity in boreholes and therefore may only play a role in the uppermost levels of the 314 aquifer (broken bedrock and possibly the upper part of the fissured bedrock) for which hydraulic 315 activity in boreholes is difficult to record due to their shallow depths and denser

316 weathering/pervasive fracturing.

At macro-scale (catchment and regional scale, i.e. 100 m to km; Fig. 1b), cartographic lineaments attributed to glacial erosion preferentially affecting regional geological boundaries and deep weathering corridors were found to be typically associated with the measured pre-Alpine NE–SW trend (Comte *et al.* 2012).

- 321
- 322 Mineral analysis

323

324 Mineralogical investigations reveal that phyllosilicate minerals are dominated by muscovite and 325 chlorite (primary) and illite and montmorillonite (secondary clays produced by weathering of chlorite 326 and muscovite) (Caulfield et al. 2014). Major non-clay minerals comprise quartz and feldspar. 327 Relative proportions of clay minerals versus the total clay mass fraction are presented in Table 2. 328 The massive (unweathered) bedrock encountered in the deep boreholes was found to contain 329 negligible secondary clays, with only primary chlorite and muscovite observed. The fissured zone in 330 boreholes GO2 and GO3 contained an approx. equal mixture of illite and montmorillonite with lesser 331 amounts of chlorite and muscovite, and in GO1, relatively even quantities of the four minerals. This 332 suggests a moderate weathering of the fissured zone primary clays in GO1, whilst in GO2 and GO3, 333 they have undergone major weathering as evidenced by the almost complete absence of chlorite. 334 Over 80% of clays in the broken zone were muscovite and the remaining 20% approx. equal

proportions of illite and montmorillonite. Because weathering is most active in the broken zone, it
was concluded that secondary clays are transported in suspension downward from the broken zone
and redeposited in the underlying fissured zone (Comte *et al.* 2012; Caulfield *et al.* 2014).

Table 2 also details the total clay mass fraction as derived from natural gamma ray logs. The
relative proportions of clay minerals and the total clay mass fraction allowed calculation of the total
CEC of the different aquifer units using the model of Revil *et al.* (1998). The total clay content of the

- 341 different aquifer units (broken, fissured and massive bedrock) was not significantly different;
- however due to significant differences in clay mineralogy, the resulting CEC is different. The fissure
- 343 bedrock, with a higher proportion of illite and montmorillonite (secondary weathering clays
- 344 characterised by high CECs) consistently produced the highest bulk CEC values. The massive bedrock
- dominated by muscovite and chlorite (primary clays characterised by low CEC values) produced the
 lowest bulk CEC. This CEC distribution is relevant to the interpretation of ERT results and application
- 347 of the petrophysical models in which it is a key input parameter.
- 348

349 Geophysical structure

350

351 Results of the ERT investigations provide cross-sectional variations of electrical resistivity along 352 the borehole hillslope transect (Figure 4), revealing a high degree of aquifer heterogeneity. As 353 electrical resistivity is primarily controlled by porosity (fracture porosity in the present context), clay 354 content and mineralogy, its spatial variations can be interpreted as reflecting changes in lithology, 355 pore water saturation, open fracture density and weathering intensity. Resistivity values range from 356 about 100 ohm.m to just under 10 000 ohm.m, i.e. three orders of magnitude. Highest resistivities 357 (>> 1000 ohm.m) were observed at the base of the transect and reflect the massive micaschist unit 358 beneath the weathered front, characterised by a low density closed fractures and small quantities of 359 secondary weathering clays. Some equally high resistivity values were also obtained in the very 360 shallow subsurface in the higher elevation part of the profile. Correlations with the depth of the 361 water table in borehole cluster GO1 and observed spring lines at its base suggested that this 362 uppermost, high resistivity unit corresponds to unsaturated, clay-poor psammite-schist. The lowest 363 resistivity values (<500 ohm.m) were obtained at a relatively shallow depth in the flood plain and 364 corresponded to the clay-till and alluvium overburden. Intermediate resistivity values (500-2500 365 ohm.m) were observed throughout the whole transect at shallow to intermediate depths (0 to about 366 50 m below surface) and characterised the heterogeneous weathered/fractured schists (broken and 367 fissured bedrock). The relatively continuous unit with values ranging ~300-1000 ohm.m overlying the 368 massive bedrock are interpreted as fissured bedrock characterised by a high density of open 369 fractures possibly filled with high-CEC weathering clays as suggested by the clay mineralogy. This 370 fissured layer was particularly deep at three locations along the transect; at the following horizontal distances from NW end of the profile of X=400-500 m, 750-850 m and 1100-1300 m. Since they 371 372 correlate well with the regional Grampian lineaments (Fig. 1), these locations are interpreted as 373 deep fractured zones that facilitate deep weathering. At these three locations a slightly more 374 resistant layer was observed overlying the fissured layer, with resistivities in the range of 500-2500 375 ohm.m. This unit is interpreted as the broken schists (transition zone) where the weathered clays 376 have been leached and transported to the underlying fissured zone.

378 Flow properties

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380 Pumping and recovery tests in every borehole provided estimates of (equivalent isotropic) 381 hydraulic conductivities K for bedrock and overburden (Table 3). A clear trend of decreasing hydraulic conductivity with depth is observed. The broken bedrock (transition zone) displayed mean 382 values close to 10⁻¹ m/d, the fissured bedrock close to 10⁻² m/d with large variability of over two 383 orders of magnitude and the massive (deep) bedrock around 6×10^{-3} m/d. The overburden (glacio-384 385 fluvial and alluvium deposits) present in the valley floor displayed much higher mean K values of ~9 386 m/d. These values partly agree with values from previous hydraulic tests by Moe et al. (2010). These 387 authors obtained much higher mean K for the broken bedrock due to one very high value (of about 7 388 m/d) obtained in the valley floor at GO3, which may be representative of the interface between the 389 overburden glacio-fluvial deposits and the bedrock. They also provided values for the deep massive 390 bedrock that are similar to those in the overlying fissured (shallow) bedrock.

391 Kmax, Kmin and the anisotropy angles (angle of Kmax to the horizontal plane, in °) in the 2D 392 vertical plane of the studied transect were further derived based on the measured orientation of the 393 dominant, hydraulically active fracture sets (see previous Section); a primary set ~70° dipping SSE 394 and secondary set ~75° dipping NNW, and another set ~35° dipping NNW which is not clearly 395 hydraulically active at large depths, but is probably active in the weathered near-surface. The 396 anisotropy ratio Kmin/Kmax (unitless) was assumed to decrease with depth from 0.5 to 0.1 (see 397 Section Methods). The calculated anisotropy angle increased with depth from a relatively low angle 398 in the broken bedrock as a result of the equal contribution of the three fracture sets (75° NW) to a 399 dominant steep angle in the deep bedrock (70° SE, parallel to dominant hydraulically active Variscan-400 Alpine fabrics). Fig. 5 plots the values of *Kmin*, *Kmax* along with the previous equivalent isotropic K 401 estimates (Moe et al. 2010; Comte et al. 2012). The values of Kmax were close to the values 402 obtained from hydraulic tests, which were assumed to preferentially mobilise groundwater from the 403 fractures responsible for Kmax (direction of maximum anisotropy). 404

405 Storage properties

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407 Quantitative information on the heterogeneity of aquifer storage properties was provided by the 408 joint analysis of hydraulic test results and geophysical data (MRS and ERT). Hydraulic test results are 409 limited to single estimates of specific yield (Sy). Moreover, Sy values obtained from measurements in 410 pumped wells in absence of observation wells are usually inaccurate due to possible well bore 411 storage effects. However they did provide indicative orders of magnitude as a guide, with broken 412 bedrock displaying Sy values around 3-5 %, fissured bedrock of about 0.1 % and massive bedrock 413 around 0.001 %, i.e. over an order of magnitude decrease between each layer with increasing depth. 414 Geophysical data provided higher resolution spatial information on the variability of storage 415 properties. MRS, despite an instrumental sensitivity that currently does not reliably quantify less 416 than 1 % water content, detected appreciable quantities of water above 2% in the central and low 417 part of the transect (Fig. 6). The highest water contents were detected at MRS8 in the valley floor 418 with values between 2 and 6 % at shallow depths (0-25 m) corresponding to the occurrence of the 419 relatively porous overburden materials and the broken bedrock. Similar water content values were 420 also obtained at MRS4, upslope of GO2, at depths (2-15 m) consistent with the area of deep 421 weathering identified by ERT (X=400-600 m on Fig. 4) and characterised by significant deepening of 422 the broken bedrock layer. In the upper part of the transect around GO1, no appreciable water 423 content was detected by MRS suggesting water content < 1 %. This is also the case for the two 424 soundings carried out between GO2 and the valley floor. Overall, the MRS results show relatively 425 good agreement with the depth delineation of the broken bedrock and overburden layers from ERT 426 (Fig. 6 and 7a,b). For the overburden in the valley floor, MRS water content values of approx. 5-6 % 427 were obtained while the broken bedrock (transition zone) yielded values of 2-4 %. At depth below 428 the base of the broken layer (as delineated with ERT), MRS mostly provided either water content 429 values below the normal limit of detection of the methods (1 % water content) or values with high 430 inversion uncertainty. This suggests that the fissured and massive bedrock had water content values 431 < 1 %. Note that the broken bedrock, which is likely present in the upper part of the transect was 432 mostly unsaturated in the summer (dry) conditions when both ERT and MRS surveys were 433 undertaken.

434 The Waxman and Smits model (Fig. 7c), provided spatial variations of (effective) porosities 435 consistent with MRS results, and porosity values of about half the MRS water contents. Values 436 ranging from 3-7 % were obtained in the overburden/broken bedrock of the valley floor while the 437 broken layer of the mid and upper part of the transect displayed values ranging from 1 to 3 %, 438 decreasing with distance to valley floor. For the fissured layer, we obtained values between 1 % 439 beneath the valley floor to less than 0.1 % in the upper part of the transect, with a relatively 440 progressive decrease from valley bottom to hilltop. Porosities lower than 0.03 % were obtained in 441 the massive, unweathered bedrock layer. As a comparison, Archie's model (not presented here),

which is applicable in clay-free materials, provided much higher, unrealistic porosity values of about
an order of magnitude (i.e. about ten times) higher, ranging 1 to 40 %. This confirmed that Archie's
model is not applicable in this type of environment due to significant clay content.

The slightly lower (effective) porosities values obtained from the Waxman and Smits model as compared to the MRS water content values, of approximately a factor of two, are consistent with the findings from Vouillamoz *et al.* (2012, 2014) who reported similar difference factors in low porosity (< 3 %) clay-rich bedrock aquifers. This supports the assumption, used for subsequent modelling of residence times, that bedrock effective porosities may reasonably be assumed as equal to Waxman and Smits porosities.

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452 Aquifer recharge

454 Application of the water table fluctuation using *Sy* values derived from MRS provided recharge 455 values that vary along the hillslope (Table 4), from an average of 163 mm/y in the hillslope to an 456 average of 287 mm/y in the valley bottom, which correspond to about 10% and 20% or the rainfall, 457 respectively. These values are higher than previous values reported by Cai & Ofterdinger (2016) due 458 to estimated *Sy* values higher than the literature values used by previous authors.

459 460

461 Data integration and numerical groundwater modelling

462463 *Methods*

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465 A 2D numerical groundwater model (equivalent porous medium EPM) was constructed in line 466 with the ERT transect in order to assess the influence of the structural heterogeneity, i.e. the spatial 467 variations in flow (permeability, anisotropy) and storage (porosity) properties on groundwater flow 468 paths and residence time. EPM models are commonly used for studying groundwater dynamics in 469 fractured bedrock aquifers at catchment and hillslope-scale (Ball et al. 2014; Welch & Allen 2014; 470 Kolbe et al. 2016). EPM models use integrated hydraulic properties data (hydraulic conductivities, 471 storativities) from hydraulic test solutions and anisotropy analysis (which are also EPM models). 472 They are also a lot less computationally expensive than discrete fracture network (DFN) models for 473 which the current hillslope-scale knowledge on the actual distribution and properties (length, 474 aperture, roughness) of fracture networks is insufficient for DFN model application.

We used the finite element code FEFLOW v6.2. The model domain was 2D vertical (cross section; Fig. 8) with a length slightly shorter than the ERT transect (1200 m; the ERT region south of the river was not included). The ground surface was obtained from the DEM and therefore identical to the ERT profile. The maximum elevation (NW of transect) was 183 m and the minimum river elevation (SW of transect) was 31.5 m, i.e. the elevation of the river. The base of the model was fixed at -50 m relative to mean sea level. The mesh comprised 5366 triangular elements and 2974 nodes.

481 The model domain was structured in five layers: alluvium (only present in the floodplain), broken 482 bedrock, fissured bedrock, massive bedrock, and a hypothetical very low conductivity/porosity 483 bottom layer (Figure 8), within which homogeneous hydrogeological parameters were applied. Two 484 alternative numerical models were applied using different levels of hydrogeological knowledge of 485 the site (aquifer structure, hydraulic properties, and recharge). Model 1 used the information 486 obtained from boreholes only. It considered a uniform layered structure defined from lateral 487 interpolation of the hydrogeological units as identified in boreholes and initially interpreted by Moe 488 et al. (2010) prior to extensive deployment of geophysical surveys. This model represents the 489 common, most simple catchment/regional scale conceptualisation of basement aquifers (Fig. 8a). 490 Model 2 incorporated the additional knowledge obtained from the interpretation of geophysical 491 surveys (Fig. 8b). Comparing the two models aimed at illustrating the added value provided by the 492 geophysical data. Details of the set up of both models were as follows.

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494 Model 1 considered a tabular spatial distribution of the different aquifer units that mimics the 495 topography (Figure 8a). Recharge values were based on borehole-based estimates by Cai & 496 Ofterdinger (2016). Model hydraulic conductivities and porosities were assigned from layer-specific 497 hydraulic test results from Moe et al. (2010) and Comte et al. (2012), respectively (Table 5). Effective 498 porosities, used for residence time calculations, were assumed to be equal to the specific yield 499 values (Sy) from borehole data. As Model 1 initially provided a poor fit to observed groundwater 500 heads, a recalibration had to be performed through increasing recharge (x5) and decreasing 501 hydraulic conductivities (/5).

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503 Model 2 was based on the geometry of the different aquifer units from the interpretation of the 504 geophysical data (ERT and MRS). Recharge values were the new values calculated with the water 505 table fluctuation method using the specific yield data derived from the MRS surveys (see Section 506 Aquifer characterisation). Hydraulic conductivities K (Table 5) were applied as anisotropic and 507 assigned from joint interpretation of revised hydraulic test data Comte et al. (2012), geophysical 508 data and fracture analyses (see Section Aquifer characterisation). Effective porosities (Table 5) were 509 assigned using the porosity values obtained from joint application of ERT and MRS (Figure 7 and 510 Table 5).

511

512 Both models were assigned values of longitudinal and transversal dispersivity of 30 and 3 m 513 respectively, as typically reported for fractured aquifers for the scale of this study (hundreds of m to 514 km-scale flow paths; Neuman 2005; Schulze-Makuch 2005; Zhou et al. 2007). With regard to the 515 boundary conditions other than the recharge described above, both models also computed the top 516 surface as a seepage face, with recharge (entering flux) switching automatically to discharge (exiting 517 flux) during simulations when the water table was equal or higher than the top surface. Additionally, 518 the river was applied as a constant head of 31.5 m, which is the average observed river level. In 519 terms of groundwater ages, water entering from recharge was assigned a constant age of 0 years. 520 Models were run in transient flow and age conditions with initial heads equal to the topographic 521 elevation and initial ages of 0 years, until reaching a dynamic steady state. Flow and groundwater 522 age were solved simultaneously. Groundwater age simulations in FEFLOW are treated as dissolved 523 transport and account for the applied values of effective porosities and dispersivities. This, contrary 524 to isochrone calculations on flow paths (streamlines), allows for water dispersion and mixing of 525 groundwater ages. Flow path simulations were run on simulated flow fields to highlight average flow 526 paths from recharge (top surface) to discharge points (river and seepage areas).

527

528 Numerical modelling results

530 Models evaluation

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532 The goodness-of-fit of the different models, i.e. the discrepancy between predicted and observed 533 heads was assessed through root-mean-square error (RMSE) calculations. Model 1, which computed 534 previously published values of hydraulic properties (Moe et al. 2010) and recharge (Cai & 535 Ofterdinger 2016) initially failed to reproduce hydraulic heads that fit the observed head in the 536 boreholes. Calculated head values were significantly lower than observed, of up to 125 m difference 537 at GO1 (uphill), due to either underestimation of the recharge or overestimation of the aquifer 538 hydraulic conductivities, but close to observations downhill due to the constraint by the river fixed 539 head. Across the whole transect this model produced a high total RMSE of 71.6 m. The recalibration 540 of Model 1, through increasing the recharge and decreasing K both by a factor five, produced a 541 reasonable fit with a RMSE of 3.7 m. Models 2 directly reproduced reasonably well the hydraulic 542 heads observed in the boreholes GO1, GO2 and GO3 with a RMSE of 4.0 m. For Model 2 however, 543 the simulated heads in the deep GO1 and GO2 boreholes were somewhat higher than those

- observed. Because these two boreholes are within or close to deep weathered/fractured zones (as
- identified by geophysics), a 3D model allowing deep and lateral drainage would be expected to
 correct for this mismatch by lowering heads in the deep units. The model comparison demonstrates
- 547 that both Model 1 (final recalibrated version) and Model 2 equally honour the observed heads,
- 548 however only Model 2 also honours the observed aquifer properties, structure and recharge values.
- 549 Simulation of groundwater fluxes, flow paths and residence time distributions 550

551 For both Model 1 and Model 2, Darcy's fluxes were highest in the broken bedrock and decreased 552 with depth (Fig. 9a,d). The deep weathered/fractured zones identified by geophysics and computed 553 in Model 2 allowed for thicker areas of high flux (Fig. 9d), especially in the vicinity of these zones (X 554 300-600 m; 750-850 m; 110-1300 m). In terms of budget (Fig. 10), in the case of Model 1 almost 80 % of the flow rate in the aquifer, originating from the recharge, transited through the broken 555 556 bedrock layer (transition zone) and fissured layer (shallow zone), with over 60% in the broken 557 bedrock only. The proportion was lower for Model 2, with about 70% flowing through broken and 558 fissured bedrock and about 50% through the broken bedrock only. This suggests that overall 559 hydrogeological heterogeneity favours deeper groundwater flow in the fissured and massive 560 bedrock (< 20 % for Model 1 vs. about 25 % for Model 2).

561 This is further confirmed by the average flow path (streamlines) simulations (Fig. 9b,e). In the 562 case of Model 1, most flow paths were sub-horizontal and restricted to the shallow broken layer (Fig. 563 9b). A limited number of flowlines travel through the fissured and massive layers. For Model 2, 564 which incorporated lateral variations in the thickness of the aquifer layers, flow paths were more 565 evenly distributed with depth, with higher contributions from the fissured and massive layers (Fig. 566 9e). They were also characterised by undulations as a result of lateral variations of weathering 567 thicknesses and the anisotropy of conductivity with an increased vertical component to the 568 groundwater flow. In detail, Model 2 simulations showed sub-vertical or oblique groundwater flow 569 from recharge locations where weathering/fracturing is poorly developed. In locations with 570 extensive weathering/fracturing, groundwater directions changed upwards suggesting that these 571 locations may act as drainage structures at catchment and regional scale. These model regions were 572 also associated with groundwater discharge through seepage, which agree with the spring lines 573 locations observed in the field (Fig. 4). Seepage appeared to occur specifically in areas where the two 574 following conditions are met: (1) upwelling groundwater flow associated to decrease in thickness of 575 the broken bedrock on the downgradient side of deep weathered/fractured 'channels' and (2) 576 presence of topographic low also associated to less competent bedrock in these zones. In contrast 577 the absence of deep weathering zones in Model 1 resulted in higher and more evenly spread 578 seepage along the hillslope.

579 Groundwater age simulations for the two models (Fig. 9c,f and Fig. 11) provided ages increasing 580 both laterally from hill top to the valley and with depth. Youngest groundwater was obtained in GO1T and GO3SS and oldest in GO3S, GO3D and GO2D (Fig. 11). Model 1 and Model 2 both resulted 581 582 in simulated ages less than about 10 years and 50 years, respectively, in boreholes and seepages 583 areas. When compared to Model 1, which did not incorporate lateral heterogeneity, Models 2 584 showed greater age mixing with depth whereas the former produced steeper age gradient with 585 depth in the upper aquifer. Model 2 ages are consistent with independent Tritium data (Pilatova 2013) that showed (1) Tritium concentrations in boreholes within the range 2.5-17.6 TU reflecting a 586 587 mixture of sub-modern water (< 1TU; prior 1952) and modern water (5-15+ TU, i.e. < 5 to 10 years) 588 (Figure 11), (2) decreasing Tritium values with depth reflecting larger component of older waters 589 with depth, (3) highest Tritium values (>14 TU) in the upper bedrock at GO1T well (X=~100 m) and 590 GO2S well (X=~600 m) reflecting high proportion of young water associated to recharge as well as 591 descending to sub-horizontal groundwater flow and (4) lowest Tritium values (2.5 TU) encountered 592 at the deep GO3D well (X=~1200 m) reflecting larger component of older upwelling groundwater. 593 The distribution of residence times obtained for Model 2, the best parameterised model

accounting for the geophysical heterogeneity, is also broadly consistent with previous catchment- or

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595 hillslope-scale studies in fractured hard rock aquifers. Robins & Smedley (1994) reported modern 596 groundwater tritium ages in the fractured basement of Jersey. Jaunat et al. (2012) reported CFC-SF6 597 residence times lower than 50 years in weathered/fractured gneiss of the French Basque Country for 598 similar flow path lengths. Banks et al. (2009) reported CFC ages of less than 40 years in a hillslope 599 transect underlain by weathered/fractured metasediments. Lapworth et al. (2013) reported CFC-600 SF6-3H mean residence times of 32-65 years in deeply weathered catchments of the West African 601 basement. Kolbe et al. (2016) modelled mean transit times of 40 years in a granite-gneiss catchment 602 of ~10 km-long with similar mean flow path lengths of ~300 m. The modelling results also 603 corroborate recent findings by Ameli et al. (2016) who showed the major impact of subsurface 604 heterogeneity on groundwater residence time distribution in a well-studied hillslope transect in 605 glacial till in Sweden.

606 607

608 General discussion

610 Study implications for groundwater resilience to climate change and contaminants, and 611 catchment management approaches

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613 The results indicate that overall low and depth-decreasing porosities, together with high 614 hydraulic gradients, give relatively short residence times for groundwater from recharge to seepage 615 or river discharge. In the upper bedrock (broken bedrock/transition zone; 1 to 30 m thick), where 616 over 50% of groundwater flow occurs, models results suggest groundwater flow paths of between 617 10-100 m long and groundwater ages of < 1 year. This implies that groundwater in this aquifer unit is 618 sensitive to both weather seasonality and extreme events (winter/summer recharge condition and 619 drought/floods). This groundwater mostly contributes to diffuse seepage, which is then collected by 620 agricultural drains. As such, point and diffuse contaminants in groundwater are expected to affect 621 surface water quickly, on timescales of days to months. In the deeper bedrock (fissured/shallow and 622 massive/deep bedrock; depths higher than 10-50 m), model results yield longer flow paths (100-623 1000 m) and older groundwater ages (from one year to several decades). This implies higher 624 groundwater resilience to extreme weather conditions, seasonality and incidental contaminant 625 exposures, but not to long-term (multi-decadal) climate change and persistent contamination, such 626 as (i) changes in recharge due to long-term changes in rainfall and evapotranspiration and (ii) diffuse 627 (e.g. agricultural nutrients) contamination. As these deeper bedrock units mostly contribute to river 628 flow as well as the most significant seepages areas/drains, the surface water network at base flow 629 (i.e. when mostly supported by groundwater) is also expected to be more vulnerable to long-term 630 climate change and contamination. 631 Model comparison (Model 1 vs. Model 2) further highlights the importance of adequately 632 accounting for aquifer heterogeneity when using models to predict the response of 633 weathered/fractured rock catchments to climate and land use change as well as contaminations. 634 Using information from borehole observations only, which does not allow for adequately capturing 635 spatial variations in weathering/fracturing, leads to underestimation of the contribution of deep 636 aquifer units to catchment water balance and discharge to river/surface water bodies. This also

leads to underestimating groundwater residence times and exaggerating both ground and surface
 water sensitivity to climate variability and contaminations. In contrast, better accounting for aquifer

639 heterogeneity as revealed here by geophysical surveys, results in higher resilience of groundwater

resources to climate variability and surface contaminations. This is of importance when applyinggroundwater models with the aim of accurately informing short to long-term catchment

642 management and policy.

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646 Study limitations and implications for residence time distributions

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648 The hydrogeophysical and numerical modelling works conducted in this study reveal the major 649 role played by hydrogeological heterogeneities on groundwater flowpaths and transit time 650 distributions at 2D hillslope scale providing important insights into catchment scale groundwater 651 processes in weathered/fractured aquifers. The methodology however has some limitations with 652 respect to accurately representing 3D heterogeneities and their influence on flow and transit times 653 at smaller (i.e. borehole) or larger (i.e. catchment or region). Specifically, the hillslope 654 characterisation and modelling work conducted do not represent; (1) three-dimensional 655 groundwater flow such as lateral or deep drainage due to deep weathering structures possibly 656 significant at the catchment/regional scale; (2) individual fracture networks and associated fracture 657 flow processes; (3) accurate 2D/3D clay distribution and cementation factor; (4) uncertainty in 658 estimates of hydrogeological porosity from geophysical (ERT and MRS) porosity; (5) temporal 659 variability in recharge; (6) flow processes in the unsaturated zone above the water table.

660 With regards to the two-dimensionality of the models, it may be expected that some 661 groundwater flow at the catchment scale may take place laterally to the 2D transects especially (i) in deep weathering furrows such as these detected by geophysics at about 300-600m, 750-850m and 662 663 110-1300m; (ii) in more local preferential flow paths along the NE-SW fractured systems. Not 664 accounting for these 3D processes, the current modelling approach is likely to lead to underestimate 665 deep groundwater flow as well as mixing and resulting groundwater ages, and overestimate seepage 666 (along the hillslope). Full 3D geophysical (acquisition and inversion) and numerical modelling 667 approaches are recommended in basement catchments where strong 3D heterogeneity is expected. 668 The expected increased accuracy is however at the cost of much higher requirement in terms of 669 acquisition and modelling time and resource, including computational.

Increased accuracy in modelling results may also be obtained by implementing discrete fractured
 network (DFN) modelling approaches. The structural and geophysical data may be used to support
 the computation of fracture orientation and density, which would allow direct computation of
 fracture permeability and aperture. A DFN approach requires implementation of computationally
 expensive 3D models.

Application of ERT to derive porosities requires information on spatial variations of bedrock clay
 content and clay mineralogy. Direct and high resolution 3D characterisation of clay properties
 through sampling/coring is challenging, but such resolution may be achieved indirectly through use
 of alternative geophysical methods such as the induced polarisation (IP).

In line with previous works in similar settings, we have assumed that aquifer effective porosity
was half the MRS water content and equal to ERT porosity. Should this assumption be erroneous,
effective porosity values higher than the ERT total porosity would result in higher groundwater
residence times. More research is recommended to constrain the relationship between ERT/MRS
porosity and hydrogeological porosities (storativity and effective porosity).

The temporal variability in recharge has not been accounted for and would be required to better understanding seasonality in groundwater contribution to seepage, drains and stream. It may, in addition, enhance groundwater mixing and, therefore groundwater residence times. Similarly, neglecting flow processes in the unsaturated zone primarily result in underestimating groundwater mixing and ages in areas where the unsaturated zone is thicker.

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691 Conclusion

The study highlights the strong control of geological heterogeneities on groundwater flow and residence times in metamorphic rock catchments in temperate regions. It also demonstrates the high value of surface geophysical data and fracture and clay analysis for the parametrisation of numerical groundwater models in complex aquifers.

697 The investigations have revealed a high degree of heterogeneity in the distribution of 698 hydrogeological properties at the hillslope scale. Through 2D equivalent porous media modelling 699 using FEFLOW, it has been shown that the high spatial variation of hydraulic conductivity and 700 porosity results in deeper groundwater flow paths (with an increased in the vertical flow component 701 with depth) as compared to simple layered conceptual models based on borehole data only. They 702 also result in older groundwater ages through enhanced mixing and dispersion caused by 703 heterogeneities and anisotropies of hydraulic conductivity. Groundwater ages along the hillslope 704 were simulated to be of modern ages, i.e. less than 50 years, consistent with available Tritium data. 705 The integrated approach presented, using both borehole and surface geophysical surveys, is

shown to help parametrise numerical groundwater models that honour the observed data without
requiring significant parameter recalibration. Such robustly parameterised models offer
straightforward application in catchment water management, to investigate in detail the
contribution of groundwater to the catchment hydrological function, as well as the impact of climate
change and contaminants on groundwater.

The results presented here suggest that the uppermost weathered/broken part of the aquifer, which is a major contributor to hillslope discharge, is sensitive to extreme hydrological events and seasonal climate fluctuations as well as point and diffuse contamination. The deeper, less weathered part of the aquifer is a significant contributor to river flow and major hillslope seepage areas and would be more sensitive to long-term (decadal) climate fluctuations and persistent, diffuse contaminants.

717 It is suggested that the 2D steady-state hillslope approach conducted provides only a minimum 718 estimate of groundwater ages at the catchment scale. This is due to underestimation of 719 groundwater mixing favoured by possible 3D structures and seasonal recharge. Full 3D 720 characterisation and modelling approaches, although more expensive in terms of data and 721 computational resources, are required to further improve our understanding of groundwater flow 722 and residence times. Recommendations include: (1) implementation of 3D MRS and ERT surveys and 723 inversion, including 3D characterisation of clay mineralogy through alternative geophysical methods 724 such as IP; (2) resolution and computation of fracture networks and fracture flow; (3) application of 725 transient, variably saturated model for better accounting of mixing and processes in the unsaturated 726 zone; (4) more research on the relationship between geophysical (ERT and MRS) porosity and 727 hydrogeological porosity.

728 729

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Tables and captions 989

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991 Table 1. Summary and source of Waxman and Smits model parameters used to derive porosity values 992 for the different aquifer units

	Bulk resistivity [*]	Water	Water electrical	Cementation	Cation exchange
	[ohm.m]	temperature [†]	conductivity [†]	factor [‡]	capacity§
		[°C]	[S/m]	[unitless]	[meq/100g]
Overburden	< 500	12.1 - 16.0	0.027 - 0.045	2.5	2.0-6.0 [¶]
Broken zone	500 – 2500	12.1 - 16.0	0.027 - 0.045	2.5	2.0-6.0
Fissured zone	300 - 1000	12.8 – 14.2	0.029 - 0.050	2	5.2 – 18.7
Massive zone	1000 - 10 000	12.7 – 13.1	0.033 – 0.057	1.5	1.2 – 1.7

* 2D distribution obtained from ERT (see Results section).

+ average values (summer 2009) recorded in boreholes (low temporal variability).

‡ from Hartmann and Beaumont (1999).

§ values calculated from mineralogical analysis and natural gamma log (see Results section).

¶ same values as for the broken zone due to lack of in situ data.

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Table 2. Clay compositions in the different boreholes and resulting bedrock clay weight fraction and 995 996 cation exchange capacity (CEC)

	Muscovite [*] % vs tot. clay	Chlorite [*] % vs tot. clay	Illite [*] % vs tot. clay	Montm. [*] % vs tot. clay	Nat. γ [†] cps	Clay weight fraction‡ %	Total CEC [‡] meg/100g
G01				, ,			<u> </u>
Broken zone (transition)	58	27	7	8	116	51	6.0
Fissured zone (shallow)	40	20	22	18	111	51	11.7
Massive zone (deep)	66	34	0	0	102	43	1.7
GO2							
Broken zone (transition)	75	11	8	5	101	40	3.2
Fissured zone (shallow)	18	3	47	32	102	48	18.7
Massive zone (deep)	82	18	0	0	116	47	1.2
GO3							
Broken zone (transition)	77	14	5	4	78	30	2.0
Fissured zone (shallow)	28	5	40	27	41	16	5.2
Massive zone (deep)	82	18	0	0	116	47	1.2

* dominant clay minerals that affects natural gamma logging and rock bulk CEC. CEC (muscovite)~1 meq/g; CEC (illite)~10 meq/g; CEC (illite)~20 meq/g; CEC (montmorillonite)~90 meq/g (multiple sources, see Methods section). +average natural gamma count per seconds from borehole logging.

‡ Clay weight fraction and CEC calculated from individual clay CEC and natural gamma according to Revil et al. (1998).

1001 Table 3. Summary of isotropic hydraulic conductivities obtained from pumping and recovery test 1002 interpretation and comparison with previous results

Hydrogeological unit	Thickness from ERT [m]	Mean <i>K</i> [m/d]	K range [m/d]	Mean <i>K</i> from previous works [*] [m/d]	K range from previous works [*] [m/d]
Overburden	10 (in valley floor)	8.7	-	9	-
Broken bedrock (transition z.)	15-60	1.3×10 ⁻¹	1×10 ⁻¹ – 2×10 ⁻¹	1	7×10 ⁻² – 7
Fissured bedrock (shallow z.)	5-40	1.0×10 ⁻²	3×10 ⁻³ – 4×10 ⁻²	1 10 ⁻²	1×10 ⁻³ – 4×10 ⁻²
Massive bedrock (deep z.)	>40	6.0×10 ⁻³	4×10 ⁻³ – 8×10 ⁻³	1 10 ⁻²	7×10 ⁻³ – 6×10 ⁻²

* initial hydraulic testing after drilling (Moe et al. 2010) using Horslev (infiltration tests) and Theis-Jacob (pumping tests)

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 Table 4. Recharge values recalculated using the water table fluctuation methods and the specific
 yield values derived from MRS. Borehole name suffixes SS (subsoil), T (transition) are derived from 1006

1007 the codes listed in section Hydrogeological setting

	Rainfall [*] [mm/y]	Water table cumulated rise [*] [m]	Specific yield [†] [%]	Calculated recharge [mm/y]	Calculated recharge [% of rainfall]
Hillslope					
GO1T 2010-2011	1134	13.6	0.75	102	9
GO1T 2011-2012	1433	19.7	0.75	147	11
GO2T 2010-2011	1134	8.8	1.57	138	13
GO2T 2011-2012	1433	16.8	1.57	264	19
Average				163	13
Valley floor					
GO3SS 2010-2011	1134	7.4	3	222	21
GO3SS 2011-2012	1433	11.7	3	351	24
Average				287	23

* from Cai & Ofterdinger (2016)

⁺ from MRS, assuming specific yield = 0.5*MRS water content

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	Kmax [*]	Kmin/Kmax	Anisotropy	Porosity [†]	Recharge [‡]
	[m/d]	[unitless]	angle [°]	[%]	[mm/y]
Model 1					
Overburden	9	1	Isotropic	20	Valley floor: 95 (480)§
Broken zone	1 (0.2) [§]	1	Isotropic	4	Hillslope: 75 (370)§
Fissured zone	0.01 (0.002) [§]	1	Isotropic	0.2	NA
Massive zone	0.01 (0.002) [§]	1	Isotropic	0.01	NA
Model 2					
Overburden	8.6	1	Isotropic	7	Valley floor: 285
Broken zone	0.14	0.5	75 ° dip NW	3	Hillslope: 165
Fissured zone	0.014	0.3	85 ° dip SE	0.1-0.5	NA
Massive zone	0.008	0.1	65 ° dip SE	0.01	NA
**			(0010) (

* from Moe et al. (2010) for Model 1; from Comte et al. (2012) for Model 2

+ from Comte et al. (2012) for Model 1; from Waxman and Smits' model and MRS results for Model 2 (Figure 7)

‡ from Cai & Ofterdinger (2016) for Models 1; recalculated values for Model 2 (Table 4)

§ values in brackets are final values after model calibration to observed heads in boreholes (K/5 and Recharge*5)

1011 **Figure captions** 1012 1013 Fig. 1. Study site physical setting maps; (a) site location within the Irish basement geological 1014 framework (modified from Geological Survey of Ireland 2006); (b) catchment boundary with location 1015 of hydrological monitoring infrastructures; (c) local interpretative geological map with location of the 1016 borehole clusters, the ERT profile and the MRS soundings.(a) and (b) modified from Comte et al. 1017 (2012). 1018 1019 **Fig. 2**. Generic hydrogeological conceptual model of weathered/fractured rocks aquifers in the 1020 context of the Irish terminology (modified from Comte et al. 2012). 1021 1022 Fig. 3. Fracture pole density distributions (Schmidt net lower hemisphere projection) and orientations 1023 (fracture azimuth rose diagrams) from outcrops and boreholes (acoustic televiewer probe); blue 1024 points and arrows show the hydraulically active fractures unambiguously identified in boreholes 1025 (Nitsche 2014; modified from Comte et al. 2012). 1026 1027 Fig. 4. ERT results (a) and interpreted conceptual model of the weathered/fractured aquifer (b). 1028 1029 Fig. 5. Comparison of hydraulic conductivities obtained from; (a) initial isotropic hydraulic test 1030 interpretations from Moe et al. (2010) used in Model 1; (b) refined isotropic interpretation from 1031 Comte et al. (2012); and (c-e) anisotropic K values used in Model 2, (c) K_{max}, (d) K_{equivalent}, and (e) K_{min}. 1032 Fig. 6. Vertical distribution of MRS water content for the 8 MRS sounding (see locations on Figure 1) 1033 1034 and comparison with the aquifer conceptual model units delineated from ERT (Figure 4). 1035 1036 Fig. 7. Spatial variations of storage properties derived from ERT and MRS geophysical data; (a) ERT 1037 resistivity model with location of approximate volume of investigation of the MRS soundings; (b) MRS 1038 water content logs; (c) ERT porosity calculated from Waxman & Smits' model. Hatched areas indicate 1039 the unsaturated zone for which saturated Archie and Waxman & Smits models used are not 1040 applicable. 1041 1042 Fig. 8. Conceptual aquifer geometries implemented in the numerical models: (a) generic tabular 1043 structure of weathered/fractured layers from borehole interpretation (Model 1); (b) complex layered 1044 structure derived from geophysical data reconciled with borehole logs (Model 2). 1: Overburden 1045 (alluvial and glacial sediments); 2: Broken bedrock (transition zone); 3: Fissured (shallow) bedrock; 4: 1046 Massive (deep) bedrock; 5: Substratum (very low productivity). 1047 1048 Fig. 9. Simulation results for Model 1 (a,b,c) and Model 2 (d,e,f) showing groundwater seepage rates 1049 at the model surface, Darcy's fluxes variations across the transect (**a**,**d**), groundwater mean 1050 flowpaths (**b**,**e**), and groundwater ages (**c**,**f**). 1051 1052 Fig. 10. Relative distributions of groundwater flow rate (as % of total flow) in the four conceptual 1053 aquifer units (overburden, broken, fissured and massive bedrock) for the two models considered. The 1054 massive bedrock here also includes the deeper levels shown in Figure 8. 1055 1056 Fig. 11. Modelled groundwater ages in the different boreholes for the two model cases. The hatched 1057 area indicates groundwater ages that are older than the modern period of high atmospheric Tritium 1058 levels, and inconsistent with measured Tritium concentrations in boreholes samples.

- 1 Catchment-scale heterogeneity of flow and storage properties in a
- 2 weathered/fractured hard rock aquifer from resistivity and magnetic
- 3 resonance surveys: Implications for groundwater flow paths and residence
- 4 times distributions
- 5 6

7

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15 Running Title

- 16 Flow and storage properties of fractured rocks
- 17

18 Abstract

- 19 Groundwater pathways and residence times are controlled by aquifer flow and storage properties,
- 20 which are characterised by high spatial heterogeneity in weathered/fractured hard rock aquifers.
- 21 Building on earlier work in a metamorphic aquifer in NW Ireland, new clay mineralogy and
- 22 geophysical data analyses provided high spatial resolution constraints on the variations of aquifer
- 23 properties. Groundwater storage values derived from magnetic resonance sounding and electrical
- resistivity tomography were found to largely vary laterally and with depth, by orders of magnitude.
- 25 Subsequent implementation of numerical, 2D-hillslope groundwater models showed that
- 26 incorporating heterogeneity from geophysical data in model parametrisation led to best fit to
- 27 observations as compared to a reference model based on borehole data only. Model simulations
- 28 further revealed that; 1/strong spatial heterogeneity produces deeper, longer groundwater flow
- 29 paths and higher age mixing in agreement with the mixed sub-modern/modern ages (mostly <50
- 30 years) provided by independent tritium data; 2/areas with extensive weathering/fracturing are
- 31 correlated with seepage zones of older groundwater, due to changes in the flow directions, and are
- 32 likely to act as drainage structures for younger groundwater on a catchment or regional scale.
- 33 Implications for groundwater resilience to climate extremes and surface pollution are discussed along
- 34 with recommendations for further research.

35 Weathered/fractured hard rock aquifers underlie over 20% of the global land surface (Sharp

36 2014) and are characterised by a high degree of structural heterogeneity and overall low

37 productivity. In recent years, water managers and policy-makers have moved to adopt a catchment

38 scale approach to the integrated management of surface and subsurface water resources (EU Water

39 Framework Directive 2000/60/EC), including the UK (UKTAG 2011) and Ireland (Daly *et al.* 2016).

- 40 Understanding spatial variations of aquifer hydraulic properties at the catchment-scale, which
- 41 dictates groundwater flow pathways and residence times, is crucial to inform catchment
- 42 management plans. Yet, resolving such spatial heterogeneity in fractured bedrock remains very
- challenging due to the typically scattered nature of direct observations points (boreholes and
 outcrops) that usually do not have sufficient spatial coverage to capture the scale of heterogeneity
- 45 (De Marsily *et al.* 2005; Neuman 2005).

46 To address the lack of spatial resolution in heterogeneous fractured rock catchments, traditional 47 direct testing techniques in boreholes, such as hydraulic testing and geophysical logging, are 48 increasingly combined with indirect and more spatially integrative investigation methods, including 49 tracer testing (e.g. Klepikova 2016), geophysical imaging (ground- or airborne-based) (Comte et al. 50 2012; Shakas et al. 2016; Day-Lewis et al. 2017) and remote sensing (Cassidy et al. 2014; Frances et 51 al. 2014). The use of geophysics has long proven effective in resolving, at catchment scale, the 52 heterogeneity of fractured rock aquifers (Holbrook et al. 2014; Robinson et al. 2016). The Electrical 53 Resistivity Tomography (ERT) method is known to be efficient at imaging spatial variability in 54 weathering, geological heterogeneity and fracture patterns (e.g. Chandra et al. 2010; Rainer et al. 55 2007). All ERT studies, however, stress the importance of *a priori* information, especially borehole 56 data and outcrop mapping to support its hydrogeological interpretation (Skinner & Heinson 2004; 57 Comte *et al.* 2012).

58 A number of studies have demonstrated the benefits of using ERT in combination with other 59 geophysical methods; in particularly with the magnetic resonance sounding (MRS) that complements 60 imaging of heterogeneity by ERT with lower resolution but more quantitative information on water 61 storage. Both methods have, for instance, been used to map groundwater occurrence and develop 62 hydrogeological conceptual and numerical models in weathered basement aquifers (Frances et al. 63 2014; Baltassat et al. 2005) and monitor groundwater recharge (Descloitres et al. 2008). As yet, 64 however, most of these studies have focused on low latitude regions with deep and relatively water-65 productive weathering horizons (saprolite) that produce strong MRS and ERT responses for relatively 66 simple, layered aquifer geometries. There are fewer examples in higher latitude catchments with a 67 glacial legacy, such as in Ireland, where most of the saprolite (relatively high storage, porous layer) is 68 absent, exposing only the fractured (low storage) and structurally complex bedrock. In addition, 69 these geophysical approaches still remain either (i) not systematically applied in catchment 70 groundwater studies or (ii) applied qualitatively, i.e. used to inform aquifer heterogeneity conceptual 71 models rather than to quantify spatial variations in aquifer properties (permeability and porosity).

72 This needs further consideration in highly heterogeneous basement rock catchments.

73 In Ireland, the fractured rock aquifers provide a good analogue of temperate region bedrock 74 aquifers with a glacial legacy. The island of Ireland is underlain by over 60% of hydrogeologically 75 poorly-productive fractured bedrock (Moe et al. 2010), either cropping out directly or covered by 76 superficial glacio-fluvial and/or alluvial sediments. Most of this poorly-productive bedrock is 77 composed of various grades of metamorphic (basement) rocks; from low-grade metasediments to 78 high-grade gneiss-migmatites and granitoids. Groundwater in these rocks, despite their overall low 79 productivity, is nonetheless crucial for maintaining river base flow during dry periods and supporting 80 aquatic ecosystems and small-scale rural water supply (DCCAE 2017). Over the last decade, the 81 extension of the Irish national groundwater monitoring network to poorly productive basement 82 aquifers as part of implementing the European Union's Water Framework Directive (EPA 2006; Moe 83 et al. 2010) has stimulated hydrogeological research through Irish Government-funded projects. 84 Among them, the Griffith Poorly-productive Aquifers Project (2007-2014), on which this work is 85 based, aimed to improve the understanding of groundwater flow regimes in fractured rock aquifers

86 and the contribution of groundwater to catchment water balance (Comte et al. 2012; Cassidy et al. 87 2014; Caulfield et al. 2014; Cai & Ofterdinger 2016).

This paper presents an overview of the latest research in a micaschist catchment in Co. Donegal, 88 89 Republic of Ireland, with the aim of resolving catchment-scale spatial variations of aquifer 90 properties. It synthesises previously published research on aquifer typology and well-scale hydraulics 91 (Comte et al. 2012; Cassidy et al. 2014), bedrock weathering (Caulfield et al. 2014) and aquifer 92 storage properties (Legchenko et al. 2017), augmented with the latest results from quantitative 93 interpretation of geophysical data (ERT and MRS) to provide a robust spatial understanding of flow 94 and storage property fields. This work further explores the integration of the existing knowledge on 95 the heterogeneity of aquifer properties using numerical models to assess groundwater flow paths 96 and residence time distributions. Finally, we use the results to discuss the impact of climate change 97 and contaminant transport on groundwater resources and catchment management, as well as 98 further recommendations for improved groundwater modelling. 99

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101 Hydrogeological setting

103 Geology

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The Gortinlieve catchment (5 km²), Co. Donegal, NW Ireland (Figure 1) is underlain by Late 105 106 Precambrian micaschists and psammites of intermediate metamorphic grade (low amphibolite 107 facies). These belong to the Grampian terrane in Co. Donegal as part of the Southern Highland 108 Group, spanning through Ireland and Scotland. They originate from turbidite sequences deposited c. 109 550 Ma BP (McConnell & Long 1997; Caulfield et al. 2014) and subsequently subjected to poly-phase deformation and metamorphism during the Caledonian orogeny. The Caledonian tectonic regime 110 111 associated with the closure of the lapetus Ocean (Grampian phase, early Ordovician) generated the 112 current regional NE–SW oriented structures (Chew 2009) characterised by fault scarps and 113 cartographic/topographic lineaments visible in the upper catchment (Fig. 1). The later Taconic phase 114 (late Ordovician) generated the current WNW-ESE orientation deformational structures and 115 imprinted the retrograde amphibolite metamorphic facies (McConnell & Long 1997). The current, 116 Alpine, strike-slip tectonic regime, is reflected by reactivation of NE-SW fracture orientations and 117 further creation of a general NW-SE trend (Worthington & Walsh 2011; Cooper et al. 2012). Localised kaolinite rich Tertiary lateritic horizons preserved on Grampian rocks of western Ireland 118 119 (Legg et al. 1985) provide evidence that the basement was during this time at least partially 120 exhumed and undergoing tropical weathering. Subsequent Quaternary glaciations have resulted in 121 further erosion, including the removal of the upper levels of the weathering profiles, most 122 pronounced in high topographic areas, and in the deposition of heterogeneous glacio-fluvial material 123 (clay-till and sand and gravel) in low-lying areas and valley bottoms.

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125 Hydrogeology and borehole instrumentation

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127 Annual rainfall in the area ranges from 1000 to 1200 mm and mean temperature annually ranges 128 from 6 to 14 °C (Met Eireann 2017). The catchment comprises a headwater stream network of a 129 Carrigans River, a tributary of the River Foyle that discharges into the Atlantic Ocean, approx. 40 km 130 NE of the catchment (Caulfield et al. 2014). The Gortinlieve catchment was instrumented with monitoring boreholes by the Irish Environmental Protection Agency (EPA) in 2006 as part of the 131 132 national groundwater monitoring programme. Three borehole clusters were sited in a linear 133 hillslope transect at high (GO1, 174 m above mean sea level; hereafter noted m amsl), intermediate 134 (GO2, 88 m amsl) and low (GO3, 33 m amsl) elevations within the catchment. Individual clusters 135 contain up to 4 boreholes, each isolated and screened across different depth-distinctive zones 136 commonly encountered in Irish bedrock aguifers. The initial classification of Moe et al. (2010)

137 conceptually described these respective intervals as; subsoil SS (average interval 1–3 m below

- ground surface (bgs); only present in Gortinlieve at the valley bottom), transition zone T (average
- interval 4–5 m bgs), shallow bedrock S (average interval 8–19 m bgs) and deep bedrock D (average
- interval 30–67 m bgs). The conceptual model was further refined by Comte *et al.* (2012) in order to
 help reconcile geological features with geophysical constraints (Fig. 2): overburden deposits (cf.
- subsoil); broken bedrock (cf. transition zone); fissured bedrock (cf. shallow bedrock); massive
- 143 bedrock (cf. deep bedrock). The Irish EPA monitors water levels at 15-minute intervals in each
- borehole using automatic data loggers. The site was also equipped with an automated tipping
- 145 bucket rain gauge (AEG 100) since October 2010 for the duration of the project.
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148 Aquifer characterisation methods

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150 Structural, mineralogical and hydraulic investigations

Detailed analysis of fracture patterns and clay occurrence in the bedrock was carried out in order to establish the micro- to meso-scale structural controls on groundwater flow, and to provide constraints on larger (meso- to macro) scale hydraulic and geophysical interpretations and numerical modelling.

156 Fracture orientations were measured on maps, outcrops, boreholes and guarry exposures. 157 Regional structural trends were determined from interpretation of the geological map (Smith 1991; 158 Long et al. 1992; McConnell & Long 1997) and the 20-m resolution digital elevation model (Ordnance 159 Survey of Ireland). During field mapping, the main fracture parameters measured were strike, dip 160 magnitude, dip azimuth with a minimum of 30 fracture measurements for each sample site, 161 subsequently plotted as rose diagram and Schmidt net pole density distribution. The acquisition and plotting methodology are described in details in Comte et al. (2012) and Nitsche (2014). Borehole 162 163 fractures were analysed using acoustic televiewer logs. Among these, hydraulically active fractures 164 were identified through cross-correlation with electrical conductivity logs whereby marked changes 165 in water electrical conductivity at the depth of observed fractures was interpreted as fracture flow 166 (Nitsche 2014).

Mineralogical and petrographic data are summarised by Caulfield et al. (2014). Representative 167 field outcrop and recovered borehole core samples were further characterised in this study to 168 quantify the relative proportions of identified phyllosilicate minerals to assess their influence on 169 170 geoelectrical properties. The mineral modes of the dominant basement lithologies (psammitic 171 schists uphill vs. micaschists downhill) were quantified by petrographic point-counting (500 points 172 per slide). These compositions were taken to represent fresh, primary (unweathered) bedrock. Free 173 phyllosilicate minerals (mineral grains not bound in the competent fresh bedrock structure) were 174 separated and collected from weathered samples by repeated washing with distilled water and dried 175 at 50°C. Clay sized (<2 µm) fractions were obtained following the method outlined in Caulfield et al. 176 (2014), and references therein. The relative proportions of primary (muscovite, chlorite) and 177 secondary (illite, montmorillonite and illite/montmorillonite admixtures) were determined via 178 thermogravimetric analysis (TGA) using a Netzsch libra thermogravimetric analyser, in conjunction 179 with XRD and FTIR results from Caulfield et al. (2014). Samples were heated from 200-880°C to 180 determine sequential dehydroxylation water loss from the clay mineral fraction. Mineral proportions 181 (volume %) were converted to weight %. Using the method of Revil et al. (1998), the total cation 182 exchange capacity (CEC) of the different geological units (transition, shallow, deep bedrock) at each of the three sites (GO1-3) were calculated from (i) the relative mass proportion of clay minerals, (ii) 183 184 their individual CEC, obtained from literature (Swineford 1955; Carroll 1959; Wiklander 1964; 185 Thomas 1976; Ridge 1983; Revil et al. 1998; Crain 2000; Gillespie et al. 2001; Ellis & Singer 2007; 186 Henn et al. 2007) and (iii) the total clay content derived from natural gamma ray logs from each 187 borehole.

188 Aquifer pumping and recovery tests were implemented in every borehole to provide local values 189 of equivalent hydraulic conductivity and storativity for each borehole. The hydraulic testing 190 methodology is described in details in Comte et al. (2012). Pumping test were conducted at a 191 constant rate (2 to 30 L/min depending on the borehole) and both pumping and recovery curves 192 were jointly interpreted using AQTESOLV Pro v4.5 and applying a range of adapted analytical 193 solutions (with regards to the known aquifer structure and the borehole technical characteristics) 194 including single porosity/permeability models. Hydraulic conductivities (K) were calculated from 195 transmissivity values using observed unit aquifer thicknesses from borehole data, ERT data and 196 geophysical logging data.

K in fractured rocks is typically anisotropic, i.e. directionally dependent implying different K 197 198 values in different directions, with the maximum hydraulic conductivity Kmax following the direction 199 of the hydraulic active fractures and the minimum hydraulic conductivity *Kmin* being orthogonal to 200 the hydraulic active fractures. This further implies that the anisotropy angle in fractured rocks, 201 because of the presence of steep fractures, has a higher vertical component than in most 202 sedimentary rocks where the direction of Kmax is often sub-horizontal, parallel to the sedimentary 203 bedding (in this case Kmax is usually denoted Kh) and the direction Kmin is subvertical (and in this 204 case denoted Kv). In this work, values of hydraulic conductivities provided by hydraulic tests were 205 further analysed in terms of anisotropy in 2D (along the studied transect). As hydraulic test values 206 were assumed equivalent to isotropic hydraulic conductivities, the integration of fracture analysis 207 data allowed for the definition of Kmax Kmin and anisotropy angles (by definition the angle of Kmax to the horizontal) based on the measured orientation of the hydraulically active fractures. The 208 209 anisotropy ratio Kmin/Kmax (unitless) was derived by considering it to decrease with depth from 0.5 210 to 0.1. At depth, where the steep fracture sets dominated (see section Aquifer characterisation 211 results) and produced a stronger anisotropy, the ratio was set to 0.1, whereas at shallow depths, 212 where other fracture sets and pronounced weathering contribute to flow, particularly in the broken 213 bedrock, anisotropy was expected to weaken and was therefore set to 0.5. The anisotropy angle was 214 derived as an average of the individual angles of the dominant fracture sets for each bedrock 215 conceptual unit (broken, figured, massive).

- 217 *Geophysical investigations*
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19 Electrical resistivity tomography and petrophysical models for porosity estimation

221 A geophysical survey using electrical resistivity tomography ERT was carried out to provide a 222 catchment-scale conceptual understanding of the bedrock heterogeneity and its hydrogeological 223 significance. ERT was applied along the hillslope transect of the three well clusters using a Syscal Pro 224 72 resistivity meter. As described in Comte et al. (2012), the acquisition was carried out with 60 225 electrodes at 5-m unit electrode spacing, subsequently expanded through roll-along. The dipole-226 dipole (DD) and multi-gradient (mGD) quadripole configurations ran sequentially, ultimately allowing 227 a depth of investigation of 50–60 m (Edwards 1977). After noise removal, DD and mGD apparent 228 resistivity data were jointly inverted using RES2DINV v3.58. The inversion provided a 2D model of 229 specific resistivities in which model regions that are insensitive to the input data were removed 230 using the depth of investigation (DOI) method of Oldenburg & Li (1999). The best model retained for hydrogeological interpretation was obtained for the 5th iteration, giving an absolute error between 231 232 observed and calculated apparent resistivities of 9.6%.

In addition to the structural information provided by the interpretation of spatial resistivity variation, a petrophysical model was applied to the resistivity model in order to derive bedrock porosity. Due to the substantial amount of clay minerals (both primary and secondary) in the psammites and the micaschists, Archie's model (Archie 1942) was not applicable and instead, the Waxman and Smits (1976) model was used. The Waxman and Smits model relates the bulk (specific) resistivity obtained by inversion to the aquifer total porosity (which is the unknown to resolve), the

- pore water saturation (equal to one below the water table), the rock matrix cementation factor, the groundwater electrical conductivity and temperature, and in addition to Archie's Model, the total rock CEC, which was obtained from mineralogical investigations and natural gamma logs (see section *Aquifer characterisation results*). The model is described in Revil et al. (1998) and was applied to estimate the 2D (hillslope) distribution of total porosity along the transect imaged by ERT. Table 1 summarises the values of the parameters used for the different hydrogeological units.
- 245
- 246 Magnetic Resonance Sounding
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248 The magnetic resonance sounding (MRS) method (Legchenko 2013) was applied at 11 points 249 along the ERT transect and close to the borehole clusters (Fig. 1). The MRS surveys provided 1D 250 (vertical) profiles of water content. MRS measurements were performed in two surveys, in 2010 and 251 2016 using the NUMISPLUS and NUMISPOLY instruments, respectively, both developed by IRIS 252 Instruments (France). As described in more detail in Legchenko et al. (2017), figure-eight square 253 loops were applied in three configurations: two cable turns of 25-m-side, one turn of 37.5-m-side, 254 and one turn of 50-m-side, with a reference loop in 2016 to improve the signal to noise ratio, which 255 is typically low in fractured rock environments, due to a low water content. Eight out of the 11 MRS 256 soundings had acceptable signal-to-noise ratios; the remaining three are not considered in this work. 257 1D inversion of MRS data was carried out using the SAMOVAR software package. It used the 258 Tikhonov regularization method (Legchenko & Shushakov 1998) and the uncertainty in the inversion 259 results was examined applying Monte-Carlo simulations (Legchenko et al. 2017). Geological 260 interpretation of MRS results was performed taking into account the thicknesses of the aquifer units

- 261 262
- 263 Estimation of aquifer storage parametres from ERT porosity and MRS water content 264

(broken and fissured zone) identified with ERT as structural constraints.

265 ERT porosity and MRS water content were used to derive aquifer specific yield *Sy* and effective 266 porosity *n_e* subsequently used for recharge estimation, and numerical groundwater modelling, 267 respectively. By definition, *n_e* is the interconnected porosity controlling groundwater mass transport 268 and residence times. Also by definition, *Sy*, also called drainage porosity, is the mobile pore water 269 not retained by capillarity forces and controls water table fluctuations in unconfined aquifers.

In fractured rocks, where porosity is mostly the result of the opening of fractures and foliation planes by (groundwater-aided) weathering, the occurrence of closed (unconnected) pore space is limited (Singhal & Gupta 2010). This implies that total porosity, obtained from ERT, can be considered a close estimate of the effective porosity n_e , which was our assumption for subsequent modelling.

The MRS water content is commonly reported (Lubczynski & Roy 2004) as being correlated with, and representing an intermediary between, the drainage porosity *Sy* (mobile pore water non retained by capillarity forces) and the effective porosity n_e (immobile and mobile interconnected pore water). Recent studies carried out in low porosity fractured environments (Vouillamoz *et al.* 2012, 2014) have reported MRS water content values linearly correlated to, although higher than *Sy* (within a factor 2) estimated through pumping tests. Consequently, for subsequent recharge calculations we have taken *Sy* at half (0.5 times) the measured MRS water content.

- 282
- 283 Recharge estimations

Recharge values were calculated with the water table fluctuation method applicable to recharge
estimation in unconfined aquifers (Healy & Cook 2002). The method requires water table time-series
as well as values of the specific yield at depth where the water table fluctuates. We used the
piezometric fluctuations in the shallowest boreholes (GO1T, GO2T and GO3SS) as reported in Cai &
Ofterdinger (2016) for the period 2010-2012. For the specific yield, we used the MRS derived values

- obtained at the approximated depth of the water table (see previous Section) from the MRS
 soundings nearest to the boreholes; MRS1 and MRS2 for GO1T, MRS4 and MRS5 for GO2T, and
 MRS8 for GO3SS.
- 293
- 294 Aquifer characterisation results
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296 Aquifer structure and weathering

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298 Fracture analysis

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300 Measurements of hydraulically active fracture from both outcrops and boreholes at the site 301 showed dominant meso-scale (millimetre to metre) fracture occurrences and orientations which 302 correlate with established past tectonic regimes in Ireland (Fig. 3). Two dominant hydraulically active 303 fracture sets were identified: a dominant set with fractures oriented E-W to NE-SW with a dip angle 304 of about 60-80° S to SE; and a less frequent set with fractures also oriented E-W to NE-SW but with a 305 dip angle of about 60-90° N to NW. The dominant fracture set may be attributed to the strike-slip 306 regional fracturing associated with Alpine compression during the Palaeocene, whereas the second 307 set may be attributed to the compressive fabric of the Neo-Proterozoic to late Carboniferous 308 orogenies (Grampian, Caledonian and Variscan), which also characterise the regional-scale 309 lineaments (Fig. 1) and Lower Carboniferous normal faults (Worthington & Walsh 2011), reactivated 310 by the Alpine strike-slip regime (Cooper et al. 2012). The most frequently recorded fracture set is 311 oriented WNW–ESE with a low dip angle of 35 ° NNE corresponding to the highly weathered 312 Dalradian schistosity plane (see below). However, this set does not appear to be clearly associated 313 with hydraulic activity in boreholes and therefore may only play a role in the uppermost levels of the 314 aquifer (broken bedrock and possibly the upper part of the fissured bedrock) for which hydraulic 315 activity in boreholes is difficult to record due to their shallow depths and denser

316 weathering/pervasive fracturing.

At macro-scale (catchment and regional scale, i.e. 100 m to km; Fig. 1b), cartographic lineaments attributed to glacial erosion preferentially affecting regional geological boundaries and deep weathering corridors were found to be typically associated with the measured pre-Alpine NE–SW trend (Comte *et al.* 2012).

- 321
- 322 Mineral analysis

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324 Mineralogical investigations reveal that phyllosilicate minerals are dominated by muscovite and 325 chlorite (primary) and illite and montmorillonite (secondary clays produced by weathering of chlorite 326 and muscovite) (Caulfield et al. 2014). Major non-clay minerals comprise quartz and feldspar. 327 Relative proportions of clay minerals versus the total clay mass fraction are presented in Table 2. 328 The massive (unweathered) bedrock encountered in the deep boreholes was found to contain 329 negligible secondary clays, with only primary chlorite and muscovite observed. The fissured zone in 330 boreholes GO2 and GO3 contained an approx. equal mixture of illite and montmorillonite with lesser 331 amounts of chlorite and muscovite, and in GO1, relatively even quantities of the four minerals. This

suggests a moderate weathering of the fissured zone primary clays in GO1, whilst in GO2 and GO3,
 they have undergone major weathering as evidenced by the almost complete absence of chlorite.

334 Over 80% of clays in the broken zone were muscovite and the remaining 20% approx. equal

335 proportions of illite and montmorillonite. Because weathering is most active in the broken zone, it

336 was concluded that secondary clays are transported in suspension downward from the broken zone

and redeposited in the underlying fissured zone (Comte *et al.* 2012; Caulfield *et al.* 2014).

Table 2 also details the total clay mass fraction as derived from natural gamma ray logs. The
 relative proportions of clay minerals and the total clay mass fraction allowed calculation of the total
 CEC of the different aquifer units using the model of Revil *et al.* (1998). The total clay content of the

- 341 different aquifer units (broken, fissured and massive bedrock) was not significantly different;
- however due to significant differences in clay mineralogy, the resulting CEC is different. The fissure
- 343 bedrock, with a higher proportion of illite and montmorillonite (secondary weathering clays
- 344 characterised by high CECs) consistently produced the highest bulk CEC values. The massive bedrock
- dominated by muscovite and chlorite (primary clays characterised by low CEC values) produced the
- lowest bulk CEC. This CEC distribution is relevant to the interpretation of ERT results and applicationof the petrophysical models in which it is a key input parameter.
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349 Geophysical structure

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351 Results of the ERT investigations provide cross-sectional variations of electrical resistivity along 352 the borehole hillslope transect (Figure 4), revealing a high degree of aquifer heterogeneity. As 353 electrical resistivity is primarily controlled by porosity (fracture porosity in the present context), clay 354 content and mineralogy, its spatial variations can be interpreted as reflecting changes in lithology, 355 pore water saturation, open fracture density and weathering intensity. Resistivity values range from 356 about 100 ohm.m to just under 10 000 ohm.m, i.e. three orders of magnitude. Highest resistivities 357 (>> 1000 ohm.m) were observed at the base of the transect and reflect the massive micaschist unit 358 beneath the weathered front, characterised by a low density closed fractures and small quantities of 359 secondary weathering clays. Some equally high resistivity values were also obtained in the very 360 shallow subsurface in the higher elevation part of the profile. Correlations with the depth of the 361 water table in borehole cluster GO1 and observed spring lines at its base suggested that this 362 uppermost, high resistivity unit corresponds to unsaturated, clay-poor psammite-schist. The lowest 363 resistivity values (<500 ohm.m) were obtained at a relatively shallow depth in the flood plain and 364 corresponded to the clay-till and alluvium overburden. Intermediate resistivity values (500-2500 365 ohm.m) were observed throughout the whole transect at shallow to intermediate depths (0 to about 366 50 m below surface) and characterised the heterogeneous weathered/fractured schists (broken and 367 fissured bedrock). The relatively continuous unit with values ranging ~300-1000 ohm.m overlying the 368 massive bedrock are interpreted as fissured bedrock characterised by a high density of open 369 fractures possibly filled with high-CEC weathering clays as suggested by the clay mineralogy. This 370 fissured layer was particularly deep at three locations along the transect; at the following horizontal distances from NW end of the profile of X=400-500 m, 750-850 m and 1100-1300 m. Since they 371 372 correlate well with the regional Grampian lineaments (Fig. 1), these locations are interpreted as 373 deep fractured zones that facilitate deep weathering. At these three locations a slightly more 374 resistant layer was observed overlying the fissured layer, with resistivities in the range of 500-2500 375 ohm.m. This unit is interpreted as the broken schists (transition zone) where the weathered clays 376 have been leached and transported to the underlying fissured zone. 377

378 Flow properties

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380 Pumping and recovery tests in every borehole provided estimates of (equivalent isotropic) 381 hydraulic conductivities K for bedrock and overburden (Table 3). A clear trend of decreasing hydraulic conductivity with depth is observed. The broken bedrock (transition zone) displayed mean 382 values close to 10⁻¹ m/d, the fissured bedrock close to 10⁻² m/d with large variability of over two 383 orders of magnitude and the massive (deep) bedrock around 6×10^{-3} m/d. The overburden (glacio-384 385 fluvial and alluvium deposits) present in the valley floor displayed much higher mean K values of ~9 386 m/d. These values partly agree with values from previous hydraulic tests by Moe et al. (2010). These 387 authors obtained much higher mean K for the broken bedrock due to one very high value (of about 7 388 m/d) obtained in the valley floor at GO3, which may be representative of the interface between the 389 overburden glacio-fluvial deposits and the bedrock. They also provided values for the deep massive 390 bedrock that are similar to those in the overlying fissured (shallow) bedrock.

391 Kmax, Kmin and the anisotropy angles (angle of Kmax to the horizontal plane, in °) in the 2D 392 vertical plane of the studied transect were further derived based on the measured orientation of the 393 dominant, hydraulically active fracture sets (see previous Section); a primary set ~70° dipping SSE 394 and secondary set ~75° dipping NNW, and another set ~35° dipping NNW which is not clearly 395 hydraulically active at large depths, but is probably active in the weathered near-surface. The 396 anisotropy ratio Kmin/Kmax (unitless) was assumed to decrease with depth from 0.5 to 0.1 (see 397 Section *Methods*). The calculated anisotropy angle increased with depth from a relatively low angle 398 in the broken bedrock as a result of the equal contribution of the three fracture sets (75° NW) to a 399 dominant steep angle in the deep bedrock (70° SE, parallel to dominant hydraulically active Variscan-400 Alpine fabrics). Fig. 5 plots the values of *Kmin*, *Kmax* along with the previous equivalent isotropic K 401 estimates (Moe et al. 2010; Comte et al. 2012). The values of Kmax were close to the values 402 obtained from hydraulic tests, which were assumed to preferentially mobilise groundwater from the 403 fractures responsible for Kmax (direction of maximum anisotropy).

405 Storage properties

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404

Quantitative information on the heterogeneity of aquifer storage properties was provided by the joint analysis of hydraulic test results and geophysical data (MRS and ERT). Hydraulic test results are limited to single estimates of specific yield (*Sy*). Moreover, *Sy* values obtained from measurements in pumped wells in absence of observation wells are usually inaccurate due to possible well bore storage effects. However they did provide indicative orders of magnitude as a guide, with broken bedrock displaying *Sy* values around 3-5 %, fissured bedrock of about 0.1 % and massive bedrock around 0.001 %, i.e. over an order of magnitude decrease between each layer with increasing depth.

414 Geophysical data provided higher resolution spatial information on the variability of storage 415 properties. MRS, despite an instrumental sensitivity that currently does not reliably quantify less 416 than 1 % water content, detected appreciable quantities of water above 2% in the central and low 417 part of the transect (Fig. 6). The highest water contents were detected at MRS8 in the valley floor 418 with values between 2 and 6 % at shallow depths (0-25 m) corresponding to the occurrence of the 419 relatively porous overburden materials and the broken bedrock. Similar water content values were 420 also obtained at MRS4, upslope of GO2, at depths (2-15 m) consistent with the area of deep 421 weathering identified by ERT (X=400-600 m on Fig. 4) and characterised by significant deepening of 422 the broken bedrock layer. In the upper part of the transect around GO1, no appreciable water 423 content was detected by MRS suggesting water content < 1 %. This is also the case for the two 424 soundings carried out between GO2 and the valley floor. Overall, the MRS results show relatively 425 good agreement with the depth delineation of the broken bedrock and overburden layers from ERT 426 (Fig. 6 and 7a,b). For the overburden in the valley floor, MRS water content values of approx. 5-6 % 427 were obtained while the broken bedrock (transition zone) yielded values of 2-4 %. At depth below 428 the base of the broken layer (as delineated with ERT), MRS mostly provided either water content 429 values below the normal limit of detection of the methods (1 % water content) or values with high 430 inversion uncertainty. This suggests that the fissured and massive bedrock had water content values 431 < 1 %. Note that the broken bedrock, which is likely present in the upper part of the transect was 432 mostly unsaturated in the summer (dry) conditions when both ERT and MRS surveys were 433 undertaken.

434 The Waxman and Smits model (Fig. 7c), provided spatial variations of (effective) porosities 435 consistent with MRS results, and porosity values of about half the MRS water contents. Values 436 ranging from 3-7 % were obtained in the overburden/broken bedrock of the valley floor while the 437 broken layer of the mid and upper part of the transect displayed values ranging from 1 to 3 %, 438 decreasing with distance to valley floor. For the fissured layer, we obtained values between 1 % 439 beneath the valley floor to less than 0.1 % in the upper part of the transect, with a relatively 440 progressive decrease from valley bottom to hilltop. Porosities lower than 0.03 % were obtained in 441 the massive, unweathered bedrock layer. As a comparison, Archie's model (not presented here),

which is applicable in clay-free materials, provided much higher, unrealistic porosity values of about
an order of magnitude (i.e. about ten times) higher, ranging 1 to 40 %. This confirmed that Archie's
model is not applicable in this type of environment due to significant clay content.

The slightly lower (effective) porosities values obtained from the Waxman and Smits model as compared to the MRS water content values, of approximately a factor of two, are consistent with the findings from Vouillamoz *et al.* (2012, 2014) who reported similar difference factors in low porosity (< 3 %) clay-rich bedrock aquifers. This supports the assumption, used for subsequent modelling of residence times, that bedrock effective porosities may reasonably be assumed as equal to Waxman and Smits porosities.

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452 Aquifer recharge

454 Application of the water table fluctuation using *Sy* values derived from MRS provided recharge 455 values that vary along the hillslope (Table 4), from an average of 163 mm/y in the hillslope to an 456 average of 287 mm/y in the valley bottom, which correspond to about 10% and 20% or the rainfall, 457 respectively. These values are higher than previous values reported by Cai & Ofterdinger (2016) due 458 to estimated *Sy* values higher than the literature values used by previous authors.

459 460

461 Data integration and numerical groundwater modelling

462463 *Methods*

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465 A 2D numerical groundwater model (equivalent porous medium EPM) was constructed in line 466 with the ERT transect in order to assess the influence of the structural heterogeneity, i.e. the spatial 467 variations in flow (permeability, anisotropy) and storage (porosity) properties on groundwater flow 468 paths and residence time. EPM models are commonly used for studying groundwater dynamics in 469 fractured bedrock aquifers at catchment and hillslope-scale (Ball et al. 2014; Welch & Allen 2014; 470 Kolbe et al. 2016). EPM models use integrated hydraulic properties data (hydraulic conductivities, 471 storativities) from hydraulic test solutions and anisotropy analysis (which are also EPM models). 472 They are also a lot less computationally expensive than discrete fracture network (DFN) models for 473 which the current hillslope-scale knowledge on the actual distribution and properties (length, 474 aperture, roughness) of fracture networks is insufficient for DFN model application.

We used the finite element code FEFLOW v6.2. The model domain was 2D vertical (cross section; Fig. 8) with a length slightly shorter than the ERT transect (1200 m; the ERT region south of the river was not included). The ground surface was obtained from the DEM and therefore identical to the ERT profile. The maximum elevation (NW of transect) was 183 m and the minimum river elevation (SW of transect) was 31.5 m, i.e. the elevation of the river. The base of the model was fixed at -50 m relative to mean sea level. The mesh comprised 5366 triangular elements and 2974 nodes.

481 The model domain was structured in five layers: alluvium (only present in the floodplain), broken 482 bedrock, fissured bedrock, massive bedrock, and a hypothetical very low conductivity/porosity 483 bottom layer (Figure 8), within which homogeneous hydrogeological parameters were applied. Two 484 alternative numerical models were applied using different levels of hydrogeological knowledge of 485 the site (aquifer structure, hydraulic properties, and recharge). Model 1 used the information 486 obtained from boreholes only. It considered a uniform layered structure defined from lateral 487 interpolation of the hydrogeological units as identified in boreholes and initially interpreted by Moe 488 et al. (2010) prior to extensive deployment of geophysical surveys. This model represents the 489 common, most simple catchment/regional scale conceptualisation of basement aquifers (Fig. 8a). 490 Model 2 incorporated the additional knowledge obtained from the interpretation of geophysical 491 surveys (Fig. 8b). Comparing the two models aimed at illustrating the added value provided by the 492 geophysical data. Details of the set up of both models were as follows.

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494 Model 1 considered a tabular spatial distribution of the different aquifer units that mimics the 495 topography (Figure 8a). Recharge values were based on borehole-based estimates by Cai & 496 Ofterdinger (2016). Model hydraulic conductivities and porosities were assigned from layer-specific 497 hydraulic test results from Moe et al. (2010) and Comte et al. (2012), respectively (Table 5). Effective 498 porosities, used for residence time calculations, were assumed to be equal to the specific yield 499 values (Sy) from borehole data. As Model 1 initially provided a poor fit to observed groundwater 500 heads, a recalibration had to be performed through increasing recharge (x5) and decreasing 501 hydraulic conductivities (/5).

502

503 Model 2 was based on the geometry of the different aquifer units from the interpretation of the 504 geophysical data (ERT and MRS). Recharge values were the new values calculated with the water 505 table fluctuation method using the specific yield data derived from the MRS surveys (see Section 506 Aquifer characterisation). Hydraulic conductivities K (Table 5) were applied as anisotropic and 507 assigned from joint interpretation of revised hydraulic test data Comte et al. (2012), geophysical 508 data and fracture analyses (see Section Aquifer characterisation). Effective porosities (Table 5) were 509 assigned using the porosity values obtained from joint application of ERT and MRS (Figure 7 and 510 Table 5).

511

512 Both models were assigned values of longitudinal and transversal dispersivity of 30 and 3 m 513 respectively, as typically reported for fractured aguifers for the scale of this study (hundreds of m to 514 km-scale flow paths; Neuman 2005; Schulze-Makuch 2005; Zhou et al. 2007). With regard to the 515 boundary conditions other than the recharge described above, both models also computed the top 516 surface as a seepage face, with recharge (entering flux) switching automatically to discharge (exiting 517 flux) during simulations when the water table was equal or higher than the top surface. Additionally, 518 the river was applied as a constant head of 31.5 m, which is the average observed river level. In 519 terms of groundwater ages, water entering from recharge was assigned a constant age of 0 years. 520 Models were run in transient flow and age conditions with initial heads equal to the topographic 521 elevation and initial ages of 0 years, until reaching a dynamic steady state. Flow and groundwater 522 age were solved simultaneously. Groundwater age simulations in FEFLOW are treated as dissolved 523 transport and account for the applied values of effective porosities and dispersivities. This, contrary 524 to isochrone calculations on flow paths (streamlines), allows for water dispersion and mixing of 525 groundwater ages. Flow path simulations were run on simulated flow fields to highlight average flow 526 paths from recharge (top surface) to discharge points (river and seepage areas). 527

- 528 Numerical modelling results
- 529

530 Models evaluation

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532 The goodness-of-fit of the different models, i.e. the discrepancy between predicted and observed 533 heads was assessed through root-mean-square error (RMSE) calculations. Model 1, which computed 534 previously published values of hydraulic properties (Moe et al. 2010) and recharge (Cai & 535 Ofterdinger 2016) initially failed to reproduce hydraulic heads that fit the observed head in the 536 boreholes. Calculated head values were significantly lower than observed, of up to 125 m difference 537 at GO1 (uphill), due to either underestimation of the recharge or overestimation of the aquifer 538 hydraulic conductivities, but close to observations downhill due to the constraint by the river fixed 539 head. Across the whole transect this model produced a high total RMSE of 71.6 m. The recalibration 540 of Model 1, through increasing the recharge and decreasing K both by a factor five, produced a 541 reasonable fit with a RMSE of 3.7 m. Models 2 directly reproduced reasonably well the hydraulic 542 heads observed in the boreholes GO1, GO2 and GO3 with a RMSE of 4.0 m. For Model 2 however, 543 the simulated heads in the deep GO1 and GO2 boreholes were somewhat higher than those

- observed. Because these two boreholes are within or close to deep weathered/fractured zones (as
 identified by geophysics), a 3D model allowing deep and lateral drainage would be expected to
- 546 correct for this mismatch by lowering heads in the deep units. The model comparison demonstrates
- 547 that both Model 1 (final recalibrated version) and Model 2 equally honour the observed heads,
- 548 however only Model 2 also honours the observed aquifer properties, structure and recharge values.
- 549 Simulation of groundwater fluxes, flow paths and residence time distributions 550

551 For both Model 1 and Model 2, Darcy's fluxes were highest in the broken bedrock and decreased 552 with depth (Fig. 9a,d). The deep weathered/fractured zones identified by geophysics and computed 553 in Model 2 allowed for thicker areas of high flux (Fig. 9d), especially in the vicinity of these zones (X 554 300-600 m; 750-850 m; 110-1300 m). In terms of budget (Fig. 10), in the case of Model 1 almost 80 % of the flow rate in the aquifer, originating from the recharge, transited through the broken 555 556 bedrock layer (transition zone) and fissured layer (shallow zone), with over 60% in the broken 557 bedrock only. The proportion was lower for Model 2, with about 70% flowing through broken and 558 fissured bedrock and about 50% through the broken bedrock only. This suggests that overall 559 hydrogeological heterogeneity favours deeper groundwater flow in the fissured and massive 560 bedrock (< 20 % for Model 1 vs. about 25 % for Model 2).

561 This is further confirmed by the average flow path (streamlines) simulations (Fig. 9b,e). In the 562 case of Model 1, most flow paths were sub-horizontal and restricted to the shallow broken layer (Fig. 563 9b). A limited number of flowlines travel through the fissured and massive layers. For Model 2, 564 which incorporated lateral variations in the thickness of the aquifer layers, flow paths were more 565 evenly distributed with depth, with higher contributions from the fissured and massive layers (Fig. 566 9e). They were also characterised by undulations as a result of lateral variations of weathering 567 thicknesses and the anisotropy of conductivity with an increased vertical component to the 568 groundwater flow. In detail, Model 2 simulations showed sub-vertical or oblique groundwater flow 569 from recharge locations where weathering/fracturing is poorly developed. In locations with 570 extensive weathering/fracturing, groundwater directions changed upwards suggesting that these 571 locations may act as drainage structures at catchment and regional scale. These model regions were 572 also associated with groundwater discharge through seepage, which agree with the spring lines 573 locations observed in the field (Fig. 4). Seepage appeared to occur specifically in areas where the two 574 following conditions are met: (1) upwelling groundwater flow associated to decrease in thickness of 575 the broken bedrock on the downgradient side of deep weathered/fractured 'channels' and (2) 576 presence of topographic low also associated to less competent bedrock in these zones. In contrast 577 the absence of deep weathering zones in Model 1 resulted in higher and more evenly spread 578 seepage along the hillslope.

579 Groundwater age simulations for the two models (Fig. 9c,f and Fig. 11) provided ages increasing 580 both laterally from hill top to the valley and with depth. Youngest groundwater was obtained in GO1T and GO3SS and oldest in GO3S, GO3D and GO2D (Fig. 11). Model 1 and Model 2 both resulted 581 582 in simulated ages less than about 10 years and 50 years, respectively, in boreholes and seepages 583 areas. When compared to Model 1, which did not incorporate lateral heterogeneity, Models 2 584 showed greater age mixing with depth whereas the former produced steeper age gradient with 585 depth in the upper aquifer. Model 2 ages are consistent with independent Tritium data (Pilatova 2013) that showed (1) Tritium concentrations in boreholes within the range 2.5-17.6 TU reflecting a 586 587 mixture of sub-modern water (< 1TU; prior 1952) and modern water (5-15+ TU, i.e. < 5 to 10 years) 588 (Figure 11), (2) decreasing Tritium values with depth reflecting larger component of older waters 589 with depth, (3) highest Tritium values (>14 TU) in the upper bedrock at GO1T well (X=~100 m) and 590 GO2S well (X=~600 m) reflecting high proportion of young water associated to recharge as well as 591 descending to sub-horizontal groundwater flow and (4) lowest Tritium values (2.5 TU) encountered 592 at the deep GO3D well (X=~1200 m) reflecting larger component of older upwelling groundwater. 593 The distribution of residence times obtained for Model 2, the best parameterised model

accounting for the geophysical heterogeneity, is also broadly consistent with previous catchment- or

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hillslope-scale studies in fractured hard rock aquifers. Robins & Smedley (1994) reported modern
groundwater tritium ages in the fractured basement of Jersey. Jaunat *et al.* (2012) reported CFC-SF6
residence times lower than 50 years in weathered/fractured gneiss of the French Basque Country for
similar flow path lengths. Banks *et al.* (2009) reported CFC ages of less than 40 years in a hillslope
transect underlain by weathered/fractured metasediments. Lapworth *et al.* (2013) reported CFC-

600 SF6-3H mean residence times of 32-65 years in deeply weathered catchments of the West African

601 basement. Kolbe et al. (2016) modelled mean transit times of 40 years in a granite-gneiss catchment

- of ~10 km-long with similar mean flow path lengths of ~300 m. The modelling results also
 corroborate recent findings by Ameli *et al.* (2016) who showed the major impact of subsurface
- heterogeneity on groundwater residence time distribution in a well-studied hillslope transect in
 glacial till in Sweden.
- 606 607

608 General discussion

610 Study implications for groundwater resilience to climate change and contaminants, and 611 catchment management approaches

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613 The results indicate that overall low and depth-decreasing porosities, together with high 614 hydraulic gradients, give relatively short residence times for groundwater from recharge to seepage 615 or river discharge. In the upper bedrock (broken bedrock/transition zone; 1 to 30 m thick), where 616 over 50% of groundwater flow occurs, models results suggest groundwater flow paths of between 617 10-100 m long and groundwater ages of < 1 year. This implies that groundwater in this aquifer unit is 618 sensitive to both weather seasonality and extreme events (winter/summer recharge condition and 619 drought/floods). This groundwater mostly contributes to diffuse seepage, which is then collected by 620 agricultural drains. As such, point and diffuse contaminants in groundwater are expected to affect 621 surface water quickly, on timescales of days to months. In the deeper bedrock (fissured/shallow and 622 massive/deep bedrock; depths higher than 10-50 m), model results yield longer flow paths (100-623 1000 m) and older groundwater ages (from one year to several decades). This implies higher 624 groundwater resilience to extreme weather conditions, seasonality and incidental contaminant 625 exposures, but not to long-term (multi-decadal) climate change and persistent contamination, such 626 as (i) changes in recharge due to long-term changes in rainfall and evapotranspiration and (ii) diffuse 627 (e.g. agricultural nutrients) contamination. As these deeper bedrock units mostly contribute to river 628 flow as well as the most significant seepages areas/drains, the surface water network at base flow 629 (i.e. when mostly supported by groundwater) is also expected to be more vulnerable to long-term 630 climate change and contamination. 631 Model comparison (Model 1 vs. Model 2) further highlights the importance of adequately 632 accounting for aquifer heterogeneity when using models to predict the response of 633 weathered/fractured rock catchments to climate and land use change as well as contaminations. 634 Using information from borehole observations only, which does not allow for adequately capturing 635 spatial variations in weathering/fracturing, leads to underestimation of the contribution of deep 636 aquifer units to catchment water balance and discharge to river/surface water bodies. This also

leads to underestimating groundwater residence times and exaggerating both ground and surface
water sensitivity to climate variability and contaminations. In contrast, better accounting for aquifer

- heterogeneity as revealed here by geophysical surveys, results in higher resilience of groundwater
- resources to climate variability and surface contaminations. This is of importance when applying
- 641 groundwater models with the aim of accurately informing short to long-term catchment
- 642 management and policy.
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646 Study limitations and implications for residence time distributions

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648 The hydrogeophysical and numerical modelling works conducted in this study reveal the major 649 role played by hydrogeological heterogeneities on groundwater flowpaths and transit time 650 distributions at 2D hillslope scale providing important insights into catchment scale groundwater 651 processes in weathered/fractured aquifers. The methodology however has some limitations with 652 respect to accurately representing 3D heterogeneities and their influence on flow and transit times 653 at smaller (i.e. borehole) or larger (i.e. catchment or region). Specifically, the hillslope 654 characterisation and modelling work conducted do not represent; (1) three-dimensional 655 groundwater flow such as lateral or deep drainage due to deep weathering structures possibly 656 significant at the catchment/regional scale; (2) individual fracture networks and associated fracture 657 flow processes; (3) accurate 2D/3D clay distribution and cementation factor; (4) uncertainty in 658 estimates of hydrogeological porosity from geophysical (ERT and MRS) porosity; (5) temporal 659 variability in recharge; (6) flow processes in the unsaturated zone above the water table.

660 With regards to the two-dimensionality of the models, it may be expected that some 661 groundwater flow at the catchment scale may take place laterally to the 2D transects especially (i) in deep weathering furrows such as these detected by geophysics at about 300-600m, 750-850m and 662 663 110-1300m; (ii) in more local preferential flow paths along the NE-SW fractured systems. Not 664 accounting for these 3D processes, the current modelling approach is likely to lead to underestimate 665 deep groundwater flow as well as mixing and resulting groundwater ages, and overestimate seepage 666 (along the hillslope). Full 3D geophysical (acquisition and inversion) and numerical modelling 667 approaches are recommended in basement catchments where strong 3D heterogeneity is expected. 668 The expected increased accuracy is however at the cost of much higher requirement in terms of 669 acquisition and modelling time and resource, including computational.

Increased accuracy in modelling results may also be obtained by implementing discrete fractured
 network (DFN) modelling approaches. The structural and geophysical data may be used to support
 the computation of fracture orientation and density, which would allow direct computation of
 fracture permeability and aperture. A DFN approach requires implementation of computationally
 expensive 3D models.

Application of ERT to derive porosities requires information on spatial variations of bedrock clay
 content and clay mineralogy. Direct and high resolution 3D characterisation of clay properties
 through sampling/coring is challenging, but such resolution may be achieved indirectly through use
 of alternative geophysical methods such as the induced polarisation (IP).

In line with previous works in similar settings, we have assumed that aquifer effective porosity
was half the MRS water content and equal to ERT porosity. Should this assumption be erroneous,
effective porosity values higher than the ERT total porosity would result in higher groundwater
residence times. More research is recommended to constrain the relationship between ERT/MRS
porosity and hydrogeological porosities (storativity and effective porosity).

The temporal variability in recharge has not been accounted for and would be required to better understanding seasonality in groundwater contribution to seepage, drains and stream. It may, in addition, enhance groundwater mixing and, therefore groundwater residence times. Similarly, neglecting flow processes in the unsaturated zone primarily result in underestimating groundwater mixing and ages in areas where the unsaturated zone is thicker.

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691 Conclusion

The study highlights the strong control of geological heterogeneities on groundwater flow and residence times in metamorphic rock catchments in temperate regions. It also demonstrates the high value of surface geophysical data and fracture and clay analysis for the parametrisation of numerical groundwater models in complex aquifers.

697 The investigations have revealed a high degree of heterogeneity in the distribution of 698 hydrogeological properties at the hillslope scale. Through 2D equivalent porous media modelling 699 using FEFLOW, it has been shown that the high spatial variation of hydraulic conductivity and 700 porosity results in deeper groundwater flow paths (with an increased in the vertical flow component 701 with depth) as compared to simple layered conceptual models based on borehole data only. They 702 also result in older groundwater ages through enhanced mixing and dispersion caused by 703 heterogeneities and anisotropies of hydraulic conductivity. Groundwater ages along the hillslope 704 were simulated to be of modern ages, i.e. less than 50 years, consistent with available Tritium data. 705 The integrated approach presented, using both borehole and surface geophysical surveys, is

shown to help parametrise numerical groundwater models that honour the observed data without
requiring significant parameter recalibration. Such robustly parameterised models offer
straightforward application in catchment water management, to investigate in detail the
contribution of groundwater to the catchment hydrological function, as well as the impact of climate
change and contaminants on groundwater.

The results presented here suggest that the uppermost weathered/broken part of the aquifer, which is a major contributor to hillslope discharge, is sensitive to extreme hydrological events and seasonal climate fluctuations as well as point and diffuse contamination. The deeper, less weathered part of the aquifer is a significant contributor to river flow and major hillslope seepage areas and would be more sensitive to long-term (decadal) climate fluctuations and persistent, diffuse contaminants.

717 It is suggested that the 2D steady-state hillslope approach conducted provides only a minimum 718 estimate of groundwater ages at the catchment scale. This is due to underestimation of 719 groundwater mixing favoured by possible 3D structures and seasonal recharge. Full 3D 720 characterisation and modelling approaches, although more expensive in terms of data and 721 computational resources, are required to further improve our understanding of groundwater flow 722 and residence times. Recommendations include: (1) implementation of 3D MRS and ERT surveys and 723 inversion, including 3D characterisation of clay mineralogy through alternative geophysical methods 724 such as IP; (2) resolution and computation of fracture networks and fracture flow; (3) application of 725 transient, variably saturated model for better accounting of mixing and processes in the unsaturated 726 zone; (4) more research on the relationship between geophysical (ERT and MRS) porosity and 727 hydrogeological porosity.

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Tables and captions 989

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991 Table 1. Summary and source of Waxman and Smits model parameters used to derive porosity values 992 for the different aquifer units

	Bulk resistivity [*]	Water	Water electrical	Cementation	Cation exchange
	[ohm.m]	temperature [†]	conductivity [†]	factor [‡]	capacity [§]
		[°C]	[S/m]	[unitless]	[meq/100g]
Overburden	< 500	12.1 - 16.0	0.027 - 0.045	2.5	2.0 – 6.0 [¶]
Broken zone	500 - 2500	12.1 - 16.0	0.027 – 0.045	2.5	2.0 - 6.0
Fissured zone	300 - 1000	12.8 - 14.2	0.029 - 0.050	2	5.2 – 18.7
Massive zone	1000 - 10 000	12.7 – 13.1	0.033 – 0.057	1.5	1.2 – 1.7

* 2D distribution obtained from ERT (see Results section).

+ average values (summer 2009) recorded in boreholes (low temporal variability).

‡ from Hartmann and Beaumont (1999).

§ values calculated from mineralogical analysis and natural gamma log (see Results section).

¶ same values as for the broken zone due to lack of in situ data.

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Table 2. Clay compositions in the different boreholes and resulting bedrock clay weight fraction and 995 996 cation exchange capacity (CEC)

	Muscovite [*] % vs tot clav	Chlorite [*] % vs tot clav	Illite [*] % vs tot clav	Montm. [*] % vs tot clav	Nat. γ [†]	Clay weight fraction # %	Total CEC [‡]
G01					662		11104/ 1008
Broken zone (transition)	58	27	7	8	116	51	6.0
Fissured zone (shallow)	40	20	22	18	111	51	11.7
Massive zone (deep)	66	34	0	0	102	43	1.7
GO2							
Broken zone (transition)	75	11	8	5	101	40	3.2
Fissured zone (shallow)	18	3	47	32	102	48	18.7
Massive zone (deep)	82	18	0	0	116	47	1.2
GO3							
Broken zone (transition)	77	14	5	4	78	30	2.0
Fissured zone (shallow)	28	5	40	27	41	16	5.2
Massive zone (deep)	82	18	0	0	116	47	1.2

* dominant clay minerals that affects natural gamma logging and rock bulk CEC. CEC (muscovite)~1 meq/g; CEC (illite)~10 meq/g; CEC (illite)~20 meq/g; CEC (montmorillonite)~90 meq/g (multiple sources, see Methods section). +average natural gamma count per seconds from borehole logging.

‡ Clay weight fraction and CEC calculated from individual clay CEC and natural gamma according to Revil et al. (1998).

1001 Table 3. Summary of isotropic hydraulic conductivities obtained from pumping and recovery test 1002 interpretation and comparison with previous results

Hydrogeological unit	Thickness from ERT [m]	Mean <i>K</i> [m/d]	K range [m/d]	Mean <i>K</i> from previous works [*] [m/d]	K range from previous works [*] [m/d]
Overburden	10 (in valley floor)	8.7	-	9	-
Broken bedrock (transition z.)	15-60	1.3×10 ⁻¹	1×10 ⁻¹ – 2×10 ⁻¹	1	7×10 ⁻² – 7
Fissured bedrock (shallow z.)	5-40	1.0×10 ⁻²	3×10 ⁻³ – 4×10 ⁻²	1 10 ⁻²	1×10 ⁻³ – 4×10 ⁻²
Massive bedrock (deep z.)	>40	6.0×10 ⁻³	4×10 ⁻³ – 8×10 ⁻³	1 10 ⁻²	7×10 ⁻³ – 6×10 ⁻²

* initial hydraulic testing after drilling (Moe et al. 2010) using Horslev (infiltration tests) and Theis-Jacob (pumping tests)

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 Table 4. Recharge values recalculated using the water table fluctuation methods and the specific
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yield values derived from MRS. Borehole name suffixes SS (subsoil), T (transition) are derived from 1007 the codes listed in section Hydrogeological setting

	Rainfall [*] [mm/y]	Water table cumulated rise [*] [m]	Specific yield [†] [%]	Calculated recharge [mm/y]	Calculated recharge [% of rainfall]
Hillslope					
GO1T 2010-2011	1134	13.6	0.75	102	9
GO1T 2011-2012	1433	19.7	0.75	147	11
GO2T 2010-2011	1134	8.8	1.57	138	13
GO2T 2011-2012	1433	16.8	1.57	264	19
Average				163	13
Valley floor					
GO3SS 2010-2011	1134	7.4	3	222	21
GO3SS 2011-2012	1433	11.7	3	351	24
Average				287	23

* from Cai & Ofterdinger (2016)

⁺ from MRS, assuming specific yield = 0.5*MRS water content

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	Kmax [*]	Kmin/Kmax	Anisotropy	Porosity [†]	Recharge [‡]	
	[m/d]	[unitless]	angle [°]	[%]	[mm/y]	
Model 1						
Overburden	9	1	Isotropic	20	Valley floor: 95 (480)§	
Broken zone	1 (0.2) [§]	1	Isotropic	4	Hillslope: 75 (370)§	
Fissured zone	0.01 (0.002) [§]	1	Isotropic	0.2	NA	
Massive zone	0.01 (0.002) [§]	1	Isotropic	0.01	NA	
Model 2						
Overburden	8.6	1	Isotropic	7	Valley floor: 285	
Broken zone	0.14	0.5	75 ° dip NW	3	Hillslope: 165	
Fissured zone	0.014	0.3	85 ° dip SE	0.1-0.5	NA	
Massive zone	0.008	0.1	65 ° dip SE	0.01	NA	

* from Moe et al. (2010) for Model 1; from Comte et al. (2012) for Model 2

+ from Comte et al. (2012) for Model 1; from Waxman and Smits' model and MRS results for Model 2 (Figure 7)

‡ from Cai & Ofterdinger (2016) for Models 1; recalculated values for Model 2 (Table 4)

§ values in brackets are final values after model calibration to observed heads in boreholes (K/5 and Recharge*5)

1011	Figure captions
1012	
1013 1014	Fig. 1 . Study site physical setting maps; (a) site location within the Irish basement geological framework (modified from Geological Survey of Ireland 2006); (b) catchment boundary with location
1015 1016	of hydrological monitoring infrastructures; (c) local interpretative geological map with location of the borehole clusters, the ERT profile and the MRS soundings.(a) and (b) modified from Comte et al.
1017 1018	(2012).
1010	Fig 2 Generic hydrogeological concentual model of weathered /fractured rocks aquifers in the
1015 1020 1021	context of the Irish terminology (modified from Comte et al. 2012).
1021	Fig 3 Fracture note density distributions (Schmidt net lower hemisphere projection) and orientations
1022	(fracture azimuth rose diagrams) from outcrons and horeholes (acoustic televiewer probe): hlue
1023	noints and arrows show the hydraulically active fractures unambiguously identified in horeholes
1024	(Nitsche 2014: modified from Comte et al. 2012)
1025	
1020	Fig. 4 ERT results (a) and interpreted concentual model of the weathered (fractured aquifer (b)
1027	rig. 4. ENT results (a) and interpreted conceptual model of the weathered/fractured aquifer (b).
1029	Fig. 5 . Comparison of hydraulic conductivities obtained from; (a) initial isotropic hydraulic test
1030	interpretations from Moe et al. (2010) used in Model 1; (b) refined isotropic interpretation from
1031	Comte et al. (2012); and (c-e) anisotropic K values used in Model 2, (c) K _{max} , (d) K _{equivalent} , and (e) K _{min} .
1032	
1033	Fig. 6. Vertical distribution of MRS water content for the 8 MRS sounding (see locations on Figure 1)
1034	and comparison with the aquifer conceptual model units delineated from ERT (Figure 4).
1035	
1036	Fig. 7. Spatial variations of storage properties derived from ERT and MRS geophysical data; (a) ERT
1037 1038	resistivity model with location of approximate volume of investigation of the MRS soundings; (b) MRS water content loas: (c) FRT porosity calculated from Waxman & Smits' model. Hatched areas indicate
1039	the unsaturated zone for which saturated Archie and Waxman & Smits models used are not
1040	annlicable
1041	apprease.
1041	Fig. 8. Conceptual aquifer geometries implemented in the numerical models: (a) generic tabular
1042	structure of weathered (fractured lowers from berebele interpretation (Medel 1), (b) complex lowers
1045	structure of weuthereu/fructureu luyers from borenole interpretation (woder 1), (b) complex luyereu
1044	structure derived from geophysical data reconciled with borehole logs (Model 2). 1: Overburden
1045	(alluvial and glacial sediments); 2: Broken bedrock (transition zone); 3: Fissured (shallow) bedrock; 4:
1046	Massive (deep) bedrock; 5: Substratum (very low productivity).
1047	
1048	Fig. 9 . Simulation results for Model 1 (<i>a</i>,<i>b</i>,<i>c</i>) and Model 2 (<i>d</i>,<i>e</i>,<i>f</i>) showing groundwater seepage rates
1049	at the model surface, Darcy's fluxes variations across the transect (a,d), groundwater mean
1050	flowpaths (b,e), and groundwater ages (c,f).
1051	
1052	Fig. 10. Relative distributions of aroundwater flow rate (as % of total flow) in the four conceptual
1053	aquifer units (overburden, broken, fissured and massive bedrock) for the two models considered. The
1054	massive bedrock here also includes the deeper levels shown in Figure 8.
1055	, 5
1056	Fig. 11. Modelled groundwater ages in the different boreholes for the two model cases. The hatched
1057	area indicates aroundwater ages that are older than the modern period of high atmospheric Tritium
1058	levels, and inconsistent with measured Tritium concentrations in boreholes samples.





(1) After Moe et al. (2010)

(2) After Comte et al. (2012)



n = 237









mean water content
 max water content
 min water content

water content > 2 %



approx. thickness of the unsaturated zone
approx. base of the broken bedrock

(as delineated by ERT; Fig. 4)









Model 1



