



**HAL**  
open science

## **Managed Aquifer Recharge in fractured crystalline rock aquifers: Impact of horizontal preferential flow on recharge dynamics**

Madeleine Nicolas, Olivier Bour, Adrien Selles, Benoît Dewandel, Vincent Bailly-Comte, Subash Chandra, Shakeel Ahmed, Jean-Christophe Maréchal

### ► To cite this version:

Madeleine Nicolas, Olivier Bour, Adrien Selles, Benoît Dewandel, Vincent Bailly-Comte, et al.. Managed Aquifer Recharge in fractured crystalline rock aquifers: Impact of horizontal preferential flow on recharge dynamics. *Journal of Hydrology*, 2019, 573, pp.717-732. <10.1016/j.jhydrol.2019.04.003>. <insu-02093946>

**HAL Id: insu-02093946**

**<https://insu.hal.science/insu-02093946v1>**

Submitted on 22 Oct 2021

**HAL** is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.



Distributed under a Creative Commons CC BY-NC 4.0 - Attribution - Non-commercial use - International License

# 1 **Managed Aquifer Recharge in fractured crystalline rock aquifers:** 2 **impact of horizontal preferential flow on recharge dynamics**

3 Madeleine Nicolas <sup>a,b,†</sup>, Olivier Bour <sup>b</sup>, Adrien Selles <sup>a</sup>, Benoit Dewandel <sup>c</sup>, Vincent Bailly-Comte <sup>c</sup>, Subash Chandra  
4 <sup>d</sup>, Shakeel Ahmed <sup>d</sup>, Jean-Christophe Maréchal <sup>c</sup>

5 <sup>a</sup> BRGM, Univ Montpellier, Indo-French Center for Groundwater Research, Uppal Road, 500007 Hyderabad,  
6 India

7 <sup>b</sup> Univ Rennes, CNRS, Géosciences Rennes - UMR 6118, F-35000 Rennes, France

8 <sup>c</sup> BRGM, Univ Montpellier, 34000 Montpellier, France

9 <sup>d</sup> CSIR-National Geophysical Research Institute, Indo-French Center for Groundwater Research, Uppal Road,  
10 500007 Hyderabad, India

---

## 11 **Abstract**

12 To overcome water scarcity issues, Managed Aquifer Recharge (MAR) structures are currently developed in  
13 many parts of the world, including poorly permeable terrain like weathered crystalline rocks. In such geological  
14 context, characterized by relatively limited groundwater storage mainly associated with fractures located at the  
15 interface between the upper weathered layer (saprolite) and the fractured bedrock, the efficiency of MAR is  
16 poorly known. To address this question and better understand the factors that control recharge dynamics, an  
17 artificial recharge basin was implemented at the Experimental Hydrogeological Park in Telangana (South India),  
18 a well-equipped and continuously monitored site situated in Archean granitic terrain. The thickness of the  
19 saprolite and hydraulic properties are relatively well known all over the site from previous geophysical surveys  
20 and hydraulic tests.

21 To characterize recharge dynamics, recharge has been monitored in different boreholes surrounding the  
22 infiltration basin. Infiltration rates and water level data are interpreted by both a volume balance approach and  
23 different analytical solutions. In addition, a simple numerical model was used to show how the depth of the  
24 permeable interface between saprolite and granite controls recharge dynamics and observed water levels  
25 variations. Results show that the permeability of the saprolite/bedrock interface is sufficiently large to allow an

---

<sup>†</sup>Corresponding author. Email: [madeleine.nicolas@univ-rennes1.fr](mailto:madeleine.nicolas@univ-rennes1.fr)

26 efficient recharge that propagates laterally throughout the aquifer through this well connected interface.  
27 However, the variable depth of this permeable pathway controls the water level response, acting as a semi-  
28 impervious boundary, leading to remarkable water level variations. Thus, our findings show how the  
29 characteristics of the most permeable pathways control recharge dynamics in weathered crystalline rocks. In  
30 addition, we show how the depth variations of the permeable interface between saprolite and granite may be  
31 inferred from the monitoring of water level during recharge events.

## 32 **Highlights**

- 33 - Preferential flow in the weathering interface controls recharge propagation
- 34 - Variable bedrock relief causes aquifer compartmentalization and focused recharge
- 35 - Modeling shows compartment dimensions control the rate and amplitude of WL increase

36 Keywords: Managed Aquifer Recharge, Crystalline rock aquifers, Compartmentalization, Bedrock topography,  
37 Weathering interface, Numerical modeling

## 38 **Funding**

39 This work has mainly benefited from CARNOT Institute BRGM funding. The Choutuppal Experimental  
40 Hydrogeological Park has also benefited from INSU support within the H+ observatory.

## 41 **Declaration of interest**

42 None

**43 List of abbreviations**

- 44  $\alpha$  : Coefficient of proportionality between infiltration and water level in the basin (dimensionless)
- 45  $b$  : Original saturated thickness (L)
- 46  $\bar{b}$  : Average saturated thickness (L)
- 47  $E$  : Evaporation (L/T)
- 48  $h$  : Water table elevation above the base of the aquifer (L)
- 49  $h_{sim}$  : Simulated water table elevation above the base of the aquifer (L)
- 50  $h_{obs}$  : Observed water table elevation above the base of the aquifer (L)
- 51  $H$  : Water level in the basin (L)
- 52  $Inf$  : Infiltration (L/T)
- 53  $K$  : Horizontal permeability/ hydraulic conductivity (L/T)
- 54  $K_V$  : Vertical permeability/ hydraulic conductivity (L/T)
- 55  $L_b$  : Length of rectangular recharge basin (L)
- 56  $L$  : Aquifer thickness (L)
- 57  $P$  : Precipitations (L/T)
- 58  $Q_{in}$  : Supply canal input (L/T)
- 59  $R$  : Recharge rate (L/T)
- 60  $RMSE$  : Root Mean Square Error
- 61  $S$  : Storativity (dimensionless)
- 62  $t$  : Time (T)
- 63  $T$  : Transmissivity (L<sup>2</sup>/T)
- 64  $VBA$  : Volume Balance Approach
- 65  $W_b$  : Width of rectangular recharge basin (L)
- 66  $x, y$  : Cartesian coordinates with center of recharge basin as origin (L)

## 67 **1. Introduction**

68 Despite their low yields and complexity (Roques *et al.*, 2016), many regions in the world depend on fractured  
69 crystalline aquifers as the only source of freshwater (UNESCO, 1999), particularly in arid and semi-arid regions  
70 like India. Crystalline rocks cover about 70% of India's geographical area (Saha *et al.*, 2013). Furthermore, India  
71 has seen an unprecedented development of groundwater exploitation within the last 50 to 60 years. Irrigation  
72 potential increased from 6.5 M ha to 45.7 M ha (Sharma *et al.*, 2005), out of which it is estimated crystalline  
73 rock aquifer groundwater represents more than 50%, especially in South India (Planning Commission, 2011).  
74 The advent of this "Green Revolution", aimed towards increasing agricultural output and achieving food  
75 security, has also brought on a number of water scarcity and groundwater quality degradation issues (Pingali,  
76 2012; Pinstруп-Andersen & Hazell, 1985; R. Singh, 2000). Natural replenishment of groundwater reservoirs has  
77 become insufficient to keep pace with the excessive continued exploitation of groundwater resources in many  
78 regions, leading to a long-term drop of groundwater levels (Central Ground Water Board, 2013).

79 In this context, the government has set out to remediate water table depletion by increasing aquifer recharge  
80 from 9% of total rainfall under natural conditions to 15% by 2020 (Government of Andhra Pradesh, 2003)  
81 through the development of large-scale managed aquifer recharge (MAR) schemes. The Central Ground Water  
82 Board has proposed the building of 11 million MAR structures nation-wide, as well as the reparation,  
83 renovation and restoration of the already existing structures, the total costs tallying up to over 12bn USD  
84 (Central Ground Water Board, 2013). A wide spectrum of MAR techniques may be implemented to recharge  
85 the groundwater reservoir, and these are selected mainly in terms of the hydrogeological framework. For  
86 example, infiltration methods (such as percolation tanks) are prioritized in relatively permeable environments,  
87 such as alluvial formations, while recharge shafts are preferred when the aquifer is overlain by poorly  
88 permeable strata; injection wells are used if the aquifer is confined. In India, and in the state of Telangana,  
89 underlain mainly by hard rock, the most prevalent recharge structures are percolation tanks and infiltration  
90 basins (Central Ground Water Board, 2013). It was estimated in the latest census of Minor Irrigation Sources  
91 that there are 46,531 percolation tanks, all of which the Department of Irrigation has set out to restore by 2020  
92 (Irrigation & CAD Department, 2015).

93 MAR can be used to address a wide range of water management issues, including: reducing seawater intrusion  
94 or land subsidence (e.g. Masciopinto, 2013), smoothing out supply and demand fluctuations (e.g. Palma et al.,  
95 2015), protecting groundwater-dependent ecosystems (e.g. Ward & Dillon, 2012), and improving water quality  
96 through filtration and chemical and biological processes (e.g. Hamadeh et al., 2014). In this particular case,  
97 however, MAR is only expected to remediate groundwater exploitation, and water quality is not managed  
98 (Jakeman et al., 2016).

99 Overall, common belief is that MAR initiatives are a viable and suitable solution to water scarcity issues.  
100 However, there is little consensus within published studies about their efficiency and impact, especially in hard  
101 rock context. At small scales (a few square meters to a few square kilometers), results often point towards  
102 increased groundwater resources and predominantly positive impacts of artificial recharge (Massuel *et al.*,  
103 2014; Srivastava *et al.*, 2009), which benefit only landholders closest to the recharge facilities (Boisson et al.,  
104 2015b; Dillon et al., 2009). Larger-scale studies (at the watershed scale or bigger) are less common (e.g.  
105 Glendenning *et al.*, 2012), on one hand, because of the difficulty in obtaining relevant data (de Marsily *et al.*,  
106 2005), and on the other on the complexity associated with groundwater reservoirs in crystalline rock. Further,  
107 many of these studies are limited to water budget analysis and do not focus on hydrodynamic processes (e.g.  
108 Scanlon *et al.*, 2012; Boisson *et al.*, 2015b). While water budgets can provide useful guidelines for groundwater  
109 exploitation, many authors have pointed out their insufficiency in providing accurate estimates of safe yield  
110 and sustainability, highlighting the need to understand dynamic processes (Bredehoeft, 2002; Zhou, 2009).  
111 Information on dynamic processes is further necessary to make out the spatial distribution of artificial  
112 recharge.

113 Weathered crystalline rock aquifers are characterized by highly variable hydraulic properties (Acworth, 1987;  
114 Chilton & Foster, 1995; Dewandel *et al.*, 2012; Maréchal *et al.*, 2004) which result in a complex combination of  
115 diffuse recharge and preferential flows (Alazard *et al.*, 2016; Reddy *et al.*, 2009; Sukhija *et al.*, 2003), where the  
116 latter have often been shown to dominate groundwater recharge processes (Cuthbert & Tindimugaya, 2010;  
117 Gleeson *et al.*, 2009; Sukhija *et al.*, 2003). In the aquifer, horizontal preferential flows take place in the most  
118 transmissive zones of the aquifer, which consist of the main open fractures and most importantly the bedrock  
119 weathering interface at the limit between the upper weathered layer (saprolite) and the fractured granite; this

120 has been recognized by many authors (e.g. Acworth, 1987; Chilton & Foster, 1995; Dewandel *et al.*, 2006;  
121 Boisson *et al.*, 2015a). The presence of these zones leads to a vertical anisotropy of permeability (Maréchal *et*  
122 *al.*, 2004, 2003). Since the bedrock weathering interface is in general hilly, its relief may control groundwater  
123 flows, leading to aquifer compartmentalization, where exchanges between compartments depend on water  
124 levels relative to interface topography (Guihéneuf *et al.*, 2014). The exact nature of the relationship between  
125 the bedrock weathering interface relief and recharge is not yet clearly understood. While this study focuses on  
126 fractured crystalline rock, it should be noted that the existence of sub-horizontal preferential flow pathways  
127 originating from sharp vertical contrasts in transmissivity exists in other type of media as well, such as stratified  
128 aquifers (Nimmo *et al.*, 2017), and that the observations outlined in this paper may be applicable in these  
129 environments as well.

130 Artificial recharge may be affected by preferential flow paths in contrasting ways. Their existence could  
131 enhance recharge, allowing an efficient and rapid lateral transfer of percolation throughout the aquifer. On the  
132 other hand, the compartmentalization associated with these environments highlighted by previous studies,  
133 could possibly slow or stop the progression of infiltration fronts leading to focused recharge in specific areas.  
134 Monitoring of MAR infiltration front progression is therefore necessary to better understand the role of  
135 preferential flow paths on the efficiency of recharge processes in crystalline rock aquifers, not only from a  
136 quantitative standpoint but also in regards to water quality and pollution propagation issues.

137 The aim of this study is to characterize artificial recharge processes in weathered crystalline rock and analyze  
138 the complex flow dynamics of recharge inputs through the main flow paths, namely the relief of the  
139 saprolite/bedrock interface. To do so, we monitored an artificial recharge basin that has been set up in an  
140 experimental site equipped with a network of observation borewells. A water balance and simple infiltration  
141 equations were used to quantify the inputs of the recharge basin and their temporal evolution, as well as the  
142 vertical hydraulic properties. Then, the lateral progression of the infiltration front in the underlying hard rock  
143 reservoir was analyzed. Analytical solutions were used to infer the lateral hydrodynamic properties of the  
144 media, while numerical modeling allowed us to quantify the effects of basement relief on recharge in order to  
145 explain the particularities of the observed recharge process.

## 146 **2. Study site**

147 The infiltration basin is situated within the Experimental Hydrogeological Park (EHP), which is a hydrological  
148 observatory located near the Choutuppal village in the Nalgonda district (Telangana state since 2014), 60 km to  
149 the south-east of Hyderabad (Latitude: 17°17'47"N; Longitude: 78°55'12"E), in South India (Fig. 1). The site has  
150 been developed by the French Geological Survey (BRGM) in partnership with the Indian National Geophysical  
151 Research Institute (NGRI) and is part of the H+ Observatory network. Most data used in this study can be  
152 downloaded from the H+ database (<http://hplus.ore.fr/en>).

## 153 2.1 Geological setting

154 The site is located in an Archean granite setting, which represents over 80% of the total surface of the  
155 Telangana state. This granitic formation is intruded locally by geological discontinuities such as dikes or quartz  
156 reefs, but none are present on the EHP (Guihéneuf *et al.*, 2014). The fracturing of the granite is mainly  
157 characterized by sub-horizontal fractures, which can display a lateral extension of tens of meters (Guihéneuf *et*  
158 *al.*, 2014). The typical geological profile in the EHP was obtained through analysis of drilling cuttings, and  
159 generally follows the description provided in Dewandel *et al.* (2006), namely:

- 160 – A thin red soil layer (a few centimeters).
- 161 – A sandy regolith layer 0-2 m thick, made up of a sandy-clay composition with quartz grains (Dewandel *et*  
162 *al.*, 2006).
- 163 – A laminated saprolite layer of variable thickness (ranging from 0 to 20m), derived from in-situ weathering  
164 of granite. It presents a millimeter-spaced horizontal laminated structure and coarse sand-size clasts and a  
165 few preserved conductive fractures (Dewandel *et al.*, 2006). Due to its composition, the saprolite layer can  
166 reach a quite high porosity (bulk porosities are mainly between 5% and 30%), depending on the lithology  
167 of the parent rock and the degree of weathering (Dewandel *et al.*, 2006).
- 168 – Granite mainly consisting of quartz, potassium feldspars and biotite (Dewandel *et al.*, 2006). The few first  
169 meters of the granite, in contact with the laminated saprolite (i.e. the saprolite/granite interface), are  
170 highly weathered and fractured. The fracture density rapidly decreases with depth, although local  
171 transmissive fractures may be encountered up to 60 meters deep (Guihéneuf *et al.*, 2014) . The effective

172 porosity of this layer is relatively low, of about 1%, and is mainly ensured by the fissure zones (Dewandel  
173 *et al.*, 2006).

Fig. 1: Study site location and position relative to Hyderabad (a, b), the Musi River and the supply channel (c); borewell position within EHP site (d, e, f), orange points are those equipped with pressure sensors.

## 174 2.2 Hydrological setting

175 The climate of this region is semi-arid and controlled by the periodicity of monsoons. Mean annual  
176 temperature is 28°C with high temperatures of about 45°C during the dry season. The rainy season occurs from  
177 June to November for a yearly average of about 800 mm. Ephemeral streams may be present during the  
178 monsoon but are most of the time absent. Intra-seasonal water level variations depend on groundwater  
179 recharge but are generally comprised between 10 and 20 m below ground surface (m bgs) (Fig. 2). There is no  
180 apparent straightforward relation between rainfall episodes and water level variations, and some intense  
181 rainfall episodes do not elicit a groundwater response (like in 2012, 765 mm of rain, Fig. 2).

Fig. 2: Hydraulic head variations in the CH03 borewell (longest on-site observed time series) and rainfall. Recharge basin was first filled by the end of 2015 (black arrow). The horizontal gray line illustrates the limit between the saprolite and the granite determined from borehole cuttings. For information relating to technical specificities of the well, refer to Guihéneuf (2014)

182 Within the framework of the previously cited state-wise MAR project, an infiltration basin was dug on the EHP  
183 during 2015 to meet the demands of farmers in the area facing water scarcity. Land use in the vicinity of the  
184 observatory consists mostly of cotton fields, some rice paddies and a few orchards, although it is assumed  
185 pumping for irrigation does not impact water levels on the site as most pumped borewells are downstream  
186 from the site. The basin was dug using an excavator; its approximate dimensions are 120 m by 40 m, with a  
187 depth of about 2 m, effectively removing the regolith layer and extending into the saprolite. Debris was piled  
188 around the basin to create a bund and avoid spillovers. The basin is mainly supplied by a canal which deviates  
189 water from the Musi River, downstream the state capital, Hyderabad (Fig. 1). The distance traveled by the  
190 canal between the Musi River and our study site is of about 40 km. Over this distance, the canal branches out  
191 into several smaller canals, which supply a network of infiltration basins throughout the region. The water thus  
192 used to recharge aquifers is river water directly rerouted into the subsurface without any treatment  
193 whatsoever. Unfortunately, despite several attempts by various governments, the river Musi is the eighth

194 most polluted river in India, receiving nearly 645 million liters per day of sewage water (Nilesh, 2016), as well as  
195 industrial effluents let out by pharmaceutical and bulk drug companies (Cheepi, 2012; Nitin, 2018).

196 The basin was initially filled by the end of 2015 (black arrow, Fig. 2), but this period was not yet monitored. The  
197 continuous monitoring of the basin began during the following filling episode which took place in July 2016.  
198 Groundwater levels significantly increased following the filling of the basin, reaching the highest levels ever  
199 monitored on-site (Fig. 2). Note that in many countries, regulations state the need for an infiltration basin and  
200 the underlying aquifer to remain hydraulically disconnected (Bouwer, 2002; Carleton, 2010; Réfloch, 2018).  
201 This allows preservation of an unsaturated zone below the basin, which is necessary to allow aerobic processes  
202 to take place to partly prevent the propagation of contaminants. This criterion was not met during this study  
203 because the water supply relied on the sporadic opening of an upstream floodgate managed by a third party  
204 entity.

Fig. 3: Conceptual model of transmissivity ( $T$ ) and storativity ( $S$ ) profiles by depth extrapolated from several hydraulic tests performed on-site by from Boisson *et al.* (2015) and schematic representation of geological log (adapted from Boisson *et al.*, 2015a). Storativity measurements end at the top of the fissured bedrock because they could not be measured within the saprolite at the time of the study as the boreholes are fully cased down to the contact between saprolite and fissured bedrock

205 Previous works (e.g. Guihéneuf *et al.*, 2014; Boisson *et al.*, 2015a), during which a series of hydraulic tests were  
206 performed on-site, have allowed the estimation of the media's hydrodynamic properties (Fig. 3), namely the  
207 transmissivity ( $T$ ,  $\text{m}^2 \cdot \text{s}^{-1}$ ) and storativity ( $S$ ). The upper fractured granite and the fractures within the granite  
208 were shown to be the most conductive with a good storativity (Fig. 3, Boisson *et al.*, 2015a). On the other hand,  
209 the saprolite layer was characterized as poorly transmissive. It must be noted, however, that several authors  
210 have suggested the potential of preserved fractures within the weathered saprolite to contribute to  
211 preferential flow (Dewandel *et al.*, 2006; Perrin *et al.*, 2011), and that overall knowledge on saprolite  
212 properties is currently limited.

213 In sum, the most transmissive zones of the aquifer are the upper fractured zone also known as the weathering  
214 interface, laterally well connected, and some permeable fractures encountered at greater depths, which  
215 conversely have a limited extension and continuity. Because the weathering interface has a variable depth  
216 (Dewandel *et al.*, 2006; St. Clair *et al.*, 2015), groundwater flow dynamics shift depending on water table

217 elevation (Guihéneuf *et al.*, 2014). In the 2014 paper by Guihéneuf *et al.*, hydraulic tests performed under  
218 different water level conditions (high water levels and low water levels), in combination with observations on  
219 piezometric variations, revealed how aquifer compartmentalization causes this contrasting behavior. Under  
220 high-level conditions, the well-connected permeable upper granite/saprolite zone allows regional flows to take  
221 place, more specifically towards the northeast of the site (Guihéneuf *et al.*, 2014). Contrariwise, because  
222 fracture density and therefore connectivity decreases with depth, low water levels lead to a lateral  
223 compartmentalization of the aquifer. The system then shifts to an independent local flow system (Guihéneuf *et*  
224 *al.*, 2014).

225 The site is equipped with 30 borewells of different depths (Fig. 1), all of which are equipped with a casing  
226 extending into the saprolite, screened or open at the contact zone between the saprolite and the fractured  
227 granite. Tab. 1 features the technical characteristics of the site boreholes which were used for this study, and  
228 the hydraulic properties at each of these wells may be found in Guihéneuf *et al.* (2014). Note that the saprolite  
229 thickness, which ranges from 14 m to 24 m, is very variable (Tab. 1). 8 boreholes were continuously equipped  
230 with pressure and temperature sensors measuring piezometric levels and temperature during the time of this  
231 study. An additional probe was installed in the infiltration basin to monitor water levels in the basin. At the  
232 time-scales considered, the fracture system within the fractured granite and the overlying saprolite are  
233 assumed to be at equilibrium, meaning there is no significant pressure difference between the unconsolidated  
234 saprolite and the fractures in the underlying granite. This was later confirmed when superficial wells (<15 m)  
235 were dug only in the saprolite near the CH01 and CH02 boreholes (Fig. 1f) which featured virtually the same  
236 hydraulic head variations as the deeper wells. This is consistent with the unconfined response (high storativity,  
237 Fig. 3) of the interface to pumping tests under packers which confirms the role of vertical fractures in  
238 connecting overlying saprolite with underlying fractures.

239

	X (m)	Y (m)	Elevation (m asl)	Borehole depth (m)	Casing depth (m bgs)	Estimated saprolite thickness (m)
CH01	279075.05	1913316.12	365.69	73.20	23.90	24.00
CH02	279120.68	1913296.32	366.31	73.20	18.75	19.00
CH03	278906.31	1913534.27	365.63	50.30	14.20	14.70
CH08	278876.59	1913531.05	366.11	61.00	17.30	18.00
CH11	278904.75	1913564.44	365.70	56.40	21.00	19.50
CH15	278949.36	1913522.31	364.60	56.40	18.30	17.90
CH16	278956.63	1913558.42	364.61	56.40	17.30	15.20
CH18	278914.12	1913576.32	365.57	50.30	19.80	21.35

240 Tab. 1: Boreholes characteristics of the Experimental Hydrogeological Park in Choutuppall (Andhra Pradesh,  
 241 Southern India) from Guihéneuf et al. (2014). Boreholes location is provided in the UTM projected coordinate  
 242 system. Borehole depth and casing depth are given in meters below ground surface.

### 243 2.3 Water level variations in response to recharge

244 The filling of the artificial recharge basin caused water levels to rise significantly (Fig. 4). The basin remained  
 245 flooded for most of the observation period, although levels in the basin varied widely from a few cm up to  
 246 almost 3 m depending on the sporadic canal inputs. Each filling episode was followed by a recession period  
 247 ranging from a few days to a few months, becoming longer towards the end of the observation period.

248 Borewells can be divided into two groups relative to their groundwater level variations. The first, composed of  
 249 CH01 and CH02, is the cluster closest to the basin. The second comprises the boreholes in the cluster farthest  
 250 from the basin (CH03 to CH24), out of which only 3 are shown along the AB profile (Fig. 4) for clarity purposes.  
 251 The two groups are characterized by different patterns of water level variations but have similarities between  
 252 them. There are three identifiable phases for both clusters depending on the time,  $t$ , of observation, starting  
 253 from  $t=0$ , which is the beginning of the monitoring period (July 5<sup>th</sup> 2016):

- 254 - **P1**  $0 < t < 60$  days (approx.). The overall range of variations is weak. The water levels for neighboring  
 255 boreholes (CH01 and CH02) remain deep (about 15 m below surface) close to water levels observed  
 256 before basin infilling. This suggests the water table remains disconnected from the infiltration basin  
 257 during this period of time.

- 258 - **P2**  $60 < t < 80$  days (approx.). An abrupt increase in water levels, of about 15 m, is observed for  
259 neighboring boreholes (CH01 and CH02), while the basin water levels vary between 1 and 2 meters.  
260 During this period, hydraulic head reaches near-surface levels (CH02 is at 366.3 m asl and CH01 at  
261 365.7 m asl). Such variations suggest that the basin and the aquifer are becoming hydraulically  
262 connected. The increase in remote water levels (CH03 to CH24) is much weaker but shows a clear  
263 response. This phase ends at around 80 days, where an important inflection point can be observed  
264 (Fig. 4) marking a change in dynamics.
- 265 - **P3**  $t > 80$  days. For this period of time, water levels in the closest boreholes (CH01 and CH02) are very  
266 high, close to ground surface levels suggesting that the basin remains connected to the water table.  
267 Remote boreholes appear to increase strongly during the first 40 days where water levels reach a  
268 depth of only a few meters below ground. Interestingly, remote borehole levels appear to evolve in  
269 sync with very similar hydraulic heads measured in all boreholes.

270 These observations suggest that P2 is a transition phase between the disconnected phase P1 and the fully  
271 connected phase P3 during which the basin fills the whole aquifer. Accordingly, the hydraulic response for  
272 remote boreholes is delayed in relation to neighboring boreholes. When levels in the near vicinity reach a  
273 pseudo-steady-state, remote water levels begin increasing simultaneously. This indicates a sort of tipping point  
274 mechanism: the attenuation of water level increase in the near vicinity of the basin is related to the  
275 progression of the pressure front laterally.

Fig. 4: Basin water level variations (gray curve) and associated hydraulic head variations in site boreholes shown in Fig. 1. Only 3 boreholes are shown for clarity purposes, as variations between neighboring boreholes was very similar.

### 276 3. Methods

277 The methodology applied aims to characterize the hydraulic properties controlling infiltration and its lateral  
278 transmission throughout the aquifer, as well as the effects of bedrock topography on the transient aquifer  
279 response. This was done through a combination of analytical and numerical modeling using water levels in the  
280 basin, meteorological data (rainfall and potential evapotranspiration) and the observed water level response in  
281 the observation boreholes. Daily rainfall was measured on-site using an automatic weather station, and daily

282 open pan evaporation was measured and made available by ICRISAT (International Crops Research Institute for  
283 the Semi-Arid Tropics, about 80 km away from our site).

### 284 3.1 Estimating infiltration rates and vertical hydraulic conductivity

285 Infiltration ( $Inf$ ) from the basin into the aquifer was quantified to characterize recharge, and to be used as  
286 inputs to the analytical and numerical aquifer response models posteriorly described. To do so, a volume  
287 balance approach (VBA) was applied in parallel to infiltration equations that will be described hereafter.

288 The VBA approach requires knowledge of all of the water balance components: canal inflow  $Q_{in}$ , rainfall  $P$ ,  
289 evaporation  $E$ , and basin stock variations  $dH$ , all of which are in mm/day:

$$290 \quad \quad \quad Inf = Q_{in} + P - E - dH \quad (1)$$

291 It was assumed that the runoff component, considering the small impluvium surface (<1ha), was negligible.

292 Because it was not possible to measure canal inflow directly for technical reasons, the VBA approach was only  
293 applied to recession periods, where  $Q_{in}$  is assumed to be null.  $dH$  was deduced from the water levels  
294 measured continuously in the basin ( $H$ ) where  $dH = H_t - H_{t-1}$ .  $E$  was assumed to be equal to open pan  
295 potential evaporation measurements. The equation was applied at a daily time step. The VBA approach is the  
296 most complete as it accounts for all inputs and outputs to the system although it is not applicable during basin  
297 filling periods.

298 In the absence of rainfall and neglecting evaporation, one also expects infiltration rates to be simply related to  
299 water levels in the basin, with greater water levels leading to greater infiltration. To examine such  
300 relationships, infiltration rates from basin water level variations during recession periods were compared to the  
301 corresponding water levels in the basin. Thus, the relationship  $Inf = \alpha \times H$  (with  $\alpha$  as the coefficient of  
302 proportionality) was evaluated empirically. If a satisfactory relationship is obtained during recession periods,  
303 such relationship may be extrapolated to all times, including infilling periods. Comparison to VBA infiltration  
304 ensured that neglecting  $P$  and  $E$  is not too strong of an assumption. Finally, with infiltration roughly deduced  
305 from basin water levels, the canal inflow  $Q_{in}$  can be calculated during infilling periods as the remnant of Eq. 1.

306 Note that these relationships can generally be described by Darcy-type equations, directly linking infiltration to  
 307 water levels based on knowledge of saturated thickness and hydraulic conductivity (Bouwer, 2002). It was not  
 308 possible to estimate hydraulic conductivity during the initial phases (P1 and P2) because the depth of the  
 309 wetting front and the apparent hydraulic conductivity are poorly constrained in partially saturated  
 310 environments.

311 When there is a full hydraulic connection between the basin and the aquifer (P3), one may assume that the  
 312 system may be equated to a falling head permeameter during the recession periods. Flows are assumed to be  
 313 mainly vertical within the saprolite before reaching the weathering interface, which is assumed to be an a priori  
 314 much more conductive zone. It therefore is assumed that flows are mainly controlled by the saprolite's  
 315 properties, which offer the greater resistance to flow, and the head gradient over the saprolite's thickness. The  
 316 hydraulic head at the bottom of the saprolite may be given by the hydraulic head in the closest boreholes,  
 317 which are very close to ground level. Thus, the head gradient may be roughly estimated equal to  $H/L$ , i.e. the  
 318 difference in hydraulic head between the top of the saprolite ( $H + L$ ) and the bottom of the saprolite ( $L$ ), with  
 319  $L$  the thickness of the saprolite. The vertical flow can be thus described using the Darcy equation:

$$320 \quad Inf = K_V \frac{H}{L} \quad (2)$$

321 where  $K_V$  is the vertical hydraulic conductivity ( $L.T^{-1}$ ). Knowing that  $Inf \approx -\frac{dH}{dt}$  during recession periods, this  
 322 equation can be rearranged to yield  $-\frac{dH}{H} = \frac{K_V}{L} dt$ , where the boundary conditions of this problem are  $H = H_0$   
 323 at  $t = 0$ . If we integrate  $dH/H$  on the left hand side from  $H_0$  to  $H$  and the right hand side from 0 to  $t$ , then  
 324  $\ln\left(\frac{H_0}{H}\right) = \frac{K_V}{L} t$ . This yields:

$$325 \quad \frac{H}{H_0} = e^{-\frac{K_V}{L} t} \quad (3)$$

326 To check the consistency of the above equation, observed water levels were plotted for each recession  
 327 posterior to hydraulic connection (P3 phase). When verified, a linear regression allows estimating of  $-\frac{K_V}{L}$ , and  
 328 with known  $L$  hydraulic conductivity can be deduced. Finally, having estimated  $K_V$  during recession period  
 329 allows estimating infiltration also for the infilling periods using Eq. 2.

## 330        3.2        Modeling the aquifer response to infiltration

331        In a second step, we calibrated an analytical solution modeling groundwater level variations against  
332        observations to provide an estimate of the hydrodynamic parameters controlling lateral flow. The effects of the  
333        transmissive interface's relief on flow, however, were ascertained using simple numerical modeling since no  
334        analytical solution was available for this purpose. To do so, different scenarios were tested where bedrock  
335        relief and hydrodynamic properties were varied to assess the changes in groundwater response, and the  
336        overall likeness to the dynamics observed on the experimental site.

### 337                    3.2.1 Analytical solutions

338        Calibration of simple analytical solutions against observed groundwater levels can allow a first-order estimation  
339        of the horizontal hydraulic conductivity and storativity controlling groundwater flow.

340        Groundwater mechanisms are described by different mathematical models depending whether there is a  
341        hydraulic connection between the basin and the water table or not (Fig. 5). The most common analytical  
342        solutions for groundwater mounding are valid only in situations where the water table is assumed to be  
343        hydraulically disconnected from the bottom of the groundwater infiltration basin. Conversely, analytical  
344        modeling of mounds that are fully connected to any recharge structure is relatively rare and most often  
345        concerns streams or canals (e.g. Dillon & Liggett, 1983; Spanoudaki *et al.*, 2010); modeling of connected  
346        recharge basins or lagoons (e.g. Kacimov *et al.*, 2016) is much rarer. Further, if the mechanisms in place deviate  
347        from standard conditions assuming homogeneous media, then fully realistic descriptions of these processes are  
348        said to be beyond any simple analytical solutions (Alderwish, 2010). In this context, it was decided to use  
349        analytical modeling only in the initial phase of wetting and mounding (P1), when the recharge basin is clearly  
350        disconnected from the aquifer and thus flow conditions remain relatively simple and respect the initial  
351        conditions required. Because remote boreholes (CH03 to CH24) show little to no response during this phase,  
352        analytical modeling was only calibrated against observations in neighboring boreholes (CH01 and CH02), which  
353        show a clear response to basin infilling even during the initial phase of wetting and mounding (P1). We used  
354        Hantush's solution for rectangular basins (Hantush, 1967).

## 355 3.2.1.1 Hantush's solution for rectangular basins

356 In general, analytical solutions to predict the rate of growth and shape of a recharge mound solve the  
 357 governing partial differential equation describing the flow of groundwater as given by the linearized Boussinesq  
 358 equation (Warner *et al.*, 1989):

$$359 \quad K\bar{b} \left( \frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} \right) + R = S \frac{\partial h}{\partial t} \quad (4)$$

360 where the different variables and parameters are defined in Fig. 5. The solution developed by Hantush  
 361 (Hantush, 1967) simulates groundwater mound growth and decay in response to percolation, and is valid when  
 362 the top of the groundwater mound is disconnected from the bottom of the recharge basin:

$$363 \quad Z = \frac{R\bar{b}}{2S} \int_0^t \left( \operatorname{erf} \left( \frac{L_b+x}{\sqrt{4v\tau}} \right) + \operatorname{erf} \left( \frac{L_b-x}{\sqrt{4v\tau}} \right) \right) * \left( \operatorname{erf} \left( \frac{W_b+y}{\sqrt{4v\tau}} \right) + \operatorname{erf} \left( \frac{W_b-y}{\sqrt{4v\tau}} \right) \right) d\tau \quad (5)$$

364 where  $Z = h^2 - b^2$ ,  $v$  is a simplifying term with  $v = K\bar{b}/S$  and  $\tau = (t - t')$ , the time during which percolation  
 365 takes place. Initial conditions assume a horizontal water table and the boundary conditions of zero slope of the  
 366 mound profile at the center of the basin and at infinity.

Fig. 5 : Schematic representation of water level variations in response to artificial recharge from an infiltration basin featuring parameters used in analytical modeling of artificial recharge (adapted from Warner *et al.*, 1989)

367 In Eq. 5, recharge from infiltration is assumed to be constant. However, infiltration was found to vary  
 368 significantly in time. To get around this issue the approach was modified: rather than directly integrating Eq. 5  
 369 by Gauss quadrature, a convolution of the impulse response of the aquifer to  $R = 1$  and the daily infiltration  
 370 time-series was made, so that at a given time  $t$ :

$$371 \quad Z(t) = \sum_{k=1}^t Z_{k-1} + ZU(k) * R(t - k) \quad (6)$$

372 where  $ZU$  is the impulse response and  $R$  is the recharge rate assumed equal to daily infiltration. This solution  
 373 was applied and calibrated at each borehole for which hydraulic head measurements are available,  
 374 where  $x$ ,  $y$ ,  $b$ , and  $\bar{b}$  were set to match the borehole's position and initial water table conditions.

375 Hydraulic conductivity ( $K$ ) and storativity ( $S$ ) were obtained by optimizing the analytical models' performance,  
376 i.e. minimizing the root mean square error (RMSE) between  $h_{sim}$  (simulated water table elevation) and  $h_{obs}$   
377 (observed water table elevation). The 2D parameter space was thus explored in order to reach a parameter set  
378 for which the error was minimal and which fit realistic standards. The resulting 2D RMSE matrix was then  
379 analyzed to assess whether the model converged towards a unique solution, as opposed to a situation where  
380 there is equifinality of parameters, which implies there are several acceptable sets of parameters that cannot  
381 easily be dismissed and should be considered in the evaluation of uncertainty (Beven, 2006).

382 Because this solution simulates how recharge inputs move laterally throughout the aquifer, and considering  
383 the important vertical anisotropy of hydraulic conductivity in fractured crystalline aquifers (Maréchal *et al.*,  
384 2004, 2003),  $K$  is assimilated to horizontal hydraulic conductivity.

### 385 3.2.2 Numerical modeling for connected basin

386 While analytical solutions are easily and straightforwardly applied, they do not account for the geometry of the  
387 aquifer and specific boundary conditions. Under these circumstances, we tested the effect of bedrock relief on  
388 drawup, i.e. the increase of water level due to basin infiltration, using MODFLOW and its BCF2 package (which  
389 allows wetting of previously dry cells) (Harbaugh, 2005). MODFLOW is a US Geological Survey block-centered  
390 finite-difference modular flow model, and the source code is free public domain software. The model was run  
391 using FloPy (Bakker *et al.*, 2016), a Python package for creating, running and post-processing MODFLOW-based  
392 models.

393 The numerical simulations were based on very simple conceptual models. First, a synthetic scenario with  
394 constant recharge was tested in order to analyze the causal relationship between the boundary conditions  
395 imposed by the bedrock relief and the groundwater response to recharge. Synthetic results were roughly  
396 compared to field results. At this stage however the objective was not to attempt to perfectly match the  
397 observed groundwater response to infiltration. Then, in a second step aimed at better reproducing the  
398 recharge dynamics, actual estimated infiltration rates were used as inputs to the model.

399 The conceptual models tested all consist of a rectangular, unconfined aquifer overlain by a thick unsaturated  
400 zone with an infiltration basin in its central part (Fig. 6). The infiltration basin covers 6400 m<sup>2</sup> and recharges the

401 aquifer at a rate  $R$  of  $1 \times 10^{-6} \text{ m.s}^{-1}$  (about  $85 \text{ mm.day}^{-1}$ ) for the synthetic tests, or at a transient rate determined  
402 by previously calculated infiltration for the final applied scenario. Recharge is applied to the highest active cell  
403 (which is automatically determined), meaning there is no retardation or storage of water in the unsaturated  
404 zone. This is equivalent to simulating a partially or fully hydraulically connected environment, i.e. P2 and P3.  
405 The initial phase P1 was not simulated due to poorly constrained boundary conditions during this unsaturated  
406 phase, and because the MODFLOW model does not accurately accommodate transient unsaturated flow.  
407 Hydraulic parameters  $S$  and  $K$  were varied over one and an half and two orders of magnitude respectively ( $1 \times$   
408  $10^{-3} < S < 5 \times 10^{-2}$  and  $1 \times 10^{-5} < K < 1 \times 10^{-3} \text{ m.s}^{-1}$ ) to test the sensitivity of the model to these parameters  
409 while remaining in a realistic range and were later fixed respectively at  $1 \times 10^{-2}$  and  $1 \times 10^{-4} \text{ m.s}^{-1}$  to match the  
410 first-order estimation of parameters from the analytical model for the synthetic tests, or more precise values  
411 for the applied scenario. The ratio of vertical to horizontal conductivity was fixed at 1/10 according to Maréchal  
412 *et al.* (2004) results.

413 We only simulate one quadrant of the aquifer (Fig. 6a) because all four quadrants are symmetrical (therefore  
414 there is no flow between quadrants). The hydrological system was simulated using a grid of 100 rows and  
415 columns with a 20 m spacing (for a total model size of  $2000 \times 2000 \text{ m}$ , size at which it was estimated the outer  
416 boundary conditions did not affect the results for the simulated duration) and six 5 m thick layers (necessary to  
417 model stability) (Fig. 6b). To ensure that the grid size and disposition were stable tests were performed where  
418 the size of the model and of the grid cells were varied to ensure water levels near the boundary remained  
419 stable. Initially, layers 1-5 are dry, head is 5 m and only the deepest layer is saturated. The evolution of the  
420 groundwater mound over time was modeled using a transient simulation, which was run for one stress period  
421 of 300 days with a one day time step using the PCG2 solver.

Fig. 6 : Model grid (a) and layer (b) configuration used for numerical modeling (adapted from Chiang & Kinzelbach, 1992) with a conceptual 3D representation of the model's two scenarios (c)

422 Two scenarios were tested using two model configurations: a reference scenario, in which there is no  
423 compartmentalization, and a compartmentalized scenario where an impervious boundary was added (Fig. 6c).  
424 The reference scenario was used to simulate water table mounding resulting from local recharge in  
425 homogeneous conditions with a sub-horizontal bedrock. The other scenario accounted for the more complex

426 boundary conditions imposed by the basement relief, by featuring a cuboid depression at the center. The  
427 depression was aimed at representing a topographic depression at the granite/saprolite interface, as it was  
428 determined from previous Electrical Resistivity Tomography (ERT) soundings that the recharge basin is located  
429 above a bedrock depression. This was done by adding an impermeable ( $K$  and  $S = 1 \times 10^{-20}$ ) layer covering the  
430 bottom of the aquifer except at the center where the compartment is located (Fig. 6). We tested different  
431 compartment widths (ranging from 100 m to 1000 m with a 100 m interval) and heights (from 5 m to 25 m with  
432 a 5 m interval) to test their causal relationship to groundwater drawup dynamics.

433 Finally, a simulation was run using the site's actual infiltration estimations from 3.1 as inputs to account for the  
434 observed variability of inputs over time and fit the model to observations. Hydraulic parameters were set equal  
435 to those obtained with the analytical solution, only the compartment size was adjusted to fit the simulation to  
436 the observed data. This allowed us to test the model's ability to simulate the observed groundwater variations  
437 by assuming a heterogeneous basement elevation.

## 438 4. Results

### 439 4.1 Infiltration estimation and relative contributions

440 Estimating infiltration from the VBA (Eq. 1) or from water levels in the basin (Eq. 2) was almost equivalent.  
441 Rainfall and evaporation do not have a strong effect on stock variations, relative to canal inflow and infiltration.  
442 In total, it was estimated canal inflow accounted for 97% of inputs to the basin, and infiltration for 95% of  
443 outputs (Tab. 2): they are the main drivers of stock variations. Note that infiltration from recession periods was  
444 obtained using a volume balance approach, and then linked to water levels in the basin to obtain relationships  
445 that could be extrapolated to filling periods. These relationships were indeed found to be linear and are shown  
446 in full below. As explained in the previous section, canal inputs were estimated as the remnants of Eq. 1 with  
447 known infiltration.

Fig. 7 : Infiltration and canal inflow estimated for the observation period. Inflow is episodic and short-lived as it is controlled by the opening and closing of an upstream floodgate managed by a third-party entity.

448 Inputs from the canal are quite high (Fig. 7), and concentrated into short periods of time. Infiltration takes  
 449 place more homogeneously throughout the observation period, but there are still non-negligible variations that  
 450 mimic but do not follow perfectly water level variations (Fig. 7). Interestingly, when the recharge mound  
 451 reaches the base of the aquifer, between 55 and 65 days approximately, infiltration reaches its maximum value  
 452 around 350 mm/day for a water level in the basin close to 2.6 meters. Regardless of the large amounts of water  
 453 brought on afterward, and similar basin water levels, recharge remains bounded, reaching at most about 150  
 454 mm/day suggesting a decrease in the recharge potential. Another way to show this evolution is to investigate  
 455 the evolution of the relationship between daily infiltration and water levels in the basin (Fig. 8a) for each

Fig. 8 : Infiltration relative to water levels in the basin for each phase of infiltration (a) and proportionality coefficients between infiltration and basin water levels plotted as a function of time. Proportionality coefficients are obtained for each individual recession period following the equation  $Inf = \alpha \times H$  (b).

456 recession period.

457 Infiltration is roughly proportional to water levels in the basin, as expected from Eq. 2, but with different  
 458 relations for the different period of recessions. The coefficient of proportionality,  $\alpha$ , decreases strongly with  
 459 time (Fig. 8b) during phase P2. For times greater than 80 days (P3), no significant changes are observed,  
 460 however the efficiency of the infiltration seems to be reduced by a factor of 6 from the early times until 80  
 461 days. This behavior is expected when connecting the basin with the aquifer (Bouwer, 2002; Carleton, 2010)  
 462 because of the decrease in hydraulic gradient which results from the pore space filling up at the point of  
 463 hydraulic connection. This further supports the statement that P2 is a transition phase during which the basin  
 464 becomes hydraulically connected to the aquifer.

465

	Total contribution (mm)	Relative contribution (%)
Rainfall	680	3
Canal inflow	19,230	97
Total estimated input	19,900	100
Pan evaporation	960	5
Infiltration	19,140	95
Total estimated output	20,100	100

466 Tab. 2: Contributions of water budget components

467 **4.2 Vertical hydraulic conductivity**

468 Vertical hydraulic conductivity posterior to hydraulic connection was obtained using Eq. 3, from the recession  
 469 slopes shown in Fig. 9. It was found to vary slightly between each recession slope, ranging from  $7.4 \times 10^{-6} \text{ m.s}^{-1}$   
 470 to  $9.4 \times 10^{-6} \text{ m.s}^{-1}$ , averaging  $8.1 \times 10^{-6} \text{ m.s}^{-1}$ . Such values are slightly higher than what is typically expected in  
 471 the saprolite in this area, but much lower than the most transmissive pathways at the interface between  
 472 saprolite and the upper fractured granite (Fig. 3, Guihéneuf *et al.*, 2014; Boisson, *et al.*, 2015a).

Fig. 9 : Recession slopes when the basin is fully connected to the aquifer (P3) at an hourly time-step plotted on a semi-log axis where the slope is equal to  $-\frac{K_V}{L}$  (a), and close-up of the shorter recession slopes (b). Each line corresponds to a different recession slope; five different recessions are shown.

473 **4.3 Horizontal hydraulic conductivity and storativity**

474 Calibration of observed groundwater levels to simulated water levels for the disconnected phase (P1) using the  
 475 Hantush solution resulted in the parameters summed up in Tab. 3. The best model fits can be seen in Fig. 10.

	Distance to center of basin (m)	Hydraulic conductivity $K$ ( $\text{m.s}^{-1}$ )	Storativity $S$ (dimensionless)	$NRMSE$ (%)
CH01	100	$7.0 \times 10^{-5}$	$3.0 \times 10^{-2}$	1.5
CH02	67	$6.0 \times 10^{-5}$	$7.0 \times 10^{-2}$	1.4
	Geometric mean	$6.5 \times 10^{-5}$	$5.0 \times 10^{-2}$	

476 Tab. 3 : Hydraulic properties obtained from calibration of analytical simulations to observed hydraulic head

Fig. 10: Observed hydraulic head and simulated hydraulic head using analytical modeling in response to infiltration from the recharge basin when it is disconnected from the water table (P1) for the boreholes closest to the basin (CH01 and CH02)

477 RMSE was normalized by the range of variation ( $NRMSE = \frac{RMSE}{h_{max}-h_{min}}$ ). Model error was satisfactory,  $NRMSE$ s  
 478 being weaker than 5%. Averaged to all boreholes hydraulic conductivity was found to be  $6.5 \times 10^{-5} \text{ m.s}^{-1}$  with  
 479 little variability between both estimates. This estimate is clearly higher than the estimate of the hydraulic  
 480 conductivity of the saprolite obtained from water level recessions, averaging  $8.10^{-6} \text{ m.s}^{-1}$ , but very close to what  
 481 Boisson *et al.* (2015a) and Guihéneuf *et al.* (2014) obtained for the weathering interface (Fig. 3,  $\bar{T} = 1.3 \times 10^{-4}$   
 482  $\text{m}^2.\text{s}^{-1}$  and thus  $\bar{K} \approx 10^{-5} \text{ m.s}^{-1}$  the interface being a few meters thick). Storativity was found to be around  
 483  $5 \times 10^{-2}$ , which is higher than the Fig. 3 weathering interface estimate ( $\bar{S} = 9.7 \times 10^{-3}$ ), denoting a possible

484 influence of the saprolite on the apparent estimated storativity. According to the variations of storativity  
485 between CH01 and CH02, it nevertheless seems to weaken as we move away from the basin.

## 486 4.4 Effect of bedrock relief (numerical model)

### 487 4.4.1 Synthetic scenarios

488 Synthetic results obtained from numerical simulations are shown in Fig. 11. The reference model, with  
489 parameters in the range of those obtained from analytical simulations ( $K = 10^{-4} \text{ m.s}^{-1}$  and  $S = 10^{-2}$ ), showed  
490 very weak variations and a weak drawup slope (black curve on Fig. 11a and b), which cannot explain observed  
491 water level variations from the study site. Recharge dynamics cannot be modeled satisfactorily using these  
492 reasonable hydraulic parameters. Sensitivity tests (not shown in Fig. 11) were also performed on  $K$  and  $S$ .  $S$   
493 variations of one order of magnitude led to a difference lower than 2 m. Decreasing  $K$  elicited a stronger  
494 groundwater response but with a weak drawup slope, again disagreeing with observed data.

Fig. 11: Simulated hydraulic head for the reference and compartmentalized scenario under constant recharge ( $R = 10^{-6} \text{ m.s}^{-1}$ ), where the lateral extension of the compartment is varied (a) and where the vertical extension is varied (b). Set distance to basin is 100 m.  $K = 10^{-4} \text{ m.s}^{-1}$  and  $S = 10^{-2}$

495 The most realistic simulation is obtained when considering a heterogeneous basement, and thus semi-  
496 penetrating impermeable boundary conditions. The lateral extension of the depression/compartment at the  
497 center of the model controls the slope of the drawup. Smaller lateral extensions of the compartment lead to  
498 stronger slopes (Fig. 11a), and as the lateral compartment size is progressively increased, drawup dynamics  
499 approach that of the reference scenario (i.e. no boundary effect). The vertical extension of the depression does  
500 not have an impact of the slope of the drawup. It does however control the amplitude of variations (Fig. 11b),  
501 where a greater difference in altitudes leads to a greater drawup amplitude.

### 502 4.4.2 Application to the present field case

503 The application of the compartmentalized scenario to estimated infiltration inputs proved quite successful in  
504 recreating the dynamics of the closest boreholes for phase 2 and 3 (Fig. 12) using the hydraulic parameters  
505 obtained from analytical modeling (Tab. 3) and varying only the shape of the basement. We obtained a

Fig. 12: Observed hydraulic head and simulated hydraulic head using numerical modeling accounting for compartmentalization in response to infiltration from the recharge basin (gray lines) for the boreholes closes to the basin (CH01 and CH02)

506 relatively good agreement between the simulated water levels and the observed ones ( $NRMSE = 8.7\%$  and  
507  $12.1\%$  for CH01 and CH02 respectively). The strong slope of the drawup is successfully reproduced, and the  
508 following stabilization of water levels as well. A lateral extension of about 120 m seems to best recreate the  
509 recharge dynamics of the study site with the above cited parameters, although storativity was shown to be  
510 somewhat overestimated under the influence of the saprolite values owing to the borehole's proximity to the  
511 basin. Decreasing storativity to a value closer to that of the weathering interface ( $S = 9.7 \times 10^{-3}$ ) requires  
512 increasing the theoretical lateral extension of the compartment to about 200 m to correctly model  
513 observations. The calibrated vertical extension was found relatively constant, around 18 m.

514 There is a slight discrepancy between the model and the simulation, where the simulation under-estimates  
515 water levels during the drawup phase and over-estimates them during P3. This is probably because the actual  
516 system modeled is in reality much more complex than the model proposed here, as we do not account for flow  
517 in fractures, storage effects or matrix effects. The underlying geometry of the interface may also be much more  
518 complex than the way it is simulated in the model. We believe this is a testament to the interest this model  
519 brings, wherein a very simple numerical model recreated complex dynamics. Note also that it was not possible  
520 to simulate remote boreholes using the model as is. Placing the observation point outside of the compartment  
521 did recreate the delay in borehole response observed in Fig. 4, but did not allow to simulate the amplitude of  
522 the variations correctly. Addressing this issue would imply complexifying the model to account for a much  
523 more complex geometry of the weathering interface, which is currently beyond the scope of this paper.

Fig. 13: Schematic representation of hydraulic head variations above the bedrock (gray area) in response to artificial recharge from a recharge basin per phase of infiltration

524 A schematic summary of recharge dynamics is presented in Fig. 13 in a very simplified way. The hydraulically  
525 disconnected phase of groundwater mounding (P1) resembles that of a relatively homogeneous environment.  
526 When hydraulic connection begins between the infiltration basin and the aquifer, we observe boundary  
527 conditions leading to a rapid increase of water levels because infiltration is focused into a bounded domain.  
528 The hydraulic connection also slows the infiltration process. The final stage P3 is much less subject to boundary  
529 effects and a pseudo-steady-state is reached, where levels oscillate around a constant value.

## 530 **5. Discussion**

## 531        5.1        Representativity of hydraulic parameters

532 Overall infiltration efficiency, i.e. the ratio of infiltration to total inputs, is quite high (96%) in the present case,  
533 as opposed to previous studies in similar environments (Boisson *et al.*, 2015b: 56% ; Perrin *et al.*, 2009: 54% ;  
534 Singh *et al.*, 2004: 67%), but in agreement with efficiencies recorded by the Central Groundwater Board in 2011  
535 of up to 98% in hard rock environments. A possible explanation of this difference is the age of the infiltration  
536 basins. Studies showing low infiltration efficiencies focused on percolation tanks that had not been recently  
537 desilted, contrary to the basin studied here (dug in 2015) and those present in the CGWB's case studies.  
538 Further, the way in which the basin is supplied, where large amounts of water are brought in short periods of  
539 time, leads to an important water depth in the basin (up to almost 3 m). This could contribute to stronger  
540 infiltration rates linked to a higher hydraulic head. This is as opposed to typical percolation tanks, which  
541 accumulate a few centimeters of runoff over larger extended areas and thus subject to a higher amount of  
542 direct evaporation compared to infiltration. Although overall infiltration efficiency is high, infiltration rates  
543 decrease significantly with hydraulic connection because of a decrease in hydraulic gradient. Infiltration is then  
544 more spread out in time, which allows more evaporation to take place, although the amount evaporated  
545 remains much weaker than the overall infiltrated volume. This is not likely to continue to be the case if  
546 significant silting occurs.

547 This study's estimate of vertical permeability ( $K_V = 8.1 \times 10^{-6} \text{ m.s}^{-1}$ ), is slightly higher than what Boisson *et al.*  
548 (2015a) obtained at the same site for the upper saprolite layer ( $9 \times 10^{-8}$  to  $3 \times 10^{-7} \text{ m.s}^{-1}$ ), but overall in the  
549 same range as other studies performed on saprolite (e.g. George, 1992:  $7 \times 10^{-6} \text{ m.s}^{-1}$ ; Cook *et al.*, 1996:  $1 \times 10^{-6}$   
550 to  $1 \times 10^{-5} \text{ m.s}^{-1}$ ). Boisson *et al.* (2015a) used falling-head borehole permeameter tests, which render point  
551 values, whereas the large horizontal extent of the infiltration basin could allow sampling of otherwise  
552 undetectable vertical preferential flow paths (i.e. preserved fractures in the saprolite), and thus lead to an  
553 increase in hydraulic conductivity (Dewandel *et al.*, 2006).

554 Horizontal hydraulic properties were found to be in the upper range of values for fractured crystalline rock  
555 ( $K = 6.5 \times 10^{-5} \text{ m.s}^{-1}$  and  $S = 5.0 \times 10^{-2}$ ), rather corresponding to the properties of the weathering interface  
556 obtained in Boisson *et al.*, 2015a. The horizontal hydraulic conductivity of the system is thus much higher than  
557 the vertical hydraulic conductivity (about 8 times higher). This is consistent with the fact that the vertical

558 hydraulic conductivity is controlled by the less transmissive saprolite, and the lateral hydraulic conductivity by  
559 the highly transmissive weathering interface acting as a preferential flow path. Overall, analytical modeling  
560 highlighted the strong influence of preferential flow in recharge processes.

## 561        5.2        Comparison to inferred interface relief

562 Previous acquisition of Electrical resistivity tomography (ERT) profiles (Chandra *et al.*, 2009) permitted the  
563 mapping of the depth of the weathering interface at the EHP site (Fig. 14). The ERT survey was carried out using  
564 a Syscal Jr. Switch with 10m electrode spacing across 9 profiles using a Wenner-Schlumberger configuration  
565 (Chandra *et al.*, 2009). Profiles were analyzed using a routine inversion method and a threshold of about 400  
566 Ohm.m (Braun *et al.*, 2009) was set to delineate the limit between fresh and weathered rock (Chandra *et al.*,  
567 2009). Point data was then interpolated using standard kriging techniques.

Fig. 14: Depth of the upper fissured layer from ERT surveys (a) and inferred 3D conceptual model of structure (b). Hypothetical compartment delimitation is the thin black lines. For clarity purposes only the wells shown in Fig. 4 are shown here.

568 The obtained interface is significantly hilly, composed of depressions and crests (the protruding areas), where  
569 crests delineate a set of sub-drainage basins (Fig. 14a). The hilliness of the interface relief suggests the system  
570 is compartmentalized, which supports the observations made in previous works highlighting the existence of  
571 compartmentalization in crystalline environments and its effects on groundwater flow and chemistry (e.g.  
572 Guihéneuf *et al.*, 2014; Perrin *et al.*, 2011). Two main compartments were identified: one east of the site and  
573 the other north of the site, extending beyond the site limits. The lowest point of the bedrock topography  
574 appears to be north of the profile. The compartment over which the infiltration basin is situated is trefoil-like-  
575 shaped, with a diameter of about 500 m, a low point at about 42m depth and summits at about 20 m depth (so  
576 a height difference of approximately 20 m). This is similar to the dimensions obtained while fitting the  
577 numerical simulation to observed water levels (i.e. a radius of 200 m and a height difference of 18 m).

578 Using the AB profile outlined in Fig. 4, we traced the relief of the bedrock interface obtained from ERT and  
579 replaced observed water levels so as to compare observed data to simulations (Fig. 15). The depth of the  
580 interface was very similar to the depth estimated from borehole logs. Neighboring boreholes (CH01 and CH02)

581 are situated in a topographic depression, whereas remote boreholes are in a crest area, where the two  
582 compartments identified in (Fig. 14a) are separated, CH03 being at the highest point.

583 The same phases outlined in Fig. 13 can be transposed to observed data. Prior to the supposed date of  
584 hydraulic connection, (in Fig. 15  $t = 0$  and  $t = 40$ ) water level variations are weak and the water table is deep,  
585 which corresponds to P1. Then, water levels increase rapidly under the effect of the lateral boundary condition  
586 (in Fig. 15  $t = 60$  and  $t = 70$ ), corresponding to P2. This continues until water levels exceed the topographic  
587 threshold “downstream” from the basin at about 70 days after the beginning of the observation period. From  
588 that point onwards compartments become connected and water levels vary weakly but on a large extent,  
589 which corresponds to P3.

Fig. 15: Comparison of schematic representation of artificial recharge from an infiltration basin with actual hydraulic head measured on-site and aquifer geometry from ERT data

590 This conceptual model supports observations made by Guihéneuf *et al.* in 2014: the hydrogeological system is  
591 compartmentalized, and the aquifer connectivity and associated flow regime is a function of water levels  
592 relative to the well-connected bedrock interface.

### 593 5.3 Artificial recharge modeling

594 The model most commonly used to explain recharge dynamics was proposed by Bouwer (2002). In this model,  
595 two scenarios explain an infiltration basin’s dynamics. If the bottom of the basin (the clogging layer) is less  
596 conductive than the underlying aquifer then infiltration is controlled by the hydraulic conductivity of the  
597 clogging layer. The water within the basin is disconnected from the underlying aquifer, perched over the silting  
598 layer. Otherwise, infiltration is controlled by the hydraulic properties of the underlying aquifer. In this case the  
599 water table may rise until hydraulic connection takes place, and then the infiltration rate is controlled either  
600 solely by gravity and vertical flows (in deep aquifers) or by the slope of the water table mound (in shallow  
601 aquifers). This concept was developed assuming homogeneous sedimentary aquifers of infinite lateral  
602 extension.

603 Instances of variable recharge dynamics have been documented in literature, either stemming from  
604 progressive and heterogeneous silting (e.g. Racz *et al.*, 2012; Mawer *et al.*, 2016), saturation of the vadose zone

605 (Dillon & Liggett, 1983) or more rarely structural discontinuities (Massuel *et al.*, 2014). In most cases, the  
606 influence of heterogeneity on flow is limited to qualitative observations. Very few studies actually model  
607 processes with consideration of aquifer geometry (e.g. Ronayne *et al.*, 2008), probably because it is not easy to  
608 image the aquifer's geometry and practically impossible to obtain three-dimensional information about  
609 hydraulic properties, especially in hard rock context (de Marsily *et al.*, 2005). In this study we successfully  
610 quantified the role played by the basement relief (and its associated compartmentalization) on the hydraulic  
611 response of the aquifer to recharge, also highlighting the effects of it on recharge propagation and infiltration  
612 potentials. This study showed that the relief of the bedrock can produce an effect somehow analogous to the  
613 effect of impervious (or semi-impervious) boundaries on drawdown during pumping tests (de Marsily, 1986). In  
614 both cases, a change in drawdown slope relative to a situation where no impervious boundary is present is  
615 expected. To the best of our knowledge, there are no other instances in literature where this effect was  
616 identified during recharge, or as a result of a heterogeneous bedrock relief. Further, because the relief crests  
617 are only a semi-penetrating impervious boundary, the increase in drawup slope relative to the reference  
618 scenario was shown to provide information on the dimensions of the relief in question, where the larger the  
619 depression below the recharge basin the lesser the increase in drawup slope.

620 The in-depth analysis of the transient aquifer response to recharge in a heterogeneous environment can be  
621 used to improve MAR management in crystalline rock. The requirements to perform this type of analysis are (i)  
622 an approximation of the aquifer's hydraulic properties, (ii) a map of the contact zone between the saprolite and  
623 the granite, and (iii) knowledge on the (actual or desired) water volumes to be stored in the aquifer. The most  
624 challenging aspect of this methodology is obtaining a map of the subsurface, although geophysical methods are  
625 becoming increasingly accurate and accessible. One can, for example, cite the development of airborne  
626 electromagnetic methods, which allow large-scale high resolution mapping of the subsurface (Sørensen &  
627 Auken, 2004). However, if bedrock relief data is unobtainable, we also showed how monitoring groundwater  
628 levels in the vicinity of the basin can provide valuable information on bedrock topography and on the storage  
629 potentials of the aquifer. As such, not only should the infiltration basin be monitored, but also, if possible,  
630 groundwater levels around the basin. Extrapolation of our simple method could improve present-day forward  
631 models used for MAR management in such geological environments, and possibly be extended into predictive  
632 modeling. From a quantitative standpoint, consideration of the effects of boundary conditions would be useful

633 to regulate inputs in order to limit evaporation losses. If the bedrock topography is known, decision-makers can  
634 gain a greater awareness of the extent to which recharge will propagate throughout the aquifer, and choose  
635 whether to prioritize a strong localized reaction, which may be easier to retrieve at later times, or a weaker  
636 reaction over larger areas depending on the density and extent of the settlements affected by groundwater  
637 depletion and which stand to benefit from the augmentation of groundwater resources. Further, implications  
638 on water quality are also non-negligible. We highlighted the importance of preferential flow, which implies  
639 contaminant attenuation is not likely to take place rapidly enough to mitigate the negative effects of injecting  
640 wastewater into the aquifer. The dominance of advective processes over diffusive processes in hard rock  
641 (Guihéneuf *et al.*, 2017) will likely aggravate this issue. It is essential this be taken into consideration by  
642 decision-makers: the water currently used to feed the network of MAR structures state-wise originates directly  
643 from the Musi River downstream of Hyderabad and is highly polluted both by domestic and industrial effluents.  
644 Another element of risk that was identified in this study is that of hydraulic connection between the infiltration  
645 basin and the water table, to which fractured crystalline environments are particularly susceptible due to their  
646 low transmissivities and variable saprolite thickness. The best way to infer hydraulic connection is through  
647 groundwater level observation, but it is still possible to infer that hydraulic connection has taken place when  
648 infiltration rates decrease significantly (which may lead the basin to overflow or water levels in the basin to  
649 stagnate). Additionally to the loss of infiltration potentials, which may dampen the efficiency of this water  
650 stress remediation method and lead to important losses by ET, hydraulic connection may amplify the risk of  
651 groundwater contamination. If no unsaturated zone exists below the infiltration basin, aerobic processes and  
652 sorption of contaminants which play an important part in retaining and degrading contaminants, and that  
653 depend on residence times, may be impeded.

## 654 **6. Conclusions**

655 Our study of artificial recharge processes in a fractured crystalline environment highlighted the role the  
656 heterogeneity of the horizontal weathering interface depth plays on recharge. During the initial stages of  
657 wetting, when the recharge mound is disconnected from the bottom of the basin, infiltration is governed by  
658 the vertical properties of the relatively homogeneous and poorly conductive saprolite. The lateral transmission  
659 of infiltration inputs reaching the aquifer is ensured mainly by the weathering interface, thus foregrounding the

660 importance of preferential flows. Once hydraulic connection between the infiltration basin and the aquifer  
661 takes place, which is relatively rapidly due to the shallowness of the aquifer, two simultaneous phenomena are  
662 observed: infiltration potential decreases, and water levels increase strongly and rapidly. It was shown from  
663 numerical simulations that this can be interpreted as resulting from the control by the weathering interface  
664 relief, where summits act as semi-impervious boundaries, which delimit a set of hydrogeological  
665 compartments. The advancement of the infiltration front is thus impeded until the crest separating the east  
666 and north hydrogeological compartments is exceeded and the compartment is filled, which focuses infiltration  
667 inputs into smaller volumes, explaining the strong and sudden local water level increase. Once the summits are  
668 overpassed, the different hydrogeological compartments are connected, and the no-flow boundary effect  
669 becomes less critical.

670 It has yet to be confirmed if the simple analytical and numerical approach used here can be applied more  
671 generally in crystalline rock environments, but we believe this is a first step in better understanding the  
672 propagation of infiltration fronts in heterogeneous aquifers. This in turn could provide useful information for  
673 the aquifer properties and for the management of MAR structures in regards to placement, optimization, and  
674 water quality.

## 675 **Acknowledgments**

676 This study has been carried out at the Indo-French Center for Groundwater Research (BRGM-NGRI). This work  
677 has mainly benefited from CARNOT Institute BRGM funding. The Choutuppal Experimental Hydrogeological  
678 Park has also benefited from INSU support within the H+ observatory. The authors are very grateful to Marion  
679 Crenner and Mohammed Wajiduddin for their fieldwork, Joy Chaudhuri and Vidya Sagar for their geophysical  
680 work, and Yata Ramesh, Yata Muthyalu, Pittala Krishna, Nilgonda Kishtaiah and Pittala Anjaiah for their  
681 valuable on-site help. The authors are also grateful to ICRISAT for providing evaporation data. The authors are  
682 thankful to the editor Peter K. Kitanidis, the associate editor and to the reviewers (John Nimmo and an  
683 anonymous reviewer) for their constructive comments, which greatly enhanced the quality of the manuscript.

684 **References**

- 685 Acworth, R. I. (1987). The development of crystalline basement aquifers in a tropical environment. *Quarterly*  
686 *Journal of Engineering Geology and Hydrogeology*, 20(4), 265–272.  
687 <https://doi.org/10.1144/GSL.QJEG.1987.020.04.02>
- 688 Alazard, M., Boisson, A., Maréchal, J. C., Perrin, J., Dewandel, B., Schwarz, T., ... Ahmed, S. (2016). Investigation  
689 of recharge dynamics and flow paths in a fractured crystalline aquifer in semi-arid India using borehole  
690 logs: implications for managed aquifer recharge. *Hydrogeology Journal*, 24(1), 35–57.  
691 <https://doi.org/10.1007/s10040-015-1323-5>
- 692 Alderwish, A. M. (2010). Induced recharge at new dam sites—Sana’a Basin, Yemen. *Arabian Journal of*  
693 *Geosciences*, 3(3), 283–293. <https://doi.org/10.1007/s12517-009-0075-8>
- 694 Bakker, M., Post, V., Langevin, C. D., Hughes, J. D., White, J. T., Starn, J. J., & Fienen, M. N. (2016). Scripting  
695 MODFLOW Model Development Using Python and FloPy. *Groundwater*, 54(5), 733–739.  
696 <https://doi.org/10.1111/gwat.12413>
- 697 Beven, K. (2006). A manifesto for the equifinality thesis. *Journal of Hydrology*, 320(1–2), 18–36.  
698 <https://doi.org/10.1016/j.jhydrol.2005.07.007>
- 699 Boisson, A., Guihéneuf, N., Perrin, J., Bour, O., Dewandel, B., Dausse, A., ... Maréchal, J. C. (2015). Determining  
700 the vertical evolution of hydrodynamic parameters in weathered and fractured south Indian crystalline-  
701 rock aquifers: insights from a study on an instrumented site. *Hydrogeology Journal*, 23(4), 757–773.  
702 <https://doi.org/10.1007/s10040-014-1226-x>
- 703 Boisson, A., Villesseche, D., Baisset, M., Perrin, J., Viossanges, M., Kloppmann, W., ... Ahmed, S. (2015).  
704 Questioning the impact and sustainability of percolation tanks as aquifer recharge structures in semi-arid  
705 crystalline context. *Environmental Earth Sciences*, 73(12), 7711–7721. [https://doi.org/10.1007/s12665-](https://doi.org/10.1007/s12665-014-3229-2)  
706 [014-3229-2](https://doi.org/10.1007/s12665-014-3229-2)
- 707 Bouwer, H. (2002). Artificial recharge of groundwater: hydrogeology and engineering. *Hydrogeology Journal*,  
708 10(1), 121–142. <https://doi.org/10.1007/s10040-001-0182-4>
- 709 Braun, J.-J., Descloitres, M., Riotte, J., Fleury, S., Barbiéro, L., Boeglin, J.-L., ... Dupré, B. (2009). Regolith mass  
710 balance inferred from combined mineralogical, geochemical and geophysical studies: Mule Hole gneissic  
711 watershed, South India. *Geochimica et Cosmochimica Acta*, 73(4), 935–961.  
712 <https://doi.org/10.1016/j.gca.2008.11.013>
- 713 Bredehoeft, J. D. (2002). The Water Budget Myth Revisited: Why Hydrogeologists Model. *Ground Water*, 40(4),  
714 340–345. <https://doi.org/10.1111/j.1745-6584.2002.tb02511.x>
- 715 Brindha, K., Jagadeshan, G., Kalpana, L., & Elango, L. (2016). Fluoride in weathered rock aquifers of southern  
716 India: Managed Aquifer Recharge for mitigation. *Environmental Science and Pollution Research*, 23(9),  
717 8302–8316. <https://doi.org/10.1007/s11356-016-6069-7>
- 718 Carleton, G. B. (2010). *Simulation of Groundwater Mounding Beneath Hypothetical Stormwater Infiltration*  
719 *Basins. U.S. Geological Survey Scientific Investigations Report (Vol. 2010–5102)*. Retrieved from  
720 <https://pubs.usgs.gov/sir/2010/5102/support/sir2010-5102.pdf>
- 721 Central Ground Water Board. (2011). *Select Case Studies: Rain Water Harvesting and Artificial Recharge*.  
722 Ministry of Water Resources, Govt. of India.
- 723 Central Ground Water Board. (2013). *Master plan for artificial recharge to ground water in India*. Ministry of  
724 Water Resources, Govt. of India.
- 725 Chandra, S., Nagaiah, E., Kumar, D., Ahmeduddin, M., Raju, K., Mallesh, D., ... Ahmed, S. (2009). *Establishment*  
726 *of International Hydrogeological Park at Chautuppal, Nalgonda District, Andhra Pradesh Delineation of*  
727 *Aquifer Geometry using Electrical Resistivity Tomography: Phase-I. IFCGR Technical Report - Unpublished,*  
728 *27 pp.*

---

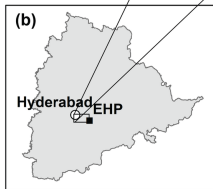
†Corresponding author. Email: madeleine.nicolas@univ-rennes1.fr

- 729 Cheepi, D. P. (2012). Musi River Pollution Its Impact on Health and Economic Conditions of Down Stream  
730 Villages-A Study. *IOSR Journal of Environmental Science, Toxicology and Food Technology*, 1(4), 40–51.  
731 <https://doi.org/10.9790/2402-0144051>
- 732 Chiang, W. H., & Kinzelbach, W. (1998). Processing Modflow. *A Simulation Program for Modelling Groundwater*  
733 *Flow and Pollution. User Manual*. Retrieved from  
734 <http://www.pmwin.net/programs/prevpm/pm4/doc/pmwin41.pdf>
- 735 Chilton, P. J., & Foster, S. S. D. (1995). Hydrogeological Characterisation And Water-Supply Potential Of  
736 Basement Aquifers In Tropical Africa. *Hydrogeology Journal*, 3(1), 36–49.  
737 <https://doi.org/10.1007/s100400050061>
- 738 Cook, P. G., Solomon, D. K., Sanford, W. E., Busenberg, E., Plummer, L. N., & Poreda, R. J. (1996). Inferring  
739 shallow groundwater flow in saprolite and fractured rock using environmental tracers. *Water Resources*  
740 *Research*, 32(6), 1501–1509. <https://doi.org/10.1029/96WR00354>
- 741 Cuthbert, M. O., & Tindimugaya, C. (2010). The importance of preferential flow in controlling groundwater  
742 recharge in tropical Africa and implications for modelling the impact of climate change on groundwater  
743 resources. *Journal of Water and Climate Change*, 1(4), 234–245. <https://doi.org/10.2166/wcc.2010.040>
- 744 de Marsily, G. (1986). *Quantitative hydrogeology*. Paris School of Mines, Fontainebleau.
- 745 de Marsily, G., Delay, F., Gonçalves, J., Renard, P., Teles, V., & Violette, S. (2005). Dealing with spatial  
746 heterogeneity. *Hydrogeology Journal*, 13(1), 161–183. <https://doi.org/10.1007/s10040-004-0432-3>
- 747 Dewandel, B., Lachassagne, P., & Krishnamurthy, N. S. (2006). A generalized 3-D geological and hydrogeological  
748 conceptual model of granite aquifers controlled by single or multiphase weathering. *Journal of*  
749 *Hydrology*, 260–284. <https://doi.org/10.1016/j.jhydrol.2006.03.026>
- 750 Dewandel, B., Maréchal, J. C., Bour, O., Ladouche, B., Ahmed, S., Chandra, S., & Pauwels, H. (2012). Upscaling  
751 and regionalizing hydraulic conductivity and effective porosity at watershed scale in deeply weathered  
752 crystalline aquifers. *Journal of Hydrology*, 416–417, 83–97. <https://doi.org/10.1016/j.jhydrol.2011.11.038>
- 753 Dillon, P. J., Gale, I., Contreras, S., Pavelic, P., Evans, R., & Ward, J. (2009). Managing aquifer recharge and  
754 discharge to sustain irrigation livelihoods under water scarcity and climate change. *IAHS-AISH Publication*,  
755 330(September), 1–12. Retrieved from [http://www.scopus.com/inward/record.url?eid=2-s2.0-](http://www.scopus.com/inward/record.url?eid=2-s2.0-78751661981&partnerID=tZOtx3y1)  
756 [78751661981&partnerID=tZOtx3y1](http://www.scopus.com/inward/record.url?eid=2-s2.0-78751661981&partnerID=tZOtx3y1)
- 757 Dillon, P. J., & Liggett, J. A. (1983). An ephemeral stream-aquifer interaction model. *Water Resources Research*,  
758 19(3), 621–626. <https://doi.org/10.1029/WR019i003p00621>
- 759 George, R. J. (1992). Hydraulic properties of groundwater systems in the saprolite and sediments of the  
760 wheatbelt, Western Australia. *Journal of Hydrology*, 130(1–4), 251–278. [https://doi.org/10.1016/0022-](https://doi.org/10.1016/0022-1694(92)90113-A)  
761 [1694\(92\)90113-A](https://doi.org/10.1016/0022-1694(92)90113-A)
- 762 Gleeson, T., Novakowski, K., & Kyser, K. T. (2009). Extremely rapid and localized recharge to a fractured rock  
763 aquifer. *Journal of Hydrology*, 376(3–4), 496–509. <https://doi.org/10.1016/j.jhydrol.2009.07.056>
- 764 Glendenning, C. J., van Ogtrop, F. F., Mishra, A. K., & Vervoort, R. W. (2012). Balancing watershed and local  
765 scale impacts of rain water harvesting in India—A review. *Agricultural Water Management*, 107, 1–13.  
766 <https://doi.org/10.1016/j.agwat.2012.01.011>
- 767 Government of Andhra Pradesh. (2003). *Andhra Pradesh Water Vision-Methods, Position papers, and District*  
768 *reports* (Vol. II). Water Conservation Mission, Government Insurance Building, Tilak road, Hyderabad,  
769 Andhra Pradesh.
- 770 Guihéneuf, N. (2014). *Structure des écoulements et propriétés de transport des aquifères cristallins fracturés et*  
771 *altérés : Application au site de Choutuppal (Inde du Sud)*. Université de Rennes 1.
- 772 Guihéneuf, N., Boisson, A., Bour, O., Dewandel, B., Perrin, J., Dausse, A., ... Maréchal, J. C. (2014). Groundwater  
773 flows in weathered crystalline rocks: Impact of piezometric variations and depth-dependent fracture  
774 connectivity. *Journal of Hydrology*, 511, 320–334. <https://doi.org/10.1016/j.jhydrol.2014.01.061>
- 775 Guihéneuf, N., Bour, O., Boisson, A., Le Borgne, T., Becker, M. W., Nigon, B., ... Maréchal, J. C. (2017). Insights

- 776 about transport mechanisms and fracture flow channeling from multi-scale observations of tracer  
777 dispersion in shallow fractured crystalline rock. *Journal of Contaminant Hydrology*, 206(May), 18–33.  
778 <https://doi.org/10.1016/j.jconhyd.2017.09.003>
- 779 Hamadeh, A. F., Sharma, S. K., & Amy, G. (2014). Comparative assessment of managed aquifer recharge versus  
780 constructed wetlands in managing chemical and microbial risks during wastewater reuse: a review.  
781 *Journal of Water Reuse and Desalination*, 4(1), 1–8. <https://doi.org/10.2166/wrd.2013.020>
- 782 Hantush, M. S. (1967). Growth and decay of groundwater-mounds in response to uniform percolation. *Water*  
783 *Resources Research*, 3(1), 227–234. <https://doi.org/10.1029/WR003i001p00227>
- 784 Harbaugh, A. W. (2005). MODFLOW-2005, The U. S. Geological Survey Modular Ground-Water Model — the  
785 Ground-Water Flow Process. *U.S. Geological Survey Techniques and Methods*, 253.
- 786 Irrigation & CAD Department. (2015). Mission Kakatiya - Mission. Retrieved August 9, 2017, from  
787 <http://missionkakatiya.cgg.gov.in/homemission>
- 788 Jakeman, A. J., Barreteau, O., Hunt, R. J., Rinaudo, J.-D., & Ross, A. (2016). *Integrated Groundwater*  
789 *Management*. (A. J. Jakeman, O. Barreteau, R. J. Hunt, J.-D. Rinaudo, & A. Ross, Eds.), *Integrated*  
790 *Groundwater Management*. Cham: Springer International Publishing. [https://doi.org/10.1007/978-3-319-](https://doi.org/10.1007/978-3-319-23576-9)  
791 [23576-9](https://doi.org/10.1007/978-3-319-23576-9)
- 792 Kacimov, A. R., Zlotnik, V., Al-Maktoumi, A., & Al-Abri, R. (2016). Modeling of transient water table response to  
793 managed aquifer recharge: a lagoon in Muscat, Oman. *Environmental Earth Sciences*, 75(4), 318.  
794 <https://doi.org/10.1007/s12665-015-5137-5>
- 795 Maréchal, J. C., Dewandel, B., & Subrahmanyam, K. (2004). Use of hydraulic tests at different scales to  
796 characterize fracture network properties in the weathered-fractured layer of a hard rock aquifer. *Water*  
797 *Resources Research*, 40(11), 1–17. <https://doi.org/10.1029/2004WR003137>
- 798 Maréchal, J. C., Wyns, R., Lachassagne, P., Subrahmanyam, K., & Touchard, F. (2003). Anisotropie verticale de la  
799 perméabilité de l'horizon fissuré des aquifères de socle : concordance avec la structure géologique des  
800 profils d'altération. *Comptes Rendus Geoscience*, 335(5), 451–460. [https://doi.org/10.1016/S1631-](https://doi.org/10.1016/S1631-0713(03)00082-8)  
801 [0713\(03\)00082-8](https://doi.org/10.1016/S1631-0713(03)00082-8)
- 802 Masciopinto, C. (2013). Management of aquifer recharge in Lebanon by removing seawater intrusion from  
803 coastal aquifers. *Journal of Environmental Management*, 130, 306–312.  
804 <https://doi.org/10.1016/j.jenvman.2013.08.021>
- 805 Massuel, S., Perrin, J., Mascré, C., Mohamed, W., Boisson, A., & Ahmed, S. (2014). Managed aquifer recharge in  
806 South India: What to expect from small percolation tanks in hard rock? *Journal of Hydrology*, 512, 157–  
807 167. <https://doi.org/10.1016/j.jhydrol.2014.02.062>
- 808 Mawer, C., Parsekian, A., Pidlisecky, A., & Knight, R. (2016). Characterizing Heterogeneity in Infiltration Rates  
809 During Managed Aquifer Recharge. *Groundwater*, 54(6), 818–829. <https://doi.org/10.1111/gwat.12423>
- 810 Nilesh, V. (2016). Hyderabad: Pollutants in Musi rise, river becomes sewage. Retrieved December 6, 2017, from  
811 [http://www.deccanchronicle.com/lifestyle/pets-and-environment/250716/hyderabad-pollutants-in-](http://www.deccanchronicle.com/lifestyle/pets-and-environment/250716/hyderabad-pollutants-in-musi-rise-river-becomes-sewage.html)  
812 [musi-rise-river-becomes-sewage.html](http://www.deccanchronicle.com/lifestyle/pets-and-environment/250716/hyderabad-pollutants-in-musi-rise-river-becomes-sewage.html)
- 813 Nimmo, J. R., Creasey, K. M., Perkins, K. S., & Mirus, B. B. (2017). Preferential flow, diffuse flow, and perching in  
814 an interbedded fractured-rock unsaturated zone. *Hydrogeology Journal*, 25(2), 421–444.  
815 <https://doi.org/10.1007/s10040-016-1496-6>
- 816 Nitin, B. (2018). A looming health threat in Hyd? How polluted Musi water finds its way back to the city. *The*  
817 *News Minute*. Retrieved from [https://www.thenewsminute.com/article/looming-health-threat-hyd-how-](https://www.thenewsminute.com/article/looming-health-threat-hyd-how-polluted-musi-water-finds-its-way-back-city-87275)  
818 [polluted-musi-water-finds-its-way-back-city-87275](https://www.thenewsminute.com/article/looming-health-threat-hyd-how-polluted-musi-water-finds-its-way-back-city-87275)
- 819 Palma, A., González, F., & Cruickshank, C. (2015). Managed Aquifer Recharge As a Key Element in Sonora River  
820 Basin Management, Mexico. *Journal of Hydrologic Engineering*, 20(3), B4014004.  
821 [https://doi.org/10.1061/\(ASCE\)HE.1943-5584.0001114](https://doi.org/10.1061/(ASCE)HE.1943-5584.0001114)
- 822 Perrin, J., Ahmed, S., & Hunkeler, D. (2011). The effects of geological heterogeneities and piezometric  
823 fluctuations on groundwater flow and chemistry in a hard-rock aquifer, southern India. *Hydrogeology*

- 824 *Journal*, 19(6), 1189–1201. <https://doi.org/10.1007/s10040-011-0745-y>
- 825 Perrin, J., Mascré, C., Massuel, S., & Ahmed, S. (2009). Tank management in Andhra Pradesh, India: Percolation  
826 versus irrigation. In *IAHS-AISH Publication* (Vol. 330, pp. 28–33). Retrieved from  
827 [https://www.scopus.com/inward/record.uri?eid=2-s2.0-  
828 78751652998&partnerID=40&md5=af84aa4600d3e9200c3c29c950020](https://www.scopus.com/inward/record.uri?eid=2-s2.0-78751652998&partnerID=40&md5=af84aa4600d3e9200c3c29c950020)
- 829 Pingali, P. L. (2012). Green Revolution: Impacts, limits, and the path ahead. *Proceedings of the National  
830 Academy of Sciences*, 109(31), 12302–12308. <https://doi.org/10.1073/pnas.0912953109>
- 831 Pinstруп-Andersen, P., & Hazell, P. (1985). The impact of the Green Revolution and prospects for the future.  
832 *Food Reviews International*. Retrieved from  
833 <http://www.tandfonline.com/doi/pdf/10.1080/87559128509540765>
- 834 Planning Commission (Government of India). (2011). *Mid-Term Appraisal Eleventh Five Year Plan 2007–2012*.  
835 New Delhi: Oxford University Press.
- 836 Racz, A. J., Fisher, A. T., Schmidt, C. M., Lockwood, B. S., & Huertos, M. L. (2012). Spatial and Temporal  
837 Infiltration Dynamics During Managed Aquifer Recharge. *Groundwater*, 50(4), 562–570.  
838 <https://doi.org/10.1111/j.1745-6584.2011.00875.x>
- 839 Reddy, D. V., Nagabhushanam, P., Sukhija, B. S., & Reddy, A. G. S. (2009). Understanding hydrological processes  
840 in a highly stressed granitic aquifer in southern India. *Hydrological Processes*, 23(9), 1282–1294.  
841 <https://doi.org/10.1002/hyp.7236>
- 842 Réfloch, A. (2018). *Compréhension expérimentale et numérique des chemins de l'eau sur le champ captant de la  
843 Métropole de Lyon*. Université Grenoble Alpes.
- 844 Ronayne, M. J., Gorelick, S. M., & Caers, J. (2008). Identifying discrete geologic structures that produce  
845 anomalous hydraulic response: An inverse modeling approach. *Water Resources Research*, 44(8), 1–16.  
846 <https://doi.org/10.1029/2007WR006635>
- 847 Roques, C., Bour, O., Aquilina, L., & Dewandel, B. (2016). High-yielding aquifers in crystalline basement: insights  
848 about the role of fault zones, exemplified by Armorican Massif, France. *Hydrogeology Journal*, 24(8),  
849 2157–2170. <https://doi.org/10.1007/s10040-016-1451-6>
- 850 Saha, D., Dwivedi, S. N., Roy, G. K., & Reddy, D. V. (2013). Isotope-based investigation on the groundwater flow  
851 and recharge mechanism in a hard-rock aquifer system: the case of Ranchi urban area, India.  
852 *Hydrogeology Journal*, 21(5), 1101–1115. <https://doi.org/10.1007/s10040-013-0974-3>
- 853 Scanlon, B. R., Faunt, C. C., Longuevergne, L., Reedy, R. C., Alley, W. M., McGuire, V. L., & McMahon, P. B.  
854 (2012). Groundwater depletion and sustainability of irrigation in the US High Plains and Central Valley.  
855 *Proceedings of the National Academy of Sciences*, 109(24), 9320–9325.  
856 <https://doi.org/10.1073/pnas.1200311109>
- 857 Sharma, B. R., Villholth, K. G., & Sharma, K. D. (2005). Groundwater Research and Management: Integrating  
858 Science into Management Decisions. In B. R. Sharma, K. G. Villholth, & K. D. Sharma (Eds.), *Groundwater  
859 Governance in Asia Series - 1*. IWA Publishing.
- 860 Singh, J., Awasthi, M. K., & Sharma, R. K. (2004). Quantification of percolation from percolation Tank. *J. Soil  
861 Wat. Conserv. India*, 3 (3&4), 128–132.
- 862 Singh, R. (2000). Environmental consequences of agricultural development: a case study from the Green  
863 Revolution state of Haryana, India. *Agriculture, Ecosystems & Environment*, 82(1–3), 97–103.  
864 [https://doi.org/10.1016/S0167-8809\(00\)00219-X](https://doi.org/10.1016/S0167-8809(00)00219-X)
- 865 Sørensen, K. I., & Auken, E. (2004). SkyTEM - a new high-resolution helicopter transient electromagnetic  
866 system. *Exploration Geophysics*, 35(3), 194. <https://doi.org/10.1071/EG04194>
- 867 Spanoudaki, K., Paschalinos, Y., Memos, C. D., & Stamou, A. I. (2010). Analytical Solution to the Stream-Aquifer  
868 Interaction Problem: A Critical Review. *Global Nest Journal*, 12(2), 126–139.
- 869 Srivastava, R. C., Kannan, K., Mohanty, S., Nanda, P., Sahoo, N., Mohanty, R. K., & Das, M. (2009). Rainwater  
870 Management for Smallholder Irrigation and its Impact on Crop Yields in Eastern India. *Water Resources*

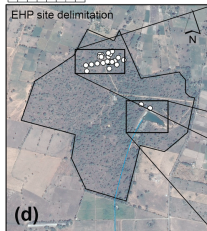
- 871            *Management*, 23(7), 1237–1255. <https://doi.org/10.1007/s11269-008-9324-y>
- 872 St. Clair, J., Moon, S., Holbrook, W. S., Perron, J. T., Riebe, C. S., Martel, S. J., ... Richter, D. d. (2015). Geophysical  
873 imaging reveals topographic stress control of bedrock weathering. *Science*, 350(6260), 534–538.  
874 <https://doi.org/10.1126/science.aab2210>
- 875 Sukhija, B. S., Reddy, D. V., Nagabhushanam, P., & Hussain, S. (2003). Recharge processes: piston flow vs  
876 preferential flow in semi-arid aquifers of India. *Hydrogeology Journal*, 11(3), 387–395.  
877 <https://doi.org/10.1007/s10040-002-0243-3>
- 878 UNESCO. (1999). *Water resources of hard rock aquifers in arid and semi-arid zones*. (J. W. Lloyd, Ed.). Paris,  
879 France: United Nations.
- 880 Ward, J., & Dillon, P. J. (2012). Principles to coordinate managed aquifer recharge with natural resource  
881 management policies in Australia. *HYDROGEOLOGY JOURNAL*, 20(5), 943–956.  
882 <https://doi.org/10.1007/s10040-012-0865-z>
- 883 Warner, J. W., Molden, D., Chahata, M., & Sunada, D. K. (1989). Mathematical analysis of artificial recharge  
884 from basins. *Journal of the American Water Resources Association*, 25(2), 401–411.  
885 <https://doi.org/10.1111/j.1752-1688.1989.tb03077.x>
- 886 Zhou, Y. (2009). A critical review of groundwater budget myth, safe yield and sustainability. *Journal of*  
887 *Hydrology*, 370(1–4), 207–213. <https://doi.org/10.1016/j.jhydrol.2009.03.009>
- 888



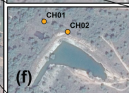
0 1 2 4 6 8 Km



0 0.1 0.2 0.4 Km

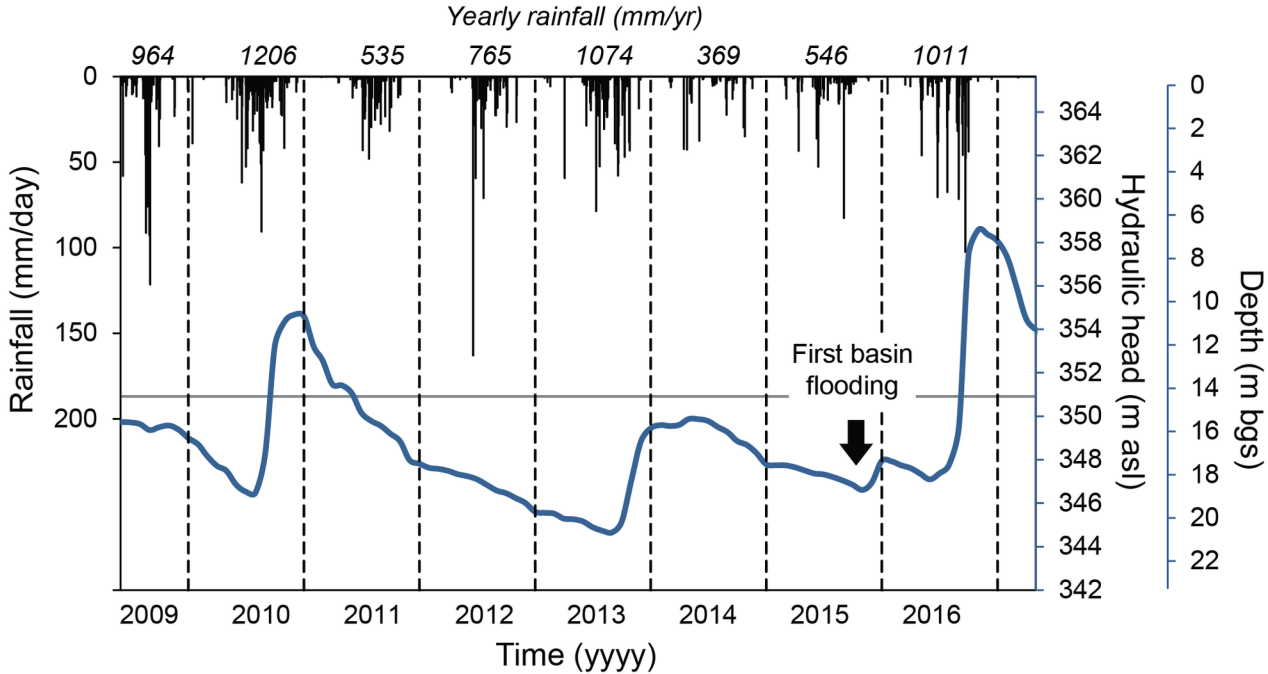


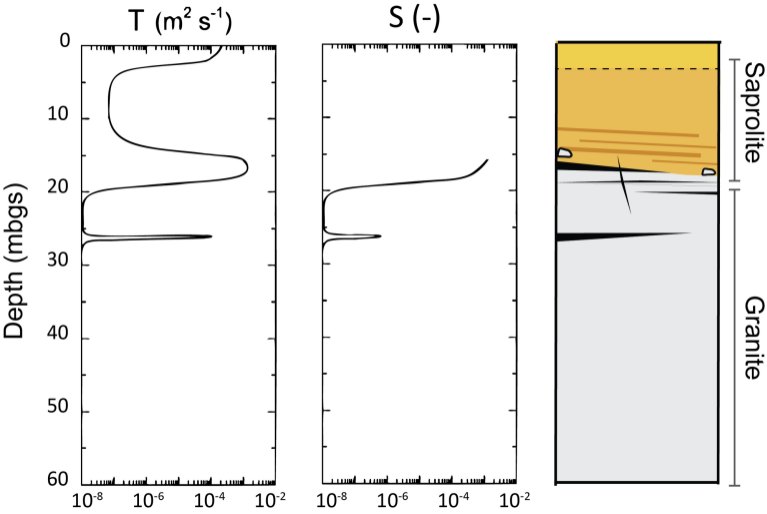
0 15 30 60 m

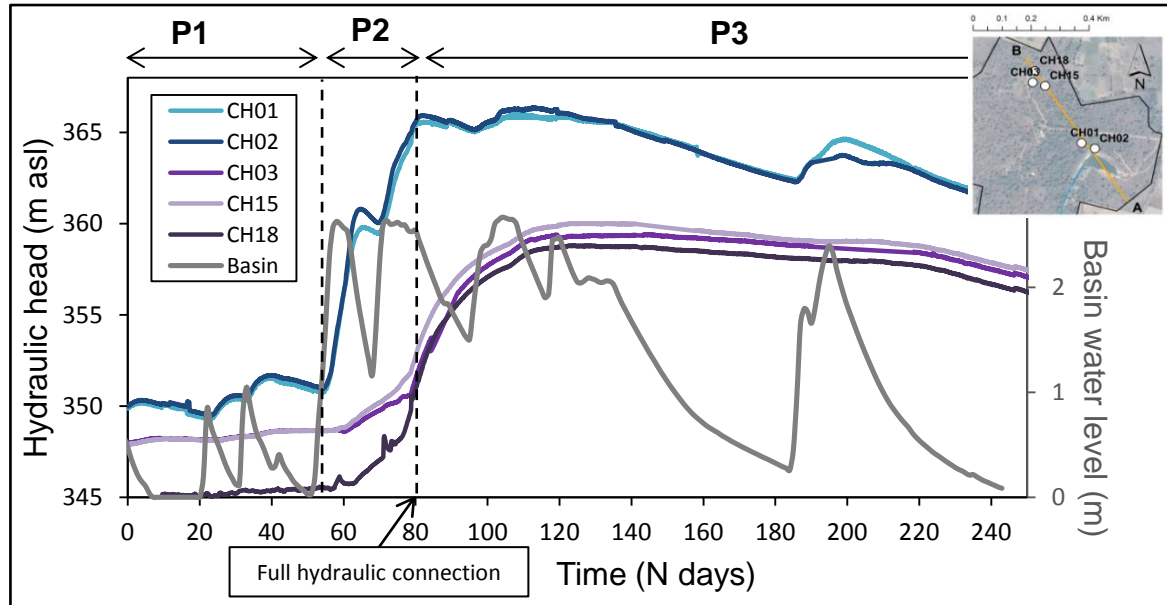


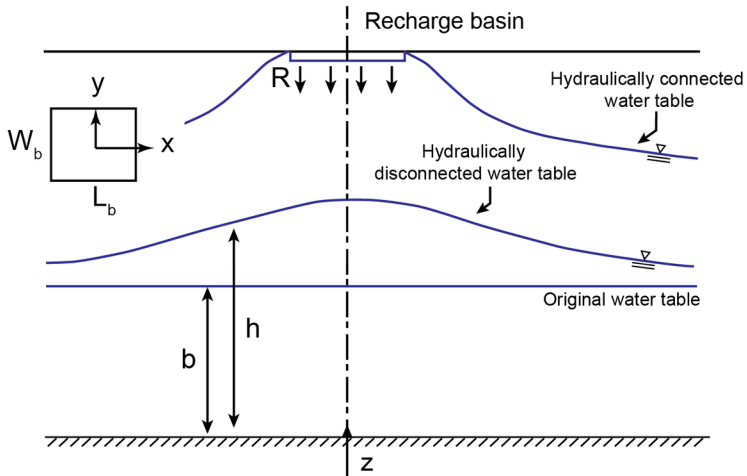
● Monitored boreholes  
○ Boreholes

0 35 70 m









$h$  = Water table elevation above the base of the aquifer (L)

$b$  = Original saturated thickness (L)

$\bar{b}$  = Average saturated thickness (L)

$L_b$  = Length of rectangular recharge basin (L)

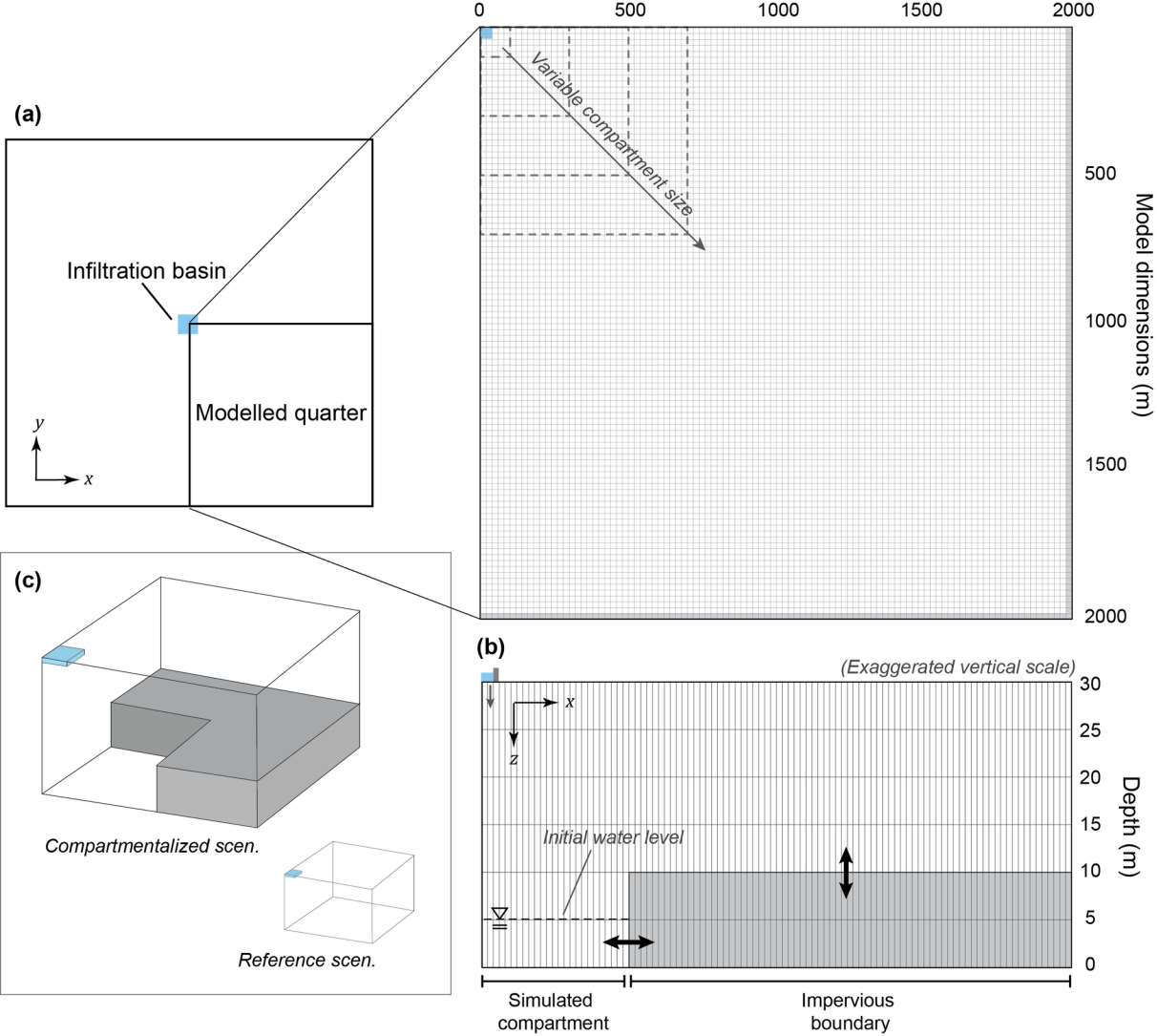
$W_b$  = Width of rectangular recharge basin (L)

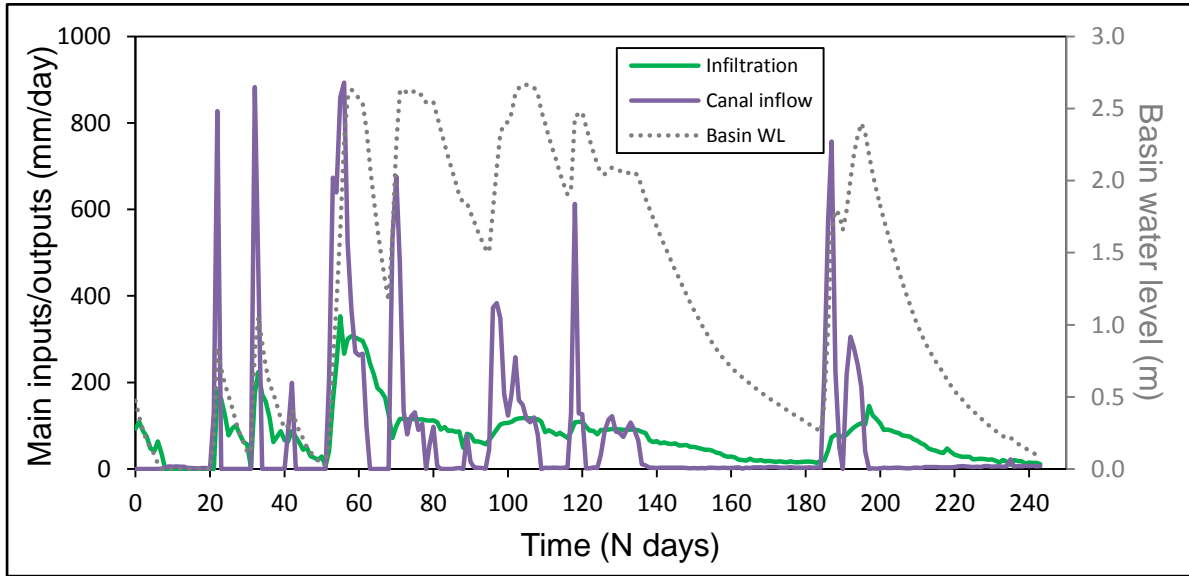
$R$  = Recharge rate (L/T)

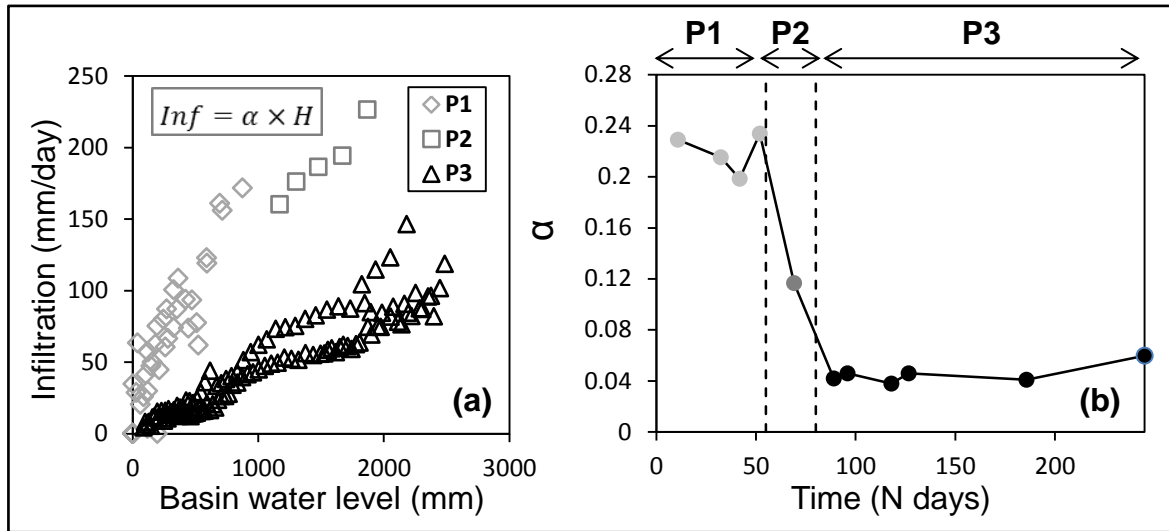
$K$  = Horizontal hydraulic conductivity (L/T)

$S$  = Storage coefficient (dimensionless)

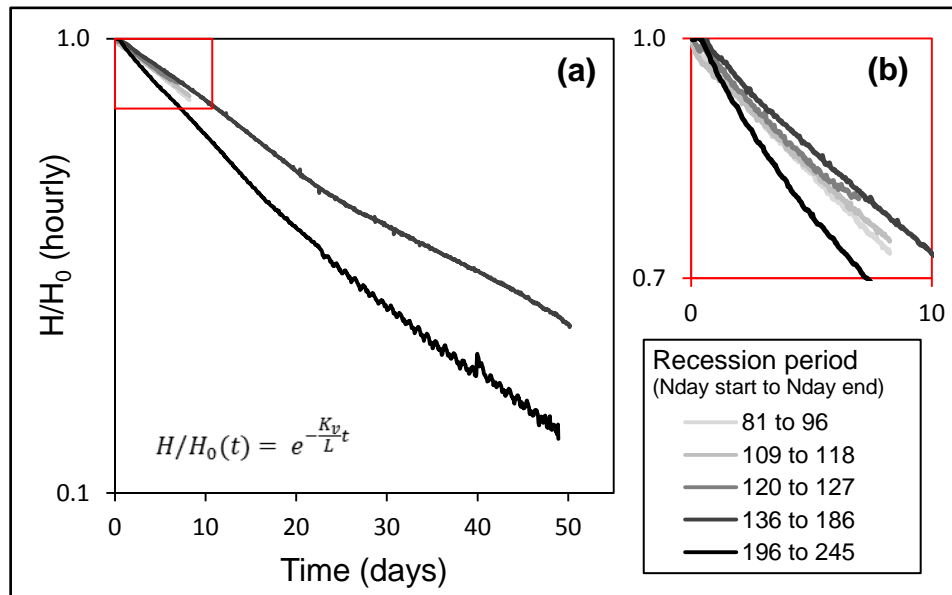
$x, y$  = Cartesian coordinates with center of recharge basin as origin (L)

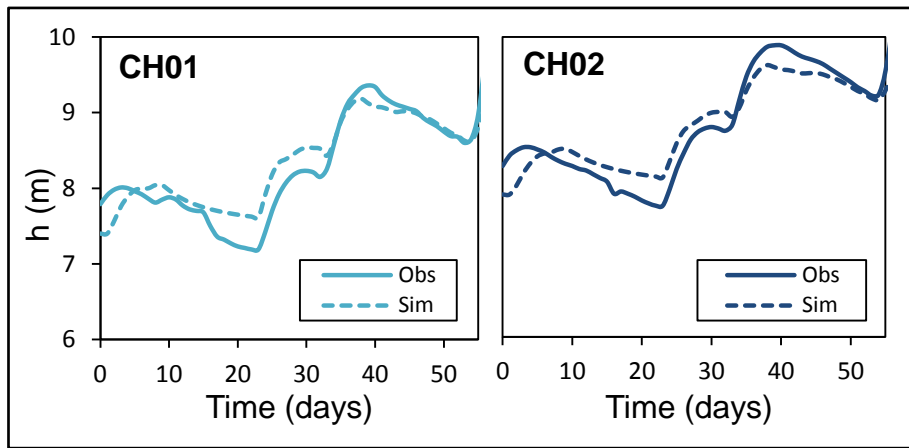


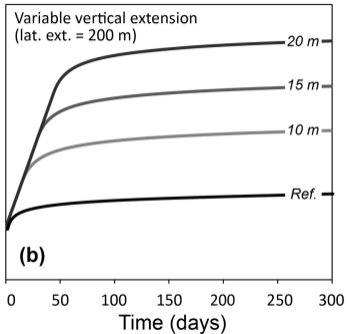
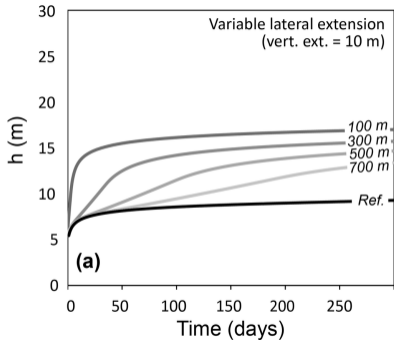


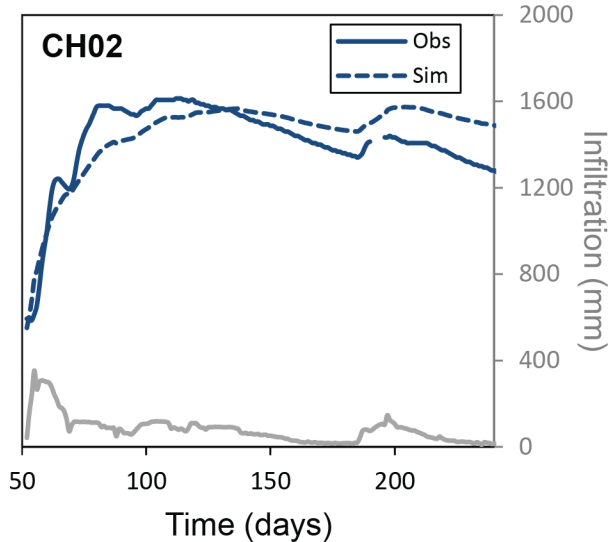
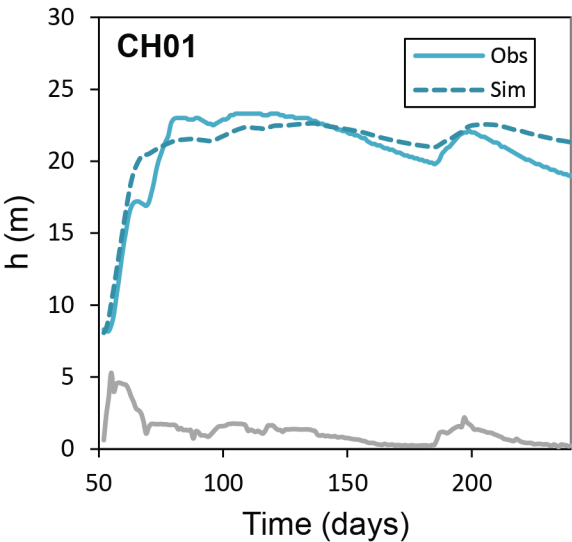


0	0.28
55	55
80	80

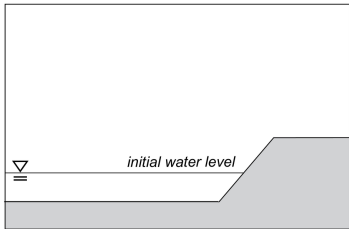




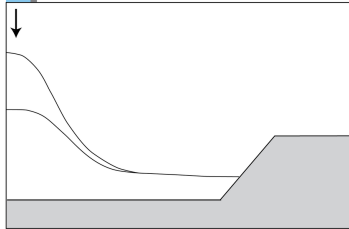




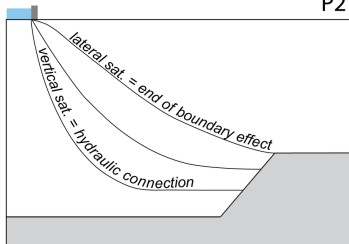
t = 0



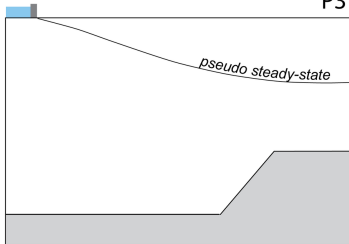
P1

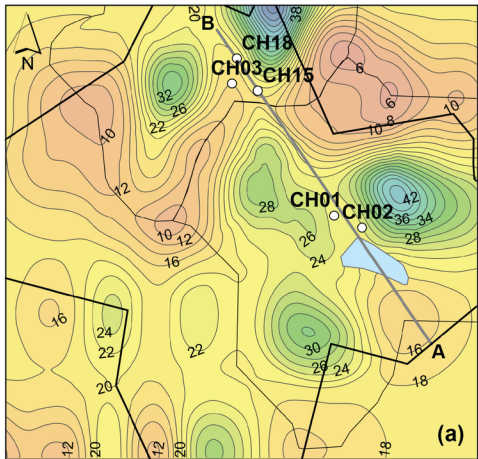


P2



P3





Interface relief depth (m bgs)

