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1 Seasonal influences on groundwater arsenic concentrations in the irrigated region of the Cambodian
2 Mekong Delta.

3

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5 Tweed S.^{1*}, Massuel S.¹, Seidel JL.², Chhuon K.³, Lun S.³, Eang KE.³, Venot JP.¹, Belaud G.¹, Babik M.⁴,
6 Leblanc M.⁴

7

8 ¹UMR G-eau, IRD, SupAgro, Montpellier, France

9 ²UMR HSM, University of Montpellier, CNRS, Montpellier, France

10 ³ITC, Phnom Penh, Cambodia

11 ⁴UMR EMMAH, Hydrogeology Laboratory, University of Avignon, France

12

13 *Corresponding author: sarah.tweed@ird.fr

14

15 **Abstract**

16

17 Similar to many southern and southeast Asian regions, the mobilisation of arsenic (As) from
18 sediments has driven a widespread contamination problem for groundwater resources in the
19 Cambodian Mekong Delta. For the first time, the seasonal changes in As concentrations and
20 potential links to groundwater pumping for irrigation in shallow aquifers of the Cambodian Mekong
21 Delta are investigated. Using environmental tracers ($\delta^{18}\text{O}$, $\delta^2\text{H}$, ^3H , major/trace ions and rare earth
22 elements) the natural and pumping-induced changes in hydrogeological processes are identified.
23 Three conceptual models are proposed: Model 1, where there is limited local recharge or low
24 recharge rates (^3H mean residence time > 60 years) and groundwater has a large range in As
25 concentrations (0.2 to 393.8 $\mu\text{g/L}$). In this semi-confined aquifer, only one of the six groundwater
26 sites has As concentrations that increase (by 10.9 $\mu\text{g/L}$) potentially due to groundwater pumping and
27 resultant mixing with high-As and low $(\text{Pr}/\text{Sm})_{\text{NASC}}$ groundwater. However, data on groundwater
28 extraction volumes is required to verify the link with irrigation practices. Model 2, where
29 groundwater is recharged by evaporated surface waters (fractionated $\delta^{18}\text{O}$ and $\delta^2\text{H}$). There are
30 moderate As concentrations (64.1-106.1 $\mu\text{g/L}$) but no significant seasonal changes even though the
31 recharging waters have relatively greater organic carbon contents during the dry season (reduced
32 Ce/Ce* anomaly). Finally model 3, where groundwater is significantly recharged by wet season
33 rainfall ($\sim 50\%$ from $\delta^{18}\text{O}$ data). There is a minor increase in As concentrations with recharge (by 6.
34 $\mu\text{g/L}$). These combined results highlight an aquifer system in the irrigated region of the Cambodian
35 Mekong Delta where As concentrations are largely impacted by natural rather than irrigation
36 processes. Seasonal-scale recharge processes control As processes where the aquifer is not confined
37 by shallow clay layers, and where the aquifer is semi-confined As concentrations largely reflect
38 longer-term natural processes.

39

40 Keywords: groundwater quality, groundwater resources, irrigation, arsenic.

41 **1. Introduction**

42

43 The impact of irrigation practices on the contamination of groundwater by arsenic (As) is a globally
44 recognised phenomenon. Whilst in many cases As is derived from natural sources, it is well
45 established that groundwater pumping can cause changes to the mixing and biogeochemical
46 processes, and thereby accentuate the As issue in groundwater resources. The extraction of
47 groundwater leading to changes in As concentrations has reportedly occurred in the United States
48 (e.g. Ayotte et al., 2011; Smith et al., 2018), Vietnam (Berg et al., 2008), and Taiwan (Liu et al., 2003).

49

50 The main aquifer of the Cambodian Mekong Delta is viewed as a groundwater system that remains
51 under relatively 'natural' conditions in terms of the extent and timing of groundwater pumped for
52 irrigation (e.g. Polya et al., 2005; Polizzotto et al., 2008; Lawson et al., 2013; 2016). Therefore, the
53 effects of groundwater pumping for irrigation on the As crisis in the Cambodian Mekong Delta has
54 not received the same attention compared to other southern and southeast Asia regions. For
55 example, in Bangladesh where the consequences of As contamination on public health are the most
56 severe worldwide (e.g. Raessler, 2018) there are several models proposed for the As release in
57 groundwater in irrigation regions. Many studies have hypothesised that irrigation withdrawals of
58 groundwater are the cause of elevated As concentrations in aquifers (e.g. Harvey et al., 2002). It is
59 suggested that pumping promoted As mobilisation in groundwater through the increase in hydraulic
60 gradients that allow the transfer of young surficial organic matter to reach greater depths in the
61 aquifer compared with 'natural' conditions (Neumann et al., 2010). Radloff et al. (2017) also found
62 that the sorption and desorption processes within the aquifer during pumping can rapidly affect As
63 mobilisation in groundwater, and for the aquifer sands in the Bengal Basin this was not highly
64 sensitive to changes in the As redox state. In comparison, Mailloux et al. (2013) analysed a site in
65 Bangladesh where reductions in hydraulic heads due to pumping and resultant increases in recharge

66 did not significantly influence groundwater redox, and it was concluded that the As mobilisation was
67 due to natural long-term processes pre-dating human perturbations.

68

69 Despite lower groundwater extraction rates in the Cambodian Mekong Delta compared with other
70 As-affected regions of southeast Asia (Lawson et al., 2013), many farmers and rural inhabitants are
71 dependent on groundwater resources, particularly during the dry season. Groundwater from shallow
72 aquifers in the upper delta region represents a strategic resource to people who cannot rely on
73 surface water, for both supplementary domestic use or for irrigation. Land irrigated by groundwater
74 in this region has reportedly increased by ~10% per year over the period 2003-2015 (Erban and
75 Gorelick, 2016). In the Vietnam Mekong Delta, a study found that the pumping of deep confined
76 aquifer systems causes inter-bedded clays to compact and release either As into solution, or release
77 dissolved organic carbon and competing ions that help to mobilise As (Erban et al., 2013). However,
78 this was an investigation of a deep aquifer system and to the best of our knowledge, the irrigation
79 pumping impacts on As in shallow aquifers of the Cambodian Mekong Delta have not yet been
80 documented.

81

82 In the Cambodian Mekong Delta, many previous studies have established that the major As
83 mobilisation processes in the shallow aquifers include the microbial decomposition of organic
84 matter and the reductive dissolution of arsenic-bearing iron minerals (Fe(III) oxyhydroxides) under
85 anaerobic conditions (Pederick et al., 2007; Rowland et al., 2007; Polizzotto et al., 2008; Quicksall et
86 al., 2008; van Dongen et al., 2008; Lawson et al., 2013; 2016; Richards et al., 2017a; 2019). However,
87 there are many different models that describe the environmental context of these processes. For
88 example, since labile organic carbon is essential for the mobilisation of As in aquifers, there have
89 been many studies devoted to analysing the sources and transfers of this carbon. In the Cambodian
90 Mekong Delta, studies have highlighted the multiple sources of organic carbon from both in-aquifer
91 sources (van Dongen et al., 2008) and surficial sources (Lawson et al., 2013; 2016; Richards et al.,

92 2019), which is used in the reductive dissolution of arsenic-bearing iron minerals (Fe(III)
93 oxyhydroxides) that mobilises As in groundwater. The release of As has also found to be slower
94 where the source of organic matter is from sediments that are older and deeper, compared with
95 recently recharged near-surface organic carbon (Polizzotto et al., 2008). Therefore, the
96 hydrodynamics of an aquifer, and modifications thereof, can significantly influence the rates of As
97 mobilisation in groundwater.

98

99 In this study, we build on the extensive As investigations already undertaken in the Cambodian
100 Mekong Delta (see Richards et al., 2019), by providing the first conceptual model of the seasonal-
101 changes in As concentrations in the shallow aquifer of the irrigated region. Groundwater from
102 typical rural sites was analysed where the resource is used for both irrigation and domestic use
103 purposes. In this study we investigate shallow groundwater along a 40 km length of the Bassac River,
104 the river that stems from the Mekong mainstream in Phnom Penh at the apex of the delta. There are
105 currently no groundwater monitoring bores or on-going measurements of hydraulic heads in this
106 region, therefore we use environmental tracers (stable isotopes, major ions, trace elements, rare
107 earth elements and tritium) to help identify key hydro(geo)logical processes. The study objectives
108 are to analyse (i) the origins of groundwater pumped for water supplies; (ii) the seasonal-scale
109 changes in groundwater chemistry; and (iii) whether or how the As concentrations seasonally
110 change during the dry season when groundwater is used for irrigation.

111

112

113 **2. Study Area**

114

115 The study area is located in the Kandal province of the Cambodian Mekong Delta, situated between
116 Phnom Penh and the border with Vietnam (Fig. 1). The tropical monsoon climate in this delta is

117 characterised by high wet season rainfall volumes derived from southwesterly weather systems, and a
118 dry season that is dominated by northeasterly weather systems (Briese, 1996).

119

120 **2.1 Hydro(geo)logy and aquifer geology**

121 South of Phnom Penh the Mekong River diverges to form the Bassac River that runs in parallel with
122 the Mekong River until they both discharge to the South China Sea in Vietnam. In the Cambodian
123 Mekong Delta, the rivers have high seasonal variations in water levels due to the large seasonal
124 differences in rainfall levels between the wet and dry seasons. For example, the Bassac River has
125 fluctuations in water levels of up to 6.6 m (between 1990-2018; Fig. 2a). Due to the flat topography
126 of the delta there are large areas that are annually inundated during the wet season rise of the river
127 water levels. Additional influences to the river water levels include tidal changes, groundwater
128 baseflow during the dry season (e.g. Rasmussen and Bradford, 1977; Polizzotto et al., 2008), and
129 river recharge to the aquifer during the wet season (e.g. Richards et al., 2018). In the study area, it
130 remains unclear whether the Bassac River interacts with shallow groundwater along its length, or
131 whether this is a spatially variable phenomenon limited by the thickness and continuity of the
132 underlying confining clay-rich sediments (as is the case in the Vietnamese section of the Mekong
133 Delta). The Bassac River has also been deepened in recent years due to river-bed mining (e.g.
134 Bravard et al., 2013), which has caused pools of up to 15 m deeper than the natural level (Brunier et
135 al., 2014; Anthony et al., 2015).

136

137 The sediments underlying the Bassac River are comprised of Quaternary alluvial sediments
138 deposited under river and marine environments. These sediments form stratified deltaic layers of
139 sand, gravel, clay and silt, and are regionally divided into two groups: the old (Pleistocene) and
140 young (Holocene) alluvium. The younger Holocene alluvium outcrops across the low elevation region
141 of the delta (<10 m ASL). In the study area this younger alluvium is the clay-rich confining layer that
142 can reach a thickness of up to ~18 m (Rasmussen and Bradford, 1977); the thickness increases

143 towards the Vietnamese coast reaching a maximum of > 100 m (Anderson, 1978). Underlying the
144 younger alluvium is the older Pleistocene alluvium, which in comparison contains less clay, coarser
145 material (sand, gravel, pebbles), and is therefore the major shallow aquifer. In the Cambodian
146 section of the delta the aquifer can be up to 100-120 m deep, whereas in the lower delta in Vietnam
147 the thickness reaches up to 450 m (Briese, 1996). Bedrock underlies and borders these alluvial
148 sediments, and in cases such as the basalt intrusions the bedrock is also interbedded with the old
149 alluvium. The bedrock is comprised of igneous, metamorphic and consolidated sedimentary rocks
150 that range in age from Precambrian to Plio-Pleistocene (Anderson, 1978). Underneath the Bassac
151 River the basement bedrocks are over 160 m from the surface (JICA, 2002).

152

153 Regionally groundwater flow in the delta is towards the south (Briese, 1996). Although groundwater
154 is an important water resource in the delta there are very few bores monitored for hydraulic head
155 data. Previous studies have identified that groundwater levels are relatively low in June and July, and
156 increase by ~5m in October and November (JICA, 2002). In the northern region of the Kandal
157 Province previous studies have highlighted that groundwater close to the river is heavily influenced
158 by river water level fluctuations, with the river's influence on groundwater levels decreasing with
159 distance and depth (Polizzotto et al., 2008; Richards et al., 2017a). However, the spatio-temporal
160 variations in groundwater elevations throughout the Cambodian Mekong Delta, particularly due to
161 groundwater pumping, are very poorly documented. The clay-rich sediments of the younger
162 alluvium forms a confining layer at the surface resulting in the underlying old alluvial aquifer acting
163 as a semi-confined system.

164

165 As in many developing countries, in both semi-arid and wet tropical environments, groundwater
166 resources in the Cambodian Mekong Delta are essential to rural communities (e.g. about 3,000
167 private wells have been identified in the study area by the Water Sanitation Program (WSP, 2019)).
168 Although groundwater resources are of poorer quality compared with river water in terms of

169 electrical conductivity and arsenic (see below results from this study), it is an unmonitored water
170 resource, and there is a cost associated with installing groundwater bores, many rural inhabitants
171 rely on accessing groundwater for both agricultural and domestic (including drinking) uses.

172

173

174 **2.2 Irrigation**

175 This region of the Mekong Delta is dedicated to farming that is heavily dictated by the wet and dry
176 seasons and the topography. Land along the river banks is slightly elevated then gently slopes down
177 in a perpendicular direction towards low-lying wetlands located further away from the main river
178 streams. Close to river banks, farmers can cultivate yearlong; orchards and vegetables dominate the
179 landscape and are irrigated especially during the dry season (Feb-May). Further away from the river,
180 farmers cultivate up to two crops of short term rice (e.g. 90-95 days crop). The first season starts
181 when the flood recedes (in November/December; this is locally called recession rice) and is mainly
182 rainfed though supplementary irrigation is common. For the second season, the cropping calendar
183 differs based on topography and water availability. Some farmers (low lying, more water available,
184 and higher vulnerability to floods) cultivate and irrigate rice during the dry season (Feb-May; this is
185 locally called dry-season rice), while others (slightly more elevated, less water available but less
186 vulnerable to floods) start cultivating rice in May with the first rains and harvest before the floods
187 come towards the end of July (supplementary irrigation is still needed; this is locally called early wet
188 season rice; Pillot, 2007).

189

190 In support of the agricultural activities in the study area there is a dense network of earthen artificial
191 channels locally call preks, acting as drainage and irrigation canals (SOFRECO, 2019). The beds of
192 these channels were usually dug several meters below the land surface, perpendicularly to the
193 Bassac, on a length of about 4 km, and are spaced 0.5 to 1 km (Fig. 1b). During high river water flows
194 the channels divert the water and sediments inland, and can either link the Bassac River with other

195 natural river streams in the delta or the water and sediments culminate in the low lying wetlands.
196 The same channels drain water back to the river during low river flow periods. The canals are used
197 for transportation, fishing and are also a major source of irrigation water (with farmers installing
198 small diesel pumps to draw water from the canals to their fields).

199

200 There has been recent rehabilitation (including deepening) of some irrigation channels in this region
201 of the delta to improve the water control and availability. Some channels were equipped with gates,
202 managed by water user associations to regulate flow from or to the river. In some cases, floating
203 pumping stations were installed on the Bassac and water is pumped in the channel during dry
204 season. Despite these changes, many of the irrigation channels (60% in 2018; SCP, 2018) cannot
205 provide river water inflows all year round, especially during the driest months. Many farmers are
206 obliged to install their own groundwater bores in order to irrigate during the dry season and for
207 domestic purposes. Groundwater is used for the irrigation of vegetables, fruit tree farms and rice
208 fields, and is mostly pumped at the end of the dry season for irrigation (~ May/June, depending on
209 the year).

210

211 **2.3 Arsenic**

212 In the Cambodian Mekong Delta, groundwater has been used to irrigate rice fields since ~ 1984 and
213 in areas where groundwater is the primary source of irrigation water the concentration of As in the
214 soils is elevated, albeit with lower health risks compared with the higher toxicity of arsenic present
215 in rice cultivated in Bangladesh (Murphy et al., 2017). For example, in the Kandal province Murphy et
216 al. (2018) reported As concentrations in soils close to wells that were approximately double the
217 concentrations of soils further from the well. Although groundwater is an essential dry season water
218 resource, it is well established (e.g. Phan et al., 2013) that As in groundwater is a health threat to the
219 rural population in the Cambodian Mekong Delta. Accordingly, As has received a lot of national
220 attention (e.g. database of 59000 records of As in groundwater in Cambodia; WSP, 2019). However,

221 in part because of the lack of in-situ groundwater monitoring data, this As database has not been
222 analysed in combination with the modified hydro(geo)logical processes due to irrigation practices. In
223 addition, as discussed previously, the natural processes influencing As concentrations are already
224 numerous and complex; including the non-uniform distribution of the eroded Himalayan sediments,
225 competing electron acceptors, and the reactivity, age and transfers of organic carbon (Papacostas et
226 al., 2008; Quicksall et al., 2008; Lawson et al., 2016; Richards et al., 2017b). An additional potential
227 source of As in the study area is the use of fertilisers. However, a recent study in Kandal by Murphy
228 et al. (2018) calculated that As loads in groundwater used for irrigation were over 3000 times the As
229 loads present in the inorganic fertilisers being used.

230

231

232 **3. Methods**

233

234 In this study, a new As dataset is analysed in conjunction with hydrochemical tracers to inform us on
235 the hydro(geo)logical processes in areas where groundwater is used for irrigation. Here we analyse
236 river bathymetry data in conjunction with hydrochemical parameters to study the potential
237 interactions between the river and shallow groundwater. The hydrochemical tracers are also used
238 to identify hydrogeological processes affecting As concentrations. Field visits were conducted during
239 June and September in 2017 and 2018 (Fig. 2b). Samples in June represent the end of the dry
240 season/start of the wet season, and it is during June that we expect to see the full impacts of the dry
241 season pumping on shallow groundwater samples. In comparison, samples collected in September
242 represent the peak of the wet season.

243

244 **3.1 River bathymetry and flow**

245 River bathymetry data was used to map sections of the river bathymetry and flow velocity from one
246 bank to the other (back and forth). The ADCP was coupled with a differential GPS in order to map

247 the exact profile track perpendicular to the riverbed. Then, the entire river flow through the section
248 was inferred. During the dry season, an acoustic Doppler current profiler (ADCP, RDI™ 600 kHz) was
249 used to map sections of the river bathymetry and measure the river flow at the northern (R10) and
250 the southern (R1) sites. In addition, a hand held pressure sensor was used to record the river depth
251 (surface water level elevation minus water column) at each of the sampling sites along the length of
252 the Bassac River. The same protocol was repeated during the wet season, but a pressure sensor was
253 also used to map sections of the river bathymetry at three locations. The mean river bed minimum
254 elevation was about 5m below sea level, but deeper sections were observed in meanders. Such low
255 elevations of river depths are not uncommon in the study area, for example previous work by Lu et
256 al. (2014) found sections in the Mekong River up to 15m below the sea level. Water level variations
257 (Fig. 2) are monitored by Mekong River Commission on a daily basis at Koh Khel (11°16'8"N, 105°
258 1'40.5"E), and on hourly basis at Chau Doc (10°42'26"N, 105° 7'38"E) at the south of the study area.

259

260 **3.2 Hydrochemistry of surface water and groundwater**

261 The lack of physical hydrogeological time-series data in the study area necessitated the use of
262 environmental tracers (major ions, stable and radiogenic isotopes, and rare earth elements) to help
263 analyse the origins and transfers of groundwater. In 2017, 10 surface water sites (including water in
264 the prek channels used for irrigation or drainage water, water from low lying wetlands (floodwater),
265 and river water) were sampled during the dry and wet seasons, and in 2018 10 groundwater and 10
266 river sites along ~40 km of the Bassac River were sampled during the dry and wet seasons (Table 1).

267

268 The locations of the groundwater and surface water samples are presented in figure 1b and c. The
269 groundwater sites were selected to represent locations close to the river's edge; all groundwater
270 samples are located at distances of 0.3-1 km from the Bassac River's edge. In addition, the sites were
271 selected based on their location within irrigated areas; privately owned bores were sampled that are
272 actively used for domestic (including drinking water supplies) and/or irrigation purposes. The

273 groundwater samples were collected using the pumps already installed in the privately owned bores
274 that pumped groundwater from near the base of the bore. The depths for the groundwater bores
275 sampled range from 18-62 m and are presented in Table 1. The bores are screened along the length
276 of the casing. Therefore sampling depths will vary depending on pumping rates and changes in
277 hydraulic conductivity values of the aquifer with depth. Groundwater bores were pumped until the
278 EC stabilised, after which the field parameters were measured and samples were collected for
279 laboratory analysis. All river samples were collected at ~50 cm depth from the middle of the river
280 using a boat. For both river and groundwater samples the electrical conductivity (EC), pH and
281 temperature were measured in the field using the field-calibrated meter WTW 3320. Water samples
282 were collected in the field for major, minor and trace elements, stable isotopes and tritium.

283

284 The major and minor ions, and stable isotopes were analysed at the Hydrochemistry Laboratory,
285 University of Avignon, France. Samples for ion analyses were filtered in the field (0.45 µm cellulose
286 nitrate filter) and an aliquot was acidified to pH < 2 (ultrapure 16N HNO₃). Anions and cations were
287 analysed using ion chromatography (Dionex; ICS1100) and ion analysis uncertainty is in the order of
288 3 %. For ion analysis, the quality control methodology is performed by cross-calculations of
289 conductivity, ion balances and saturation indices for each dataset. In addition, inter-laboratory
290 analyses are regularly undertaken. For the concentrations of N species we used groundwater
291 sampled in the Vietnam Mekong Delta to compare values (i) when analysed locally within a couple of
292 days by standard colorimetric methods and (ii) when the filtered samples were analysed a few
293 weeks later in France at the University of Avignon (ion chromatography), which is the same
294 laboratory used in this study. The results varied slightly (differences in NH₄ ≤ 2.71 mg/L) and the
295 differences were linear (e.g. for NH₄ R² = 0.953); therefore we consider that the relative changes of
296 NH₄ between different groundwater samples in this study will still be valid. However, we use a
297 higher detection limit of 2 mg/L for NH₄ than the analytical detection limit associated with the ion
298 chromatography analysis (0.01 mg/L). Results of selected cations and anions for groundwater and

299 surface water are presented in Table 1. Charge balance errors for these waters are $\leq 5\%$. Stable
300 isotope values ($\delta^2\text{H}$ and $\delta^{18}\text{O}$) were analysed using the LGR (Los Gatos Research) laser spectrometer;
301 the uncertainty is $\delta^{18}\text{O} = \pm 0.1\text{‰}$ and $\delta^2\text{H} = \pm 1\text{‰}$ (Table 1).

302

303 Samples were also analysed for Tritium (^3H) contents (Table 1) by low level liquid scintillation
304 counting after electrolytic enrichment using a scintillation counter (Tri-Carb 3180 TR/SL) at the
305 Hydrochemistry Laboratory, University of Avignon, France. Calibration of the device was undertaken
306 before analysis of these water samples, during analysis there are control samples (of “dead water”)
307 in conjunction with regular inter-comparison with IAEA analysis and monitoring the enrichment
308 factor of each electrode. The results are returned to 2-sigma and the uncertainty associated with ^3H
309 results ranges from 0.3 to 0.6 TU (Table 1). For four groundwater samples there is ^3H present but the
310 values are low (0.8-1.2 TU), and for most groundwater samples the ^3H activities were below the
311 detection limit. For these samples below the detection limit, the ^3H value noted in Table 1 is \leq the 2-
312 sigma error associated with the sample analysis (e.g. ≤ 0.7 TU for G1 in June). Consequently,
313 although during the analysis the average counts per minute (cpm) of the June G1 sample was less
314 than the cpm of the “dead” ^3H water and the detection limit at 2σ , and the average cpm of the
315 September G1 sample was greater, due to the 2σ error value we cannot say that the ^3H value of \leq
316 0.7 TU in June is significantly different to 0.8 ± 0.3 TU in September.

317 The trace element and rare earth element samples were filtered ($0.2\ \mu\text{m}$) and acidified to $\text{pH} < 2$
318 (ultrapure 16N HNO_3) in the field. The samples were then analysed at the AETE-ISO platform OSU-
319 OREME at the University of Montpellier, France, with an ICAP Q THERMO SCIENTIFIC. Trace element
320 and rare earth element results are presented in Table 2. For the June groundwater trace element
321 data presented in Table 2 a repeat analysis was undertaken, and t-test p-values were > 0.05
322 indicating that the two analyses were not significantly different. For trace element analysis, all
323 chemicals used were of analytical grade (Merck Darmstadt, Germany). Ultrapure water was used
324 throughout the analysis. For calibration, multi-elemental standard solutions of metals were

325 prepared by dilution of a 1000 ppm certified standard solution (SCP Science, Canada) of
326 corresponding trace elements. Instrumental drift was monitored and corrected by addition of a
327 multi-elemental (Be, Sc, Ge, Rh, Ir) internal standard to each sample. Reagent and procedural blanks
328 were measured in parallel to sample treatment using identical procedures. Participation in
329 laboratory inter-comparison exercise was achieved for CRM SLRS-5 (Yeghicheyan et al., 2013) and
330 for SLRS-6 (Yeghicheyan et al., 2019) to complete the quality control methodology. The accuracy of
331 the data was tested by replicate analyses of the certified river water reference standard SLRS-6 for
332 trace elements (National Research Council, Canada). SLRS-6 was analysed every 10 samples to test
333 the analysis accuracy and precision. The SLRS-6 results are within the range of certified uncertainties
334 and deviations from the certified values were generally lower than 10% except for Pr (12%) and Tm
335 (13%). Repeated analysis of the CRM SLRS-6 several months apart makes it possible to assess the
336 reproducibility and the long-term precision of the analyses for each element (Table 3). The total
337 errors ($\sqrt{\text{sample error}^2 + \text{analytical error}^2}$) for As are presented in Table 2 and are used in the
338 comparison of the seasonal changes in As concentrations at each site.

339

340

341

342 **3.3 Geochemical modelling**

343 Two-component mixing models were used to determine (i) the relative recharge to the aquifer from
344 wet season rainfall at site G1, and (ii) the relative recharge of evaporated water compared with
345 rainwater for each of the groundwater sites. The first model used to calculate wet season rainfall
346 recharge at site G1 is based on those described in Clark and Fritz (1997) and uses equation (1):

$$347 F_r = (C_{\text{mix}} - C_{\text{gw}}) / (C_r - C_{\text{gw}}) \quad (1)$$

348 Where the fraction of rainfall contribution to the groundwater aquifer (F_r) at site G1 is calculated
349 using the concentrations of the tracer for groundwater from the aquifer before recharge (C_{gw}), the

350 groundwater after recharge (C_{mix}), and for the rain water (C_r). Calculations were undertaken using
351 the stable isotope $\delta^{18}O$ values.

352

353 The second mixing model, used to calculate the fraction of evaporated pond waters recharging
354 groundwater (F_{evap}) with equation (2), was also used by Richards et al. (2018) in the Kandal province
355 ~20 km north of our study area.

$$356 F_{evap} = (C_{gw} - C_r) / (C_{evap} - C_r) \quad (2)$$

357 The calculations use both $\delta^{18}O$ and δ^2H data. Equation 2 uses the isotope values of
358 groundwater (C_{gw}), end-member isotope values of rain (C_r) and end-member isotope values of
359 evaporated water (C_{evap}). The end-member isotope values for rainfall were those values used by
360 Richards et al. (2018) from 2014 rainfall data collected in Kandal ($\delta^{18}O = -10.6$, $\delta^2H = -69.5$ ‰). The
361 end-member isotope values for evaporated water were collected in this study from a drainage pond
362 ($\delta^{18}O = -0.81$, $\delta^2H = 18.44$ ‰). Results of this modelling are presented in Table 1 for all groundwater
363 sites. Where the modelled % of groundwater sourced from evaporated water are > 65%, these
364 indicate areas where the waters are more evaporated than mean rainfall values (Richards et al.,
365 2018).

366

367

368 **4. Results**

369

370 **4.1 River flow and depth profiles**

371 In the dry season, the river discharge was measured at 660 m³/s along the upstream profile and
372 470 m³/s downstream. In the wet season, the upstream flow was 2700 m³/s and 1650 m³/s
373 downstream. This means that the discharge was reduced by 30% and 40% along the 40-km length of
374 the Bassac River in the dry and wet season respectively. The discrepancy of the flow can be
375 attributed to water being discharged through the channels towards the floodplain. It can also be the

376 result of a net contribution from the river to the underlying aquifer in wet season (i.e. when the
377 hydraulic gradient is towards the aquifer). . The measurements showed that the bathymetry of the
378 streambed was heterogeneous with depths up to 13 m below ground surface at locations R1, R10
379 and between R4 and R5, creating possible connections between the river and the semi-confined
380 aquifer.

381

382

383 **4.2 Stable isotope values of groundwater and river water**

384 There does not yet exist an extensive database for stable isotopes of local rainfall in the lower
385 Mekong Delta (Cambodia and Vietnam). The LMWLs used in this study include data from 2014
386 rainfall in the Kandal province (Richards et al., 2018) that has a lower slope (6.4) compared with the
387 LMWL from long time-series rainwater data (1968-2015) collected in Bangkok (slope 7.3) located
388 ~500 km NW of the study area (GNIP station 4845500, elevation 2 mASL, 460 samples). Stable
389 isotope data of rainfall at Bangkok is considered representative of both inter-annual and seasonal
390 variations of rainfall in the lower Mekong Delta (Le Duy et al., 2017). The Bangkok LMWL has a slope
391 that is slightly lower (7.35) compared with the GMWL slope (8) from Craig (1961). In wet and dry
392 tropical zones the seasonal changes in atmospheric moisture transported over continental land
393 mass, temperatures and precipitation volumes result in wet and dry season differences in rainfall
394 stable isotope values (Aggarwal et al., 2004). The average stable isotope values for rainfall at
395 Bangkok during the peak wet season (September and October; 93 samples) are depleted compared
396 with peak dry season values (February and March; 63 samples) that show differences of 6.1 and 45
397 ‰ for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ respectively (Fig. 3). The Mekong River water sampled in the study region is also
398 influenced by rainfall from the northern area of the Mekong River Basin, where seasonal isotopic
399 differences are further pronounced and show more depleted values. For example, at Kunming (GNIP
400 station 5677800), located 300 km east of the Mekong River Basin in China at an elevation of 1892

401 mASL, the wet season averages for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ are depleted by 9.4 and 61 ‰ respectively
402 compared with the dry season values (data from 1999-2003; Fig. 3).

403

404 Results from this study show that with the exception of groundwater from sites G7 and G8, all
405 samples lie on the meteoric water lines and the dry season groundwater stable isotope values ($\delta^{18}\text{O}$
406 = -8.6 to -7.6; $\delta^2\text{H}$ = -60 to -53 ‰) are only slightly lower compared with dry season river values ($\delta^{18}\text{O}$
407 = -7.9 to -7.8; $\delta^2\text{H}$ = -55 to -54 ‰). During the wet season the groundwater values remain the same
408 as the dry season (maximum change of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ by 0.1 and 1 ‰ respectively), with the
409 exception of site G1 where there was a decrease of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ by 1.2 and 4 ‰ respectively. This
410 indicates either that groundwater and dry season river waters are recharged by similar rainfall
411 sources, or that there are significant inflows of groundwater from this aquifer to the river during the
412 dry season. The results also indicate that groundwater was recharged under similar conditions to the
413 modern day climate. In contrast to the constant groundwater isotope values, the river values
414 increase during the wet season by 0.4-0.6 and 3-5 ‰ for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ respectively. Since the
415 southern region of the Mekong River Basin (e.g. in Cambodia) generally has rainfall with higher
416 isotope values compared with northern region of the Mekong River Basin (e.g. in China), this wet
417 season increase in the river values may correspond with greater contributions from local rainfall.

418

419 Most groundwater shows no evidence of evaporation effects, with the exception of groundwater at
420 sites G7 and G8. Here there is fractionation from the meteoric water lines (Fig. 3) that may result
421 from either evaporation of the groundwater or mixing with evaporated water (Clark and Fritz, 1997).
422 The slope of the regression line for groundwater at these bores is 5.94, which is relatively high and
423 typical of evaporation in a highly humid (> 75%) environment (Clark and Fritz, 1997). The relative
424 extent of evaporation of groundwater calculated for G7 and G8 are higher (65-80 %) compared with
425 all other groundwater sites for both the wet and the dry seasons (≤ 32 %; Table 1). Much of the
426 delta area in Cambodia is naturally inundated during the wet season and artificially inundated during

427 the dry season from irrigation water. Stable isotope values of all inundation waters are presented in
428 figure 3, and those collected from drainage and irrigation prek waters during the dry season show
429 evaporated signatures with a similar slope (5.29) and values to the groundwater samples from bores
430 G7 and G8 (Fig. 3). In addition, the area is protected by a dyke, but in 2018 the water went through
431 and could not drain back. Therefore, groundwater at G7 and G8 is potentially recharged by
432 evaporated surface waters causing the shift in groundwater isotope values to the right of the MWL.
433 Recharge of evaporated surface waters is consistent with the tritium data (below) which indicates
434 there is recent recharge at these sites, and the major ion chemistry (below) that does not indicate
435 on-going evaporation controls on groundwater chemistry.

436

437

438 **4.3 Major ion composition of groundwater**

439 Most groundwater sampled has EC values between 86 to 1207 $\mu\text{S}/\text{cm}$, which increases with
440 borehole depth (Fig. 4a). At site G3 the borehole is deeper (62 m depth from surface) compared with
441 other samples, and has EC values ranging from 4110 to 4220 $\mu\text{S}/\text{cm}$. Despite the increase in rainfall,
442 inundation waters and high river water levels during the wet season, the wet season groundwater
443 EC values do not systematically decrease; only 3 groundwater sites show wet season decreases in EC
444 (by 44 - 506 $\mu\text{S}/\text{cm}$; G1, G6, G9). On the contrary, 3 groundwater sites show increases in EC (by 50 -
445 110 $\mu\text{S}/\text{cm}$; G2, G3, G5) and 3 sites show no significant seasonal change in EC ($< 10 \mu\text{S}/\text{cm}$; G4, G7,
446 G10). These results indicate the processes controlling groundwater EC are highly spatially and
447 temporally variable.

448

449 When compared with stable isotope values, there are 3 distinct end-members with varying EC values
450 (Fig. 4b). At site G1 groundwater is recharged by recent rainfall or inundation waters resulting in
451 isotope and EC values that decrease during the wet season (EC from 592 to 86 $\mu\text{S}/\text{cm}$). The second
452 end-member is from the deeper groundwater borehole at site G3; where isotope and EC values

453 show little seasonal variation (4110 to 4220 $\mu\text{S}/\text{cm}$). Thirdly, the sites G7 and G8 are where
454 groundwater is recharged by evaporated surface water although EC remains low (248 – 293 $\mu\text{S}/\text{cm}$).
455 From these end-members, there are 2 major mechanisms resulting in increases in groundwater EC;
456 (i) due to mixing with evaporated surface waters, and (ii) due to mixing with groundwater that has
457 high EC from water-rock interactions or evapotranspiration. In addition, river waters show increased
458 EC values and decreased isotope values during the dry season highlighting the increase in inflows
459 from groundwater (Fig. 4b).

460

461 Most of these circumneutral pH groundwater samples show some evidence of water-rock
462 interactions with silica concentrations that are up to 10 % of the TDS, and range from 2.9 – 29.4
463 (average = 14.4) mg/L SiO_2 (Table 1). The water-rock interactions results in groundwater
464 characterised by HCO_3 -Ca-Mg for most groundwater samples with EC values < 700 $\mu\text{S}/\text{cm}$, except for
465 bore G10 which has notably higher concentrations of SO_4 (457 to 473 mg/L) compared with other
466 ions. Where groundwater has higher EC values (4110 - 4220 $\mu\text{S}/\text{cm}$; G3) there are relative increases
467 in Na and Cl concentrations compared to other major ions.

468

469 Nitrogen in the groundwaters in the study area is mostly < 2mg/L, but there are three groundwater
470 sites that have relatively high ammonium concentrations ($\text{NH}_4 = 8\text{-}23$ mg/L; G4, G5, G6). Prek waters
471 and floodwaters also have relatively high NH_4 concentrations (6-38 mg/L) compared with river NH_4
472 concentrations that mostly remain < 2 mg/L (Table 1). NH_4 is present in reducing waters and can
473 originate from either natural sources such as the anaerobic decomposition of organic material, or
474 anthropogenic sources including fertilisers and animal or human sewerage. Anthropogenically
475 sourced NH_4 in groundwater can therefore highlight infiltration of water from the surface or shallow
476 sub-surface. Cl/Br molar ratios higher than seawater values (~ 650 to 660; Davis et al., 1998) relative
477 to a large range in Cl concentrations (Cl/Br: 620-1970, Cl: 35-396 mg/L) can indicate contamination
478 of water from human sewerage and septic tank effluent (e.g. Katz et al., 2011). Many groundwater

479 samples in this study have Br concentrations below detection limits, and those values measured
480 show that Cl/Br molar ratios that remain relatively low (340-746; Table 1), including samples G5 and
481 G6 (Cl/Br ratios 464 and 592 respectively) that have relatively high NH_4 . This indicates little
482 contamination from anthropogenic wastewater in the aquifer. In comparison, the irrigation and
483 drainage pre-k waters have elevated Cl/Br molar ratios ranging from 628-947 (Table 1), which may
484 indicate some influence of NH_4 from fertilisers or animal/human sewerage. Therefore, a high
485 component of NH_4 in the groundwater is likely sourced from the anaerobic respiration of organic
486 material, which is discussed further below in relation to arsenic concentrations.

487

488

489 **4.4 Tritium content**

490 The river ^3H results show seasonal differences with values higher in the dry season (mean = 2.5 TU;
491 range = $2.3-2.9 \pm 0.3-0.6$ TU) compared with the wet season (mean = 1.9 TU; range = $1.8-2.5 \text{ TU} \pm$
492 $0.3-0.5$). Higher values can indicate increased contributions from rain in the northern region of the
493 Mekong River Basin during the dry season (Nguyen et al., 2007), which is consistent with the stable
494 isotope value results. Rainfall ^3H data shows that although the atmospheric contents were 10 times
495 lower than those experienced in European rainfall during the nuclear bomb pulse effect, SE Asia also
496 experience elevated atmospheric ^3H contents during the 1960s that have since declined to values
497 consistent with pre-bomb pulse conditions.

498

499 The ^3H contents of groundwater sampled in this study show that most samples were predominantly
500 recharged prior to the 1960s ($^3\text{H} \leq 0.3$ to ≤ 0.8 TU; Table 1). These results indicate that there was not
501 enough recent recharge from rainfall, inundation waters, irrigation waters or river water observed
502 over the seasonal scale for [the apparent tritium content of the bulk sample to be above the](#)
503 [detection limit.](#) This therefore highlights a relatively shallow aquifer system (depth $\sim 30-40$ m) that is
504 semi-confined. In terms of river-groundwater interactions, because there is no detectable ^3H in most

505 groundwater samples this implies that either (i) the sites are outside of the zone of influence of river
506 inflows to the aquifer, which is kept small by the seasonal reversals in hydraulic gradients, or (ii) the
507 river does not sufficiently cut through the aquitard to interact with the semi-confined aquifer.

508

509 At sites G1, G7, and G8 there is some ^3H detected, however the activities are considered too low to
510 significantly different from the other groundwater samples. At site G1, the low ^3H in the wet season
511 (0.8 ± 0.3 TU) is not significantly greater than values in the dry season (≤ 0.7 TU), even though
512 changes in major ion chemistry highlighted a wet season dilution due to recharge by rainfall or
513 inundation waters. At the site G7 there is some ^3H present during both the dry season (1.2 ± 0.3 TU)
514 and the wet season (0.8 ± 0.4 TU), however this is also not significantly different to groundwater
515 samples where the ^3H value is below detection, and similar at site G8 where there is a low activity of
516 ^3H during the dry season (1.2 ± 0.4 TU; wet season analysis was not possible due to the site being
517 inundated). The stable isotope and major ion data suggest that the water recharging the aquifer at
518 sites G7 and G8 is evaporated. This in addition to the lack of significant seasonal change in ^3H at G7
519 indicates that the recharge rates do not significantly differ between dry and wet seasons.

520

521

522 **4.5 Arsenic in groundwater**

523 All Bassac River arsenic (As) concentrations are relatively low, with wet season values ($0.7\text{-}1.3$ $\mu\text{g/L}$)
524 slightly lower than dry season values ($1.3\text{-}1.5$ $\mu\text{g/L}$). In comparison, groundwater shows a larger
525 range in concentrations, with five groundwater sites that have As values below the 10 $\mu\text{g/L}$ limit that
526 WHO (2017) provides as a provisional guide for drinking water (0.2 to 8.6 $\mu\text{g/L}$; Table 2), and one
527 groundwater site that has As values below the 50 $\mu\text{g/L}$ limit that the Cambodian government uses (\leq
528 14.6 $\mu\text{g/L}$). In comparison, the 4 other groundwater sites have high As concentrations above both
529 the WHO and Cambodian government limit (64.1 – 393.8 $\mu\text{g/L}$; Table 2). Where the mean residence
530 time of groundwater is > 60 years the concentrations of As are higher (sites G4 and G5: As = 218.5-

531 393.8 µg/L) compared with sites where the groundwater has younger residence times (< 60 years at
532 sites G7 and G8; As = 64.1-106.1 µg/L).

533

534 When taking into account the total errors of As (Table 2), there are only two sites that show
535 significant changes in As concentrations between the dry and wet seasons (Fig. 5). Site G6 shows an
536 increase in As by 10.9 µg/L during the dry season. Whereas site G1 shows an increase in As by 6.8
537 µg/L during the wet season. All other sites show no significant seasonal change in As concentrations
538 (Fig. 5).

539

540 In comparison with other trace elements, the seasonal increases in As concentrations predominantly
541 corresponds with a seasonal decrease in manganese (Mn, $R^2 = 0.5$) compared with a positive
542 correlation with iron (Fe, $R^2 = 0.5$; Fig. 6). For most groundwater, with both high and low As
543 concentrations, the HCO_3 concentrations remain relatively high and the SO_4 concentrations are
544 generally low (≤ 3 mg/L; Table 1). These results are indicative that reductive-dissolution of Fe
545 oxides/hydroxides is a driving process of high As concentrations in groundwater.

546

547

548 **4.6 Rare earth elements**

549 The results of the rare earth element (REE) concentrations are presented relative to the North
550 America Shale Composite (NASC) in figure 7. For the Bassac River, the high REE concentrations
551 during both the wet and dry seasons (Fig. 7a) indicate some interactions with the sediment load.
552 There are small decreases in Cerium (Ce, where $\text{Ce}/\text{Ce}^* = 0.6-0.8$; Table 2), which is common
553 because of the propensity for this REE to be preferentially removed from solution in oxygenated
554 river water (e.g. Leybourne and Johannesson, 2008). However, the sediment interactions do not
555 significantly change overall REE patterns, which remain flat relative to NASC. The minor seasonal

556 variations and relatively high REE concentrations compared with groundwater samples suggest that
557 groundwater inflows to the river are not detected by the REE concentrations.

558

559 The only groundwater site that resembles river REEs is at G1 during the wet season (Fig. 7a). Other
560 hydrochemical parameters, as presented above, show that this site is significantly recharged by wet
561 season rainfall and not by the river. Therefore, the similar REEs to river values at G1 are likely a
562 reflection that there are limited water-rock interactions occurring between the recharging water and
563 the lithology in the wet season. During the dry season, site G1 has a similar REE pattern to many
564 other groundwater sites where there are small increases in heavy REEs (HREEs) relative to light REEs
565 (LREEs; Fig. 7a). For example, groundwater has greater $(Yb/Nd)_{NASC}$ ratios (0.7-9.8) compared with
566 the stable river patterns (0.9-1.5; Table 2). Increases in HREEs can result from mineral dissolution
567 reactions, the preferential sorption of light LREEs (e.g. Biddau et al., 2002), and/or due to the
568 complexation of more stable HREEs in groundwater of circumneutral pH (e.g. Johannesson et al.,
569 2000). Most REEs patterns for groundwater also exhibit a downward concave pattern for the medium
570 REEs (MREEs; Fig. 7), which can be indicative of a high organic matter content of the REE sources
571 (e.g. Pouret et al., 2007; Davranche et al., 2015).

572

573 In addition to the seasonal variations observed at site G1, the groundwater sites G7, G9, G4 and G6
574 (Figs. 7b-d) also show changes in REEs between wet and dry seasons. Sites G7 and G9 both show a
575 wet season decline in the Ce/Ce^* ratios by 0.5, which is accompanied by a decrease in total REEs. At
576 sites G4 and G6, the dry season LREEs concentrations are lower (Fig. 7d), resulting in declines in the
577 $(Pr/Sm)_{NASC}$ ratios of 0.3 and 0.5 respectively (Table 2). In comparison, the groundwater sites G3, G5,
578 G2, and G10 (Figs. 7e-g) show relatively little temporal variation in both REE patterns and
579 concentrations. For example, both Ce/Ce^* and $(Pr/Sm)_{NASC}$ ratios seasonally vary by ≤ 0.2 (Table 2).

580

581 (Y/Ho)_{NASC} ratios are also included in Table 2 because this can act as an indicator of REE sorption due
582 to Yttrium (Y) having a lower affinity for surface-complexes compared with holmium (Ho). Co-
583 precipitation or sorption of REEs to iron oxyhydroxides (e.g. Möller et al., 1998) or particulate matter
584 (e.g. Nozaki et al., 1997) results in the fractionation of Y and Ho. In this study, the (Y/Ho)_{NASC} values
585 (Fig. 7h) are compared with As concentrations. The results show a negative linear correlation
586 ($R^2=0.7$) between the decreasing (Y/Ho)_{NASC} ratios and increasing As concentrations in both wet and
587 dry season groundwater samples until the groundwater has As concentrations > 100 µg/L. Where
588 groundwater has As concentrations > 100 µg/L the (Y/Ho)_{NASC} ratio increases. The enrichment of
589 (Y/Ho)_{NASC} in groundwater with low As concentrations indicates that the removal of REEs is likely
590 linked to As processes such as sorption or co-precipitation with iron oxyhydroxides.

591

592

593 **5. Discussion**

594

595 **5.1 Hydro(geo)logical processes**

596

597 ***River-groundwater interactions***

598 The hydrochemical results (³H and EC) from this study did not show much evidence of seasonal-scale
599 recharge of the Bassac River to the shallow aquifer, despite in some areas the measured river depths
600 suggest that this would be possible. Previous studies in other sites of the Cambodian Mekong Delta
601 have highlighted a connection between the Bassac and Mekong River and surrounding shallow
602 aquifers that result in linked changes in groundwater quality (e.g. Polizzotto et al., 2008; Lawson et
603 al., 2016; Richards et al., 2017a, b). Therefore, we hypothesise that the bores sampled in this study
604 were located outside of the zone of influence of recharge from the Bassac River (at distances of 0.3-
605 1 km from the river's edge). Despite the high dry season groundwater pumping rates, the river stable
606 isotope and EC values do indicate potentially dry season inflows to the river from the shallow

607 aquifer. In addition, the stable isotope results suggest an increase in local delta rainfall contributions
608 to the river during the wet season compared with greater surface water inflows from northern
609 catchments to the river in the dry season. This corresponds with previous work that has used both
610 physical models (MRC, 2009) and stable and radiogenic isotope data (Nguyen et al., 2007).

611

612 ***Groundwater recharge***

613 One groundwater site (G1) showed evidence of local recharge during the wet season by rainfall or
614 inundation waters. This site therefore is located where there is little or no confining unit at the
615 surface. The combination of lower EC, higher REE concentrations and lower stable isotopes in the
616 wet season compared with the dry season highlights an older laterally flowing groundwater system
617 that mixes with wet season recharging rainfall and/or inundation waters. The mixing calculation
618 using $\delta^{18}\text{O}$ values highlighted a high percentage (~50 %) of rain/inundation waters that mixed with
619 the groundwater system during the wet season.

620

621 A second groundwater site (G7) also showed evidence of recent recharge but, as highlighted by the
622 fractionation of stable isotope values, this is by evaporated waters. At G7, recharge occurs during
623 both the dry and wet season most likely via the nearby permanent surface waters trapped by a dyke
624 in the low elevation areas. Previous studies to the north of the study area have also found ponds and
625 wetlands to be a recharge source for shallow groundwater (e.g. Lawson et al., 2016; Richards et al.,
626 2019). Given the stagnation of these waters at the surface, the underlying sediments must act as a
627 low-infiltration layer. The relatively constant stable isotope and major ion chemistry at G7 suggest
628 that there is constant mixing between the dry and wet season recharge with lateral groundwater
629 flow.

630

631 For 7 of the 10 groundwater sites there is no detectable ^3H contents, therefore the mean residence
632 times are > 60 years old, and these sites represent shallow groundwater from a semi-confined

633 aquifer at ~30 to 62 m depth. The stable isotope and major ion hydrochemistry also indicates that
634 groundwater from this semi-confined aquifer is not locally or recently recharged by rainfall, river,
635 inundation or irrigation waters and changes in EC are controlled by variations in water-rock
636 interactions in the aquifer and in-aquifer mixing.

637

638

639 **5.2 Seasonal changes in groundwater arsenic concentrations**

640 Simplified schematics for each of the three types of groundwater recharge processes (limited recent
641 recharge, evaporated water and rainfall recharge) are presented in figure 8. At each of the
642 groundwater sites there is dry season groundwater pumping that exceeds wet season pumping. The
643 influence of groundwater pumping versus recharge processes impacting on the As concentrations in
644 groundwater is not consistent between the groundwater sites (Fig. 8) and is explored below.

645

646 ***Model 1. Limited recent recharge***

647 Model 1 is assigned to groundwater sites where the aquifer is semi-confined, there is limited
648 modern recharge ($^3\text{H} \leq 0.8$ TU), and there is no recharge from evaporated waters or recent rain
649 water. This includes 5 groundwater sites where the seasonal changes in As concentrations are not
650 significant after accounting for total errors (G2, G3, G5, G10 and G4), and 1 groundwater site where
651 there is a dry season increases in the As concentration (by 10.9 $\mu\text{g/L}$; G6). Further work, including
652 the comparison with groundwater pumping volumes, is required to determine whether the
653 difference in As concentration changes between the sites is related to groundwater pumping
654 practices.

655

656 At G6, the increase in As concentrations correlate with the period of high pumping rates during the
657 dry season. Unlike some As models proposed from sites in Bangladesh, the groundwater pumping at
658 G6 did not induce mixing with younger waters (e.g. Harvey et al., 2002), nor did it induce the dilution

659 of As concentrations (e.g. Radloff et al., 2017). The groundwater that is drawn into the bore during
660 pumping (during both the dry and wet season) has a mean residence time of > 60 years, and changes
661 in As concentrations are therefore related to seasonal-scale in-aquifer processes that are triggered
662 or accentuated during the relatively high pumping volumes of the dry season.

663

664 Groundwater bores installed for water supplies often have long screen intervals to increase the
665 pumping volumes, which is the case for the private bores sampled in this study with screens along
666 the length of the casing. On-going pumping can therefore result in the connection of otherwise
667 separate aquifer zones (Ayotte et al., 2011). At the site G6, changes in As concentrations may result
668 from greater extractions of groundwater during the dry season that draws in groundwater from
669 different aquifer zones that are otherwise not accessed during the wet season when pumping rates
670 are lower. Groundwater elevations have been found to decrease by up to 5 m in the Kandal Province
671 during the June/July irrigation period (JICA, 2002). Since groundwater flows to a bore during
672 pumping are firstly accessed from relatively high hydraulic conductivity (K) zones in the aquifer, only
673 after prolonged pumping will regions of lower K contribute to the groundwater volumes extracted.
674 Therefore, particularly towards the end of the dry season, the continued groundwater pumping can
675 cause inflows of waters with distinct chemical properties. In previous work on the shallow
676 groundwater aquifer in the Mekong Delta, at a site 8 km south of our study area, Wang et al. (2018)
677 reported that the sediments with the highest hydraulic conductivities ($K = 3.8 \times 10^{-6}$ to 4.6×10^{-6} cm/s)
678 were associated with relatively low concentrations of As (11-12 ppm) and total organic carbon (TOC:
679 0.9-1 wt %), compared with the lower K (2.6×10^{-7} to 5.9×10^{-7} cm/s) sediments that have higher As
680 concentrations (15-28 ppm) and TOC contents (4.8-5.1 wt %). Therefore, pumping-induced inflows
681 from these lower K zones can contribute greater concentrations of As to the groundwater pumped.
682 In comparison, during the wet season, pumped volumes are lower, therefore mostly higher K zones
683 in the aquifer are drawn into the aquifer, thereby reducing the As concentrations.

684

685 In addition to mixing processes directly affecting As concentrations during pumping, the resultant
686 groundwater chemistry changes can also potentially induce changes in ion sorption/desorption and
687 mineral dissolution/precipitation processes that also affect As concentrations. For example, in near-
688 neutral waters minor increases in pH can result in desorption processes that increase As
689 concentrations (Richards et al., 2019). However, in this study the increased As concentrations during
690 the dry season is associated with a minor decrease in pH compared with the wet season. Therefore
691 seasonal changes in groundwater pH are not driving changes in As concentrations.

692

693 The results from this study highlight that the increase in in-aquifer interactions between
694 groundwater from higher K/lower K zones produces high As concentrations in groundwater in the
695 dry season, which may result from changes in redox-driven processes over short time-frames
696 (McMahon et al., 2011). Reducing groundwater conditions promote the reduction of As(V) to As(III)
697 that can lead to desorption from ferric (Fe(III)) and Mn (hydr)oxides and/or sedimentary organic
698 matter (e.g. Richards et al., 2019). In addition, the release of high OC groundwater can result in the
699 respiration of DOC producing NH_4 and an increase in the reductive dissolution of Fe (hydr)oxides also
700 releasing As into solution (e.g. Lawson et al., 2016). At G6, the changes in groundwater chemistry
701 corresponds with these processes; with dry season increases in Fe concentrations (from 1.23 to
702 21.57 $\mu\text{g/L}$), and high concentrations of HCO_3 and NH_4 (Table 1).

703

704 Changes in REE patterns and ratios can indicate changes in the groundwater system that are
705 significant for As. For example, temporal changes in REEs in groundwater can result from changes in
706 the aquifer pH and redox conditions and mixing between waters (e.g. Dia et al., 2000; Seto and
707 Akagi, 2008). For model 1, the REE data confirms that sites G3, G5, G2, and G10 (Figs. 7e-g) show
708 relatively little temporal variation in mixing or redox effects on REE patterns and concentrations,
709 similar to low changes in As concentrations. In comparison there are changes in LREEs
710 concentrations and patterns at sites G4 and G6 in the dry season (Fig. 7d), which are likely to result

711 from mixing with waters that have interacted with different mineralogy and/or organic matter
712 compared with wet season groundwater. This is consistent with the As being derived from a low K
713 zone of the aquifer during pumping. However, this change in groundwater mixing during the dry
714 season only results in a significant increase in As concentrations at site G6 and not at site G4 (Fig. 8).

715

716 ***Model 2. Recharge by evaporated surface waters***

717 The second As model is based on data at site G7, where the groundwater has been recently
718 recharged and there is mixing with evaporated waters. At site G7 there is no significant change in As
719 between the dry season ($67.6 \pm 3.6 \mu\text{g/L}$) and wet season ($64.1 \pm 3.5 \mu\text{g/L}$). Similar to Model 1, this
720 site is located in an area where there are significant increases in groundwater pumping during the
721 dry compared with the wet season. In contrast with model 1, the groundwater from the shallow
722 aquifer is recharged during both the dry and wet seasons by evaporated surface waters. Although
723 recharged by evaporated surface waters, there is no evidence that the As concentrations are high
724 due to affects from evaporation. For example, As molar ratios relative to conservative ions in
725 solution (e.g. Cl) are high at G7 (dry season As/Cl: 0.008) compared to samples with low As
726 concentrations (dry season As/Cl: < 0.00005).

727

728 As concentrations are likely to be controlled by evaporated water recharge processes, and less so by
729 pumping since there are no changes in As concentrations at the seasonal-scale. This is consistent
730 with models proposed in other sites of the Cambodian Mekong Delta where there is limited
731 groundwater pumping, which showed that for groundwater outside of the influence of river inflows,
732 the high As concentrations are predominantly the result of recharge of evaporated surface waters
733 and the transfer of young organic matter into the aquifer resulting in the reductive dissolution of
734 Fe/Mn (hydr)oxides (Lawson et al., 2013, 2016; Richards et al., 2018). The low vertical transfer times
735 (ie. > 200 years; Lawson et al., 2013) of the shallow clay-rich sediments in the Mekong Delta implies

736 that the confining layer underlying the evaporated surface waters at site G7 is relatively thin, and
737 allows for transfer times to the shallow groundwater system of < 60 years (³H data, this study).

738

739 Due to the local recharge processes occurring at site G7, there is less likely to be the horizontal
740 stratification of groundwater chemistry that was proposed in model 1. Increases in dry season
741 groundwater pumping may induce greater drawdown rates of the recharging waters, and may result
742 in the observed increase in the Ce/Ce* ratio (Table 2), which in a reducing aquifer can be indicative
743 of an increase in REEs sourced from an organic-rich environment (Pouret et al., 2010; Davranche et
744 al., 2015). This implies that during the dry season the evaporative water recharge transfers more
745 organic-rich waters compared with the wet season. However, this did not result in a significant
746 increase in As concentrations in the groundwater.

747

748 ***Model 3. Rainfall recharge***

749 Model 3 is based on groundwater at the site G1. This is the only site in this study to show evidence
750 of significant recharge by wet season rainfall, which results in wet season increases in REE
751 concentrations, decreased $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values, and the dilution of many major ions. Despite this,
752 there were some ion concentrations that increased during recharge, including As (from 1.3 ± 0.1 to
753 8.6 ± 0.5 $\mu\text{g/L}$), and Fe (4 to 25 $\mu\text{g/L}$). Therefore, a driver of increases in As at site G1 is linked to wet
754 season recharge processes, and likely due to the transfers of young OC and the resultant reductive
755 dissolution of Fe(III) (hydr)oxides (as reflected in wet season increases in Fe). Similar to the model 2,
756 due to efficient recharge processes there is unlikely to be any significant horizontal stratification of
757 groundwater chemistry, and the seasonal-scale recharge indicates that at this site there is a clay-
758 window and therefore the shallow aquifer is unconfined.

759

760

761 **5.3 Future work on As models**

762 As observed in the 3 models proposed, there are multiple physical and biogeochemical processes
763 impacting groundwater in the Kandal region of the Cambodian Mekong Delta that result in the high
764 spatial heterogeneity of the groundwater As concentrations. In contrast, we observe very few sites
765 with significant seasonal changes in groundwater As concentrations. Temporal variations were only
766 observed at two sites, and the causes of these variations ranged from natural processes (significant
767 seasonal recharge at site G1), to potential anthropogenic impacts (from groundwater pumping at
768 site G6). Further work in the region requires an analysis of the seasonal changes in As concentrations
769 at sites such as G6 in combination with groundwater extraction volumes, analysis of sediments, and
770 changes in hydraulic head data. In addition, monitoring bores should be installed so that an analysis
771 of aquifer processes can be undertaken outside of the bias of pumping influences and the regional
772 extent of irrigation pumping effects on groundwater As concentrations can be assessed. This would
773 allow a more in-depth analysis of the natural versus anthropogenic processes controlling the spatial
774 and seasonal changes in As in groundwater from the irrigated area of the Cambodian Mekong Delta.
775 Finally, the on-going projects devoted to rehabilitating current water resources infrastructure (e.g.
776 deepening of the prek channels) raise questions on the potential impact on groundwater quality.
777 The deepening could result in faster vertical transfer times or clay-windows allowing surface water
778 recharge into the semi-confined aquifer, and thus may result in an increase in As mobilisation in
779 groundwater with the introduction of young organic matter.

780

781

782 **6. Conclusions**

783 Managing groundwater As contamination issues is important nationally for the rural community of
784 Cambodia in the delta, but also for the 17.5 million Vietnamese people that live down-gradient in
785 the Mekong Delta. This study presents a first insight into the seasonal variations of As
786 concentrations in groundwater in the irrigation region of the Cambodian Mekong Delta. Where the
787 aquifer is semi-confined, this provided ideal sites to consider groundwater pumping impacts on As

788 concentrations because there were no wet season recharge processes influencing changes in
789 groundwater chemistry. The results from this study show that in the shallow semi-confined aquifer a
790 lack of seasonal variations in As concentrations at most groundwater sites indicates that the wet-dry
791 variations in pumping volumes are not significantly impacting the mobilisation of As in the semi-
792 confined aquifer. There was only one site (G6) that exhibited dry season increases in As
793 concentrations, which is potentially caused by the dry season increases in groundwater pumped for
794 irrigation. In areas where the aquifer is not confined, it was the natural recharge processes (of either
795 evaporated surface waters or wet season rainfall) that controlled As concentrations in groundwater.
796 The results from this study present a conceptual model of the seasonal-scale processes affecting As
797 concentrations in the irrigated area of the Cambodian Mekong Delta, and early findings are similar
798 to other southern and southeast Asia regions such as Bangladesh where models of both natural and
799 anthropogenic processes are required to describe the As processes. In future work, information on
800 pumping volumes is essential to verify (i) whether groundwater extraction is causing the increase in
801 the As concentrations at site G6, and also (ii) whether the lack of seasonal changes in groundwater
802 As concentrations at the other sites of the semi-confined aquifer in the irrigated area is solely
803 controlled by long-term natural processes.

804

805

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811

812

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Table 1

[Click here to download Table: Table 1.docx](#)

Table 1. Field parameters, major ions, stable and radiogenic isotope results for the Bassac River, inundation and irrigation water, and groundwater in the dry and wet seasons (2017 and 2018).

Sample ID	Date	Depth (m)	pH	EC ($\mu\text{S}/\text{cm}$)	HCO ₃ ⁻ (mg/L)	Cl	F	Br	PO ₄	NO ₂	NO ₃	NH ₄	SO ₄	Na	K	Mg	Ca	SiO ₂	$\delta^{18}\text{O}$ ‰ VSMOW	$\delta^2\text{H}$ ‰ TU	³ H ‰ TU	Evaporation ^a %	Cl/Br molar ratio	
Bassac River																								
R1	13-Jun-18	7.71	209	70.8	12.2	0.06	<0.01	<0.01	<2	<2	<2	<2	14.1	9.8	2.3	5.5	19.7	5.0	-7.9	-55.3	2.6	0.3		
	26-Sep-18	7.84	112	53.7	3.8	0.05	<0.01	<0.01	<2	<2	<2	<2	2.4	3.8	1.9	2.7	12.2	5.1	-7.3	-50.4	1.8	0.3		
R2	13-Jun-18	7.75	207	78.1	12.4	0.05	<0.01	<0.01	<2	<2	<2	<2	14.2	9.9	2.4	5.7	19.9	4.9	-7.8	-55.2	n.a.			
R3	13-Jun-18	7.73	207	76.9	12.9	0.05	<0.01	<0.01	<2	<2	<2	<2	14.0	10.1	2.5	5.6	19.7	5.0	-7.9	-55.1	2.9	0.4		
	26-Sep-18	7.82	112	53.7	3.5	0.06	<0.01	<0.01	<2	<2	<2	<2	2.4	3.8	1.8	2.6	12.2	5.0	-7.3	-50.4	2.5	0.3		
R4	13-Jun-18	7.68	206	78.1	12.3	0.03	<0.01	<0.01	<2	<2	<2	<2	13.9	10.1	2.4	5.6	19.2	5.0	-7.8	-54.9	n.a.			
R5	13-Jun-18	7.73	206	78.1	12.7	0.02	<0.01	<0.01	<2	<2	<2	<2	13.9	10.5	2.8	5.6	19.6	4.8	-7.8	-54.9	2.4	0.6		
	26-Sep-18	7.79	109	52.5	3.6	0.02	<0.01	<0.01	<2	<2	<2	<2	2.7	4.0	1.5	2.7	12.3	5.0	-7.3	-50.5	2.2	0.5		
R6	13-Jun-18	7.68	206	75.6	12.2	0.01	<0.01	<0.01	<2	<2	<2	<2	14.0	9.9	2.3	5.6	19.6	4.9	-7.8	-54.7	3.0	0.3		
R7	13-Jun-18	7.68	204	74.4	12.1	0.02	<0.01	<0.01	<2	<2	<2	<2	13.5	9.8	2.2	5.5	17.8	5.0	-7.8	-54.4	2.3	0.3		
	26-Sep-18	7.43	110	54.9	4.1	0.02	<0.01	<0.01	<2	<2	<2	<2	2.8	4.0	2.5	2.8	12.7	4.9	-7.3	-50.6	1.8	0.4		
R8	13-Jun-18	7.70	201	74.4	12.1	<0.01	<0.01	<0.01	<2	<2	<2	<2	13.1	9.7	2.3	5.3	21.1	4.9	-7.8	-54.4	n.a.			
R9	13-Jun-18	7.70	199	73.2	11.8	<0.01	<0.01	<0.01	<2	<2	<2	<2	13.0	9.6	2.2	5.6	18.7	4.9	-7.8	-54.2	n.a.			
R10	13-Jun-18	7.70	197	70.8	11.6	0.00	<0.01	<0.01	<2	<2	<2	<2	13.1	9.7	2.1	5.5	18.9	5.5	-7.7	-53.8	2.4	0.5		
	26-Sep-18	7.70	110	57.3	3.9	0.02	<0.01	<0.01	<2	<2	<2	<2	2.5	4.0	1.8	2.8	12.7	4.9	-7.3	-50.9	1.9	0.4		
R2b	19-Jun-17	n.a.	159	53.7	11.5	0.04	0.02	<0.01	<2	<2	9	<0.01	8.9	2.2	4.2	14.5	n.a.	n.a.	-7.7	-55.1	n.a.		1375	
R8b	27-Sep-17	n.a.	113	56.4	3.5	0.03	<0.01	<0.01	<2	<2	4	<0.01	3.4	1.2	2.8	11.9	n.a.	n.a.	-9.5	-68.3	n.a.			
Irrigation/drainage water																								
1	15-Jun-17	n.a.	319	151.3	41.1	0.27	0.10	0.12	<2	5	38	0.1	18.0	5.9	13.0	30.7	n.a.	n.a.	-5.0	-36.0	n.a.		906	
2	19-Jun-17	n.a.	460	104.9	20.1	0.23	0.07	<0.01	<2	<2	15	<0.01	24.2	3.2	9.9	29.6	n.a.	n.a.	-4.3	-36.4	n.a.		628	
4	19-Jun-17	n.a.	301	44.4	34.5	0.08	0.09	<0.01	<2	<2	25	<0.01	31.6	7.5	8.8	13.1	n.a.	n.a.	-0.8	-18.4	n.a.		826	
5	19-Jun-17	n.a.	228	54.9	16.5	0.07	0.06	0.01	17	2	13	0.01	6.4	5.8	7.3	16.5	n.a.	n.a.	-5.6	-42.3	n.a.		662	
6	19-Jun-17	n.a.	286	88.6	19.7	0.08	0.06	<0.01	<2	<2	16	<0.01	24.5	4.2	8.6	25.1	n.a.	n.a.	-4.0	-34.4	n.a.		782	
7	19-Jun-17	n.a.	175	61.0	12.2	0.05	0.03	<0.01	<2	<2	10	<0.01	11.2	2.4	4.8	16.0	n.a.	n.a.	-6.9	-51.1	n.a.		947	
Floodwater																								
SK2	27-Sep-17	n.a.	203	60.4	4.6	0.09	<0.01	<0.01	<2	<2	6	<0.01	2.9	1.6	3.4	12.3	n.a.	n.a.	-8.3	-61.6	n.a.			
SK1	27-Sep-17	n.a.	113	83.2	12.6	0.04	<0.01	<0.01	<2	<2	16	<0.01	17.7	2.2	5.6	19.4	n.a.	n.a.	-8.8	-64.7	n.a.			
Groundwater																								
G1	14-Jun-18	38	7.1	592	290.4	18.9	0.64	0.13	<0.01	<2	<2	<2	27.2	50.3	2.2	24.9	39.6	20.3	-7.9	-55.1	≤0.7		28	340
	26-Sep-18		7.2	86	40.3	4.6	0.05	<0.01	0.86	<2	<2	<2	2.5	2.7	7.0	2.6	5.7	8.0	-9.1	-58.8	0.8	0.3	16-21	
G2	14-Jun-18	34	7.0	481	283.0	10.8	0.73	0.39	<0.01	<2	<2	<2	0.3	69.1	2.0	14.1	18.9	18.5	-8.2	-57.1	≤0.6		24	62
	26-Sep-18		7.4	546	319.6	11.7	0.40	<0.01	<0.01	<2	<2	<2	2.0	60.6	2.4	19.8	30.1	20.5	-8.0	-56.1	≤0.6		26	
G3	14-Jun-18	62	6.9	4110	402.6	1121.3	0.93	3.39	<0.01	<2	<2	<2	5.2	672.5	3.1	69.6	81.0	29.4	-8.2	-57.5	≤0.6		24	746
	26-Sep-18		7.1	4220	408.7	1234.5	0.61	4.00	<0.01	<2	<2	<2	1.2	671.0	4.3	61.0	85.0	28.9	-8.2	-57.7	≤0.5		23-24	695
G4	14-Jun-18	35	7.7	463	268.4	8.1	0.05	<0.01	0.88	<2	<2	8	<0.01	31.9	4.4	13.1	35.2	15.6	-8.0	-55.7	≤0.8		26-27	
	26-Sep-18		7.9	460	264.7	6.4	0.10	<0.01	1.21	<2	<2	9	<0.01	31.7	5.7	12.8	33.3	16.3	-8.0	-55.7	≤0.5		26-27	
G5	14-Jun-18	38	7.5	698	359.9	25.3	0.03	<0.01	<0.01	<2	<2	20	<0.01	31.2	4.6	23.6	46.9	9.0	-8.0	-56.7	≤0.5		25-26	
	26-Sep-18		7.7	748	375.8	34.7	0.06	0.17	<0.01	<2	<2	23	<0.01	36.1	6.3	24.2	45.8	8.1	-8.0	-56.5	≤0.5		25-27	464
G6	14-Jun-18	40	7.1	815	268.4	113.1	0.52	<0.01	<0.01	<2	<2	8	0.3	103.3	3.2	18.8	24.6	7.8	-8.6	-59.6	≤0.3		19-21	
	26-Sep-18		7.3	589	297.7	30.8	0.41	0.12	<0.01	<2	<2	9	<0.01	56.0	4.2	19.3	26.0	15.1	-8.6	-59.6	≤0.7		19-20	592
G7	14-Jun-18	18	6.9	284	133.0	4.2	<0.01	<0.01	<0.01	<2	<2	<2	<0.01	10.3	0.8	7.0	24.9	18.6	-4.1	-33.4	1.2	0.3	66-71	
	26-Sep-18		7.2	293	144.0	4.3	0.14	<0.01	<0.01	<2	<2	2	<0.01	11.0	1.9	7.2	25.1	20.3	-4.3	-34.4	0.8	0.4	65-69	
G8	14-Jun-18	22	6.8	248	115.9	4.4	<0.01	<0.01	<0.01	<2	<2	<2	3.0	10.7	1.0	6.2	22.2	19.2	-3.5	-28.7	1.2	0.4	73-80	
G9	14-Jun-18	35	7.4	456	263.5	13.0	0.38	<0.01	<0.01	<2	<2	<2	1.5	25.3	1.5	27.5	31.0	19.2	-8.0	-55.8	≤0.5		26-27	
	26-Sep-18		7.5	412	242.8	9.9	0.44	<0.01	<0.01	<2	<2	<2	0.6	18.9	1.9	27.0	25.4	19.8	-8.2	-56.3	≤0.5		25-26	
G10	14-Jun-18	25	7.3	1214	269.6	19.1	0.50	<0.01	<0.01	<2	<2	<2	472.8	96.5	0.8	56.5	88.6	21.0	-7.6	-53.0	≤0.4		31-32	
	26-Sep-18		7.5	1207	295.2	21.5	0.67	0.14	<0.01	<2	<2	<2	457.3	100.1	1.5	55.6	92.4	20.1	-7.6	-53.1	≤0.8		30-32	347

a: Evaporation values represent the modelled % of groundwater sourced from evaporated water (Richards et al., 2018); n.a.: not analysed.

Table 2

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Table 2. Trace element and rare earth element concentrations and ratios for the Bassac River and groundwater in the dry and wet seasons (2018).

Sample ID	Date	As	sample error	^a total error	Li	B	Al	Mn	Fe	Rb	Sr	Ba	Pb	Y	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	(Yb/Nd) _{NASC}	^b Ce/Ce*	(Y/Ho) _{NASC}	(Pr/Sm) _{NASC}	
		µg/L	±	±	µg/L	ng/L																											
Bassac River	R1	13-juin-18	1.4	0.0	0.1	2.2	18.0	11.2	0.3	11.5	3.8	126.5	40.4	0.0	28.6	17.9	29.2	6.2	24.2	4.8	1.2	5.3	0.8	4.3	0.9	2.5	0.4	2.4	0.4	1.1	0.6	0.9	0.9
		26-sept-18	1.3	0.0	0.1	0.6	6.1	7.8	16.2	11.6	1.8	55.5	31.4	0.0	18.8	13.0	23.1	3.6	18.1	3.6	0.9	4.1	0.6	3.3	0.7	1.8	0.3	1.7	0.3	1.0	0.7	1.2	0.7
	R2	13-juin-18	1.4	0.0	0.1	2.3	18.4	9.4	0.3	9.4	3.9	127.9	40.3	0.0	23.6	14.7	23.6	4.3	16.2	3.7	1.1	4.1	0.6	3.6	0.8	2.3	0.4	2.1	0.3	1.4	0.6	0.9	0.8
	R3	13-juin-18	1.4	0.0	0.1	2.2	18.0	6.7	0.3	6.3	4.1	128.8	39.5	0.0	20.6	12.8	18.3	3.5	14.3	3.6	1.1	3.6	0.5	3.0	0.6	1.9	0.3	2.0	0.3	1.5	0.6	1.0	0.7
		26-sept-18	0.9	0.0	0.1	0.7	5.7	5.8	3.5	17.0	1.6	54.2	26.4	0.0	23.8	18.6	34.1	5.3	23.4	5.5	1.3	5.4	0.7	4.3	0.8	2.4	0.3	2.1	0.3	0.9	0.7	0.9	0.7
	R4	13-juin-18	1.4	0.0	0.1	2.3	18.1	12.5	0.3	13.4	3.7	127.5	39.5	0.0	29.1	20.2	33.6	5.4	25.0	4.9	1.1	5.0	0.8	5.0	0.9	2.6	0.4	2.6	0.4	1.1	0.7	1.0	0.8
	R5	13-juin-18	1.5	0.0	0.1	2.3	18.3	14.4	1.8	16.5	3.9	126.9	39.7	0.1	66.0	26.2	46.5	7.5	35.9	8.9	2.1	11.2	1.6	10.2	1.8	5.3	0.7	4.3	0.6	1.3	0.7	1.1	0.6
		26-sept-18	0.8	0.0	0.0	0.6	5.7	6.3	1.5	12.4	1.8	54.4	31.6	0.0	28.8	18.6	27.2	4.9	21.6	5.1	1.3	5.5	0.7	4.3	0.9	2.5	0.3	2.1	0.4	1.0	0.6	1.0	0.7
	R6	13-juin-18	1.4	0.0	0.1	2.2	17.9	9.5	0.3	9.5	3.6	127.4	38.2	0.0	23.3	16.0	26.0	4.0	17.4	4.0	0.9	4.1	0.6	3.7	0.7	2.1	0.3	2.2	0.3	1.4	0.7	1.0	0.7
	R7	13-juin-18	1.4	0.0	0.1	2.3	17.6	10.7	0.3	11.1	3.6	126.3	38.6	0.0	25.9	17.2	28.3	4.4	20.2	4.4	1.0	4.9	0.7	4.2	0.8	2.3	0.4	2.1	0.3	1.1	0.7	0.9	0.7
		26-sept-18	0.8	0.0	0.0	0.5	5.5	5.0	1.2	11.5	1.7	53.7	32.5	0.0	25.1	16.9	25.1	4.4	20.2	4.7	1.1	4.7	0.7	3.8	0.7	2.3	0.3	1.9	0.3	1.0	0.6	1.0	0.7
	R8	13-juin-18	1.3	0.0	0.1	2.2	17.6	11.7	0.3	12.2	3.5	125.4	37.6	0.0	25.7	19.1	30.6	4.7	21.2	5.0	1.2	4.9	0.7	4.0	0.8	2.6	0.3	2.2	0.3	1.1	0.7	1.0	0.7
	R9	13-juin-18	1.3	0.0	0.1	2.2	17.1	11.0	0.3	11.3	3.4	123.3	36.3	0.0	25.5	18.4	30.1	4.8	21.1	5.0	1.2	5.1	0.6	4.4	0.8	2.5	0.3	2.4	0.3	1.2	0.7	0.9	0.7
	R10	13-juin-18	1.3	0.0	0.1	2.2	17.7	30.5	0.5	31.4	3.6	127.6	38.4	0.0	44.4	38.8	69.9	9.9	44.8	10.3	2.3	9.1	1.3	7.9	1.5	4.3	0.6	3.8	0.6	0.9	0.8	0.9	0.7
		26-sept-18	0.7	0.0	0.0	0.5	5.7	6.7	0.8	13.7	1.7	53.4	29.1	0.0	27.3	19.2	29.1	5.1	22.8	5.5	1.4	5.6	0.8	4.3	0.8	2.4	0.4	2.1	0.4	1.0	0.6	0.7	0.7
Groundwater	G1	14-juin-18	1.3	0.0	0.1	9.4	34.1	1.5	485.2	3.9	0.3	418.8	180.9	0.1	15.8	3.8	4.4	0.6	2.4	0.8	0.2	0.8	0.1	1.0	0.2	0.9	0.1	0.8	0.1	3.7	0.7	1.9	0.5
		26-sept-18	8.6	0.0	0.5	2.3	13.0	18.4	109.9	25.2	2.9	43.3	31.2	0.2	46.8	28.8	67.1	8.2	35.3	8.8	2.3	9.7	1.3	7.7	1.6	4.8	0.6	4.3	0.6	1.3	1.0	0.9	0.7
	G2	14-juin-18	0.6	0.0	0.0	16.4	41.6	1.4	891.6	1.7	0.6	129.3	72.7	0.2	15.7	2.0	3.4	0.3	1.5	0.6	0.2	0.6	0.1	0.9	0.3	1.2	0.2	1.0	0.2	7.3	0.9	1.6	0.4
		26-sept-18	0.5	0.0	0.0	19.6	31.4	8.7	788.1	1.4	0.6	183.8	77.7	0.1	9.5	1.8	2.2	0.2	2.3	0.5	0.1	0.5	0.1	0.4	0.1	0.5	0.1	0.6	0.1	2.8	0.7	2.5	0.4
	G3	14-juin-18	0.2	0.0	0.0	96.2	169.3	1.7	1696.1	<0.03	0.3	669.4	189.1	0.1	22.2	2.8	3.2	0.4	1.5	0.8	0.3	1.5	0.2	1.5	0.3	1.0	0.2	1.4	0.2	9.8	0.7	2.2	0.3
		26-sept-18	0.2	0.0	0.0	87.1	164.6	3.4	1609.8	3.9	0.6	678.4	172.4	0.0	18.0	3.0	2.6	0.4	2.0	0.6	0.2	0.9	0.1	0.7	0.2	0.7	0.1	0.7	0.1	3.9	0.5	3.2	0.5
	G4	14-juin-18	231.9	2.1	12.5	4.6	105.0	1.8	213.8	8.0	14.4	784.1	1004.6	0.0	2.2	4.8	0.7	0.0	0.3	0.6	0.2	<0.02	<0.05	<0.11	<0.02	<0.10	0.0	<0.08	0.0		0.4		0.0
		26-sept-18	218.5	1.5	11.7	4.5	103.9	18.3	209.4	6.5	14.0	725.2	886.5	0.0	6.4	8.7	4.8	0.7	5.7	1.5	0.3	1.0	0.1	0.7	0.1	0.4	0.1	0.5	0.1	0.8	0.4	2.0	0.3
	G5	14-juin-18	383.7	10.7	23.0	4.6	73.4	1.1	1035.8	5.9	7.3	496.0	1657.0	0.0	2.8	15.8	0.7	0.3	1.7	2.1	0.5	<0.02	<0.05	<0.11	0.0	<0.10	<0.01	<0.08	<0.01		0.1	1.7	0.1
		26-sept-18	393.8	2.0	21.0	4.8	79.8	2.3	1115.9	3.3	6.8	471.7	1756.1	0.0	2.9	3.1	1.9	0.5	3.9	1.5	0.4	1.3	0.1	0.8	0.1	0.2	0.0	0.2	0.1	0.7	0.3	1.0	0.2
	G6	14-juin-18	14.6	0.3	0.8	6.3	54.4	0.7	885.8	21.6	1.8	246.7	138.2	0.0	1.6	1.4	0.8	0.1	0.4	0.3	0.2	<0.02	<0.05	<0.11	<0.02	0.1	<0.01	<0.08	0.1		0.6		0.1
		26-sept-18	2.7	0.1	0.2	6.3	47.3	1.9	938.4	1.2	1.7	239.6	77.5	0.0	6.9	3.3	3.7	0.6	2.7	0.7	0.2	0.8	0.1	0.5	0.1	0.4	0.1	0.3	0.1	1.3	0.6	1.6	0.6
	G7	14-juin-18	67.6	0.3	3.6	3.8	16.2	0.2	450.7	657.4	2.4	145.1	383.1	0.0	0.9	9.7	15.3	2.2	9.7	2.1	0.6	2.4	0.2	1.5	0.3	0.9	0.2	1.0	0.1	1.1	0.7	0.1	0.8
		26-sept-18	64.1	0.8	3.5	3.2	17.1	0.5	392.1	8.6	3.8	136.6	334.8	0.0	1.6	2.5	1.7	0.9	3.5	0.9	0.2	0.8	0.1	0.5	0.1	0.3	0.0	0.3	0.1	1.0	0.2	0.5	0.7
	G8	14-juin-18	106.1	1.1	5.7	2.6	16.0	0.1	608.2	5095.0	3.1	121.9	200.6	0.0	0.8	1.8	<0.42	<0.01	0.2	0.5	0.4	<0.02	<0.05	<0.11	<0.02	<0.10	<0.01	<0.08	<0.01				
	G9	14-juin-18	1.1	0.1	0.1	19.2	25.8	0.7	267.1	15.8	0.2	220.5	29.4	0.0	33.7	3.6	8.0	0.8	4.2	1.0	0.4	1.6	0.2	1.9	0.4	1.5	0.2	1.0	0.2	2.6	1.0	2.3	0.6
		26-sept-18	1.4	0.1	0.1	17.3	24.7	1.2	296.4	3.1	0.6	177.8	24.9	0.0	4.1	1.1	1.1	0.2	0.9	0.3	0.1	0.5	0.1	0.3	0.1	0.2	0.0	0.1	0.0	1.6	0.5	2.1	0.4
	G10	14-juin-18	1.2	0.0	0.1	30.2	35.7	1.1	835.1	9.2	0.5	759.5	30.9	0.3	39.9	3.0	5.0	0.6	2.7	0.7	0.2	1.3	0.2	2.0	0.5	1.5	0.2	1.3	0.2	5.3	0.8	2.5	0.6
		26-sept-18	0.9	0.1	0.1	28.1	34.3	4.5	1050.3	3.2	0.6	686.5	27.9	0.0	47.2	7.7	9.1	1.0	4.3	1.2	0.4	1.9	0.3	3.1	0.7	2.6	0.3	2.1	0.2	5.1	0.7	2.0	0.6

a: $\sqrt{(\text{sample error})^2 + (\text{analytical error})^2}$; b: $\text{Ce/Ce}^* = \text{Ce}_{\text{NASC}} / (\text{La}_{\text{NASC}} / \text{Pr}_{\text{NASC}})^{0.5}$

Table 3[Click here to download Table: Table 3.docx](#)

Table 3. SLRS-6 data from this study, and CRM SLRS-6 data from Yeghicheyan et al. (2019) and certified values.

Element	SLRS-6 (n=6) ^a µg/L	2 σ STD %	CRM SLRS-6 µg/L
Li	0.53 ± 0.04	6.7%	0.53 ± 0.02
B	7.33 ± 0.34	4.6%	7.39 ± 1.28
Al	33.7 ± 0.9	2.7%	33.8 ± 2.2 ^c
Mn	2.16 ± 0.07	3.2%	2.12 ± 0.1 ^c
Fe	82.8 ± 1.6	1.9%	84.3 ± 3.6 ^c
As	0.57 ± 0.03	5.3%	0.57 ± 0.08 ^c
Rb	1.50 ± 0.05	3.3%	1.41 ± 0.05
Sr	41.21 ± 0.68	1.7%	40.66 ± 0.32 ^c
Ba	14.3 ± 0.64	4.5%	14.28 ± 0.48 ^c
Pb	0.169 ± 0.009	5.3%	0.17 ± 0.026 ^c
Y	0.133 ± 0.003	2.3%	0.128 ± 0.006
La	0.249 ± 0.004	1.6%	0.248 ± 0.012
Ce	0.295 ± 0.011	3.7%	0.293 ± 0.015
Pr	0.061 ± 0.012	11.7%	0.059 ± 0.002
Nd	0.238 ± 0.002	4.2%	0.228 ± 0.009
Sm	0.041 ± 0.002	4.9%	0.039 ± 0.002
Eu	0.008 ± 0.001	6.0%	0.007 ± 0.001
Gd	0.033 ± 0.001	3.0%	0.032 ± 0.002
Tb	0.004 ± 0.001	9.3%	0.004 ± 0.001
Dy	0.023 ± 0.001	3.3%	0.022 ± 0.001
Ho	0.004 ± 0.001	7.0%	0.004 ± 0.001
Er	0.013 ± 0.001	7.7%	0.012 ± 0.001
Tm	0.002 ± 0.001	13.0%	0.002 ± 0.001
Yb	0.012 ± 0.001	2.5%	0.011 ± 0.001
Lu	0.002 ± 0.001	7.3%	0.002 ± 0.001

^aValues from this study from 6 long term replicates;^bcompilation values from Yeghicheyan et al. (2019);and ^ccertified values (µg/L).

Figure 1
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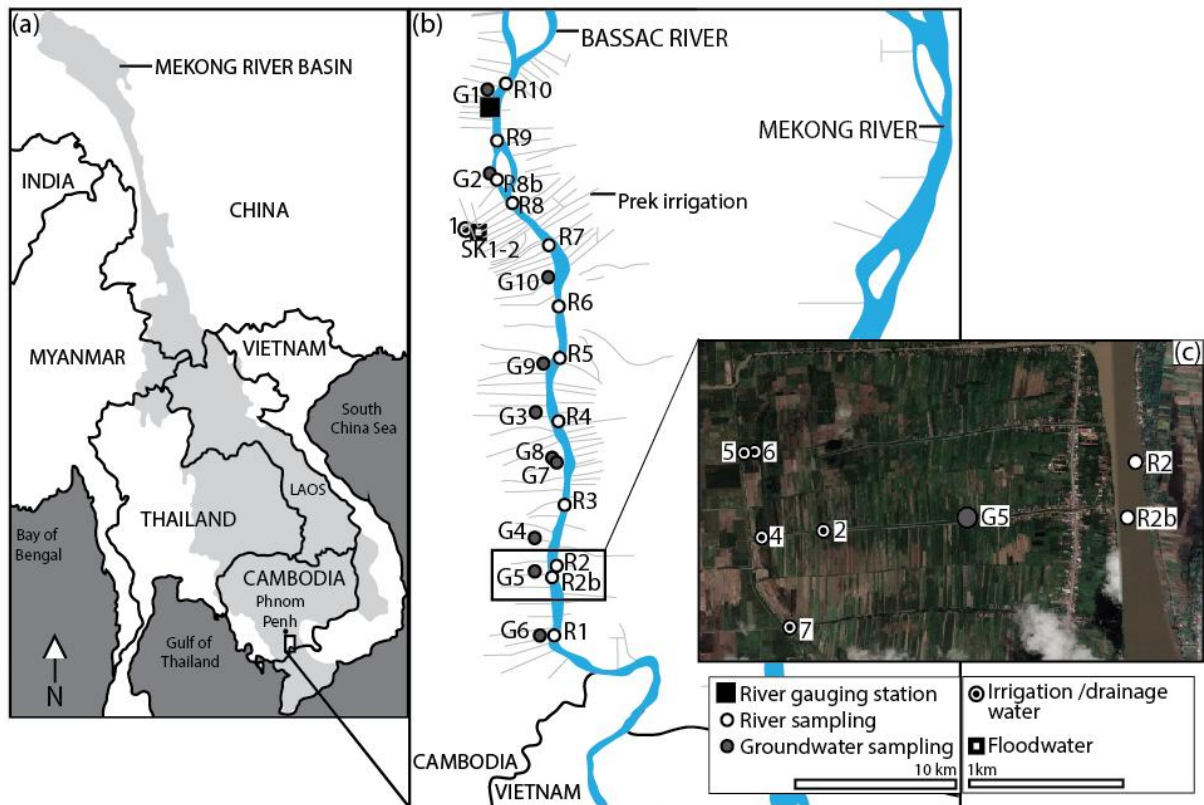


Figure 1. Map of (a) the study area located in the Cambodia Mekong Delta, Kandal province; (b) the sites for river water (R) and groundwater (G) sampling in 2018, and the river gauging station (Koh Khel) in the north of the study area; and (c) close-up view of areas sampled for irrigation/drainage prek water overlaying the image of the irrigated landscape (Google Earth, 24-June-2018).

Figure 2
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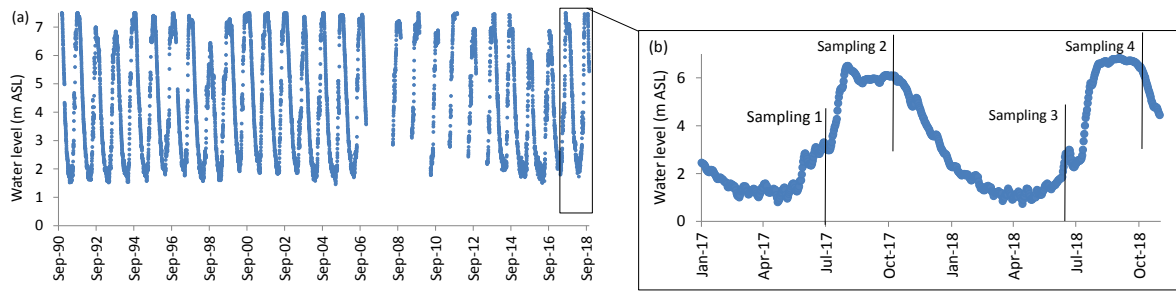


Figure 2. (a) Bassac River water levels (at Koh Khel gauging station, 33402; [Mekong River Commission, 2019](#)) from 1990 to 2018, and (b) a zoom of the river water levels during the sampling rounds in this study in 2017 and 2018.

Figure 3

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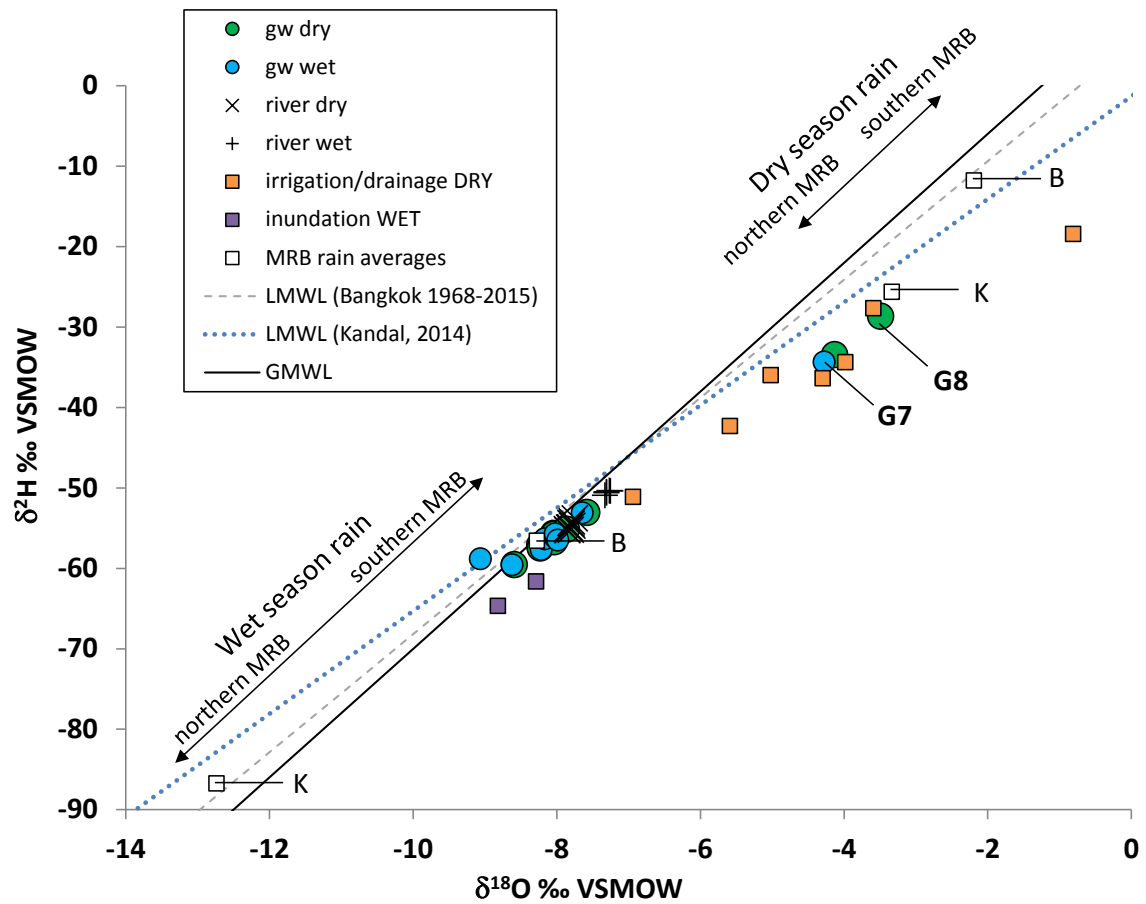


Figure 3. Stable isotope values for rain, groundwater, river, inundation, irrigation/drainage prek water collected during the dry and wet seasons 2017-2018. Values are compared with the global meteoric water line (GMWL), and the local MWLs (LMWL) from the lower delta rainfall at Bangkok (B; IAEA/WMO station 1968-2015 data) and at Kandal (K; 2014 data, Richards et al., 2018). Also presented are average values for dry season months (February and March) and wet season months (September and October) for rainfall near the southern region of the Mekong River Basin (MRB) at Bangkok, Thailand (B), and the rainfall near the northern MRB at Kunming, China (K) (data from IAEA/WMO, 2019).

Figure 4

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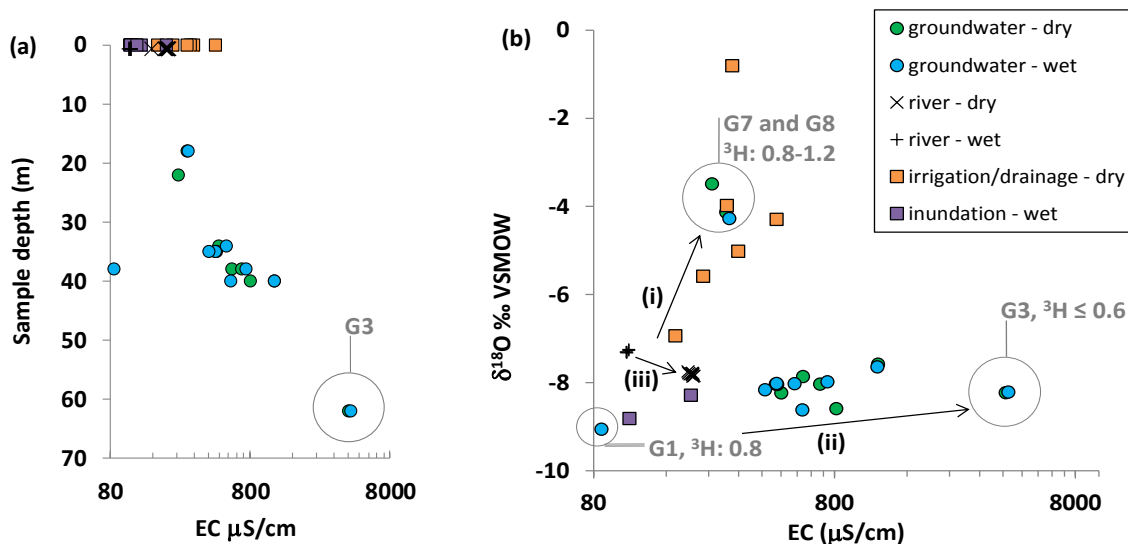


Figure 4. (a) EC values with depth to sample in the river (sampled at 0.5 m depth) and depth of the borehole in the aquifer and (b) EC and $\delta^{18}\text{O}$ values of groundwater, river, inundation, and irrigation/drainage pre-k water collected during the dry and wet seasons. This graph highlights 3 end-members for groundwater sampled at sites G1, G3, and G7/G8 that result from 2 main mechanisms controlling increases in groundwater EC; (i) evaporation or mixing with evaporated waters, and (ii) water-rock interactions, evapotranspiration or mixing with groundwater that had ^3H values below the detection limit, and 1 main mechanism increasing river EC; (iii) increased mixing with groundwater.

Figure 5
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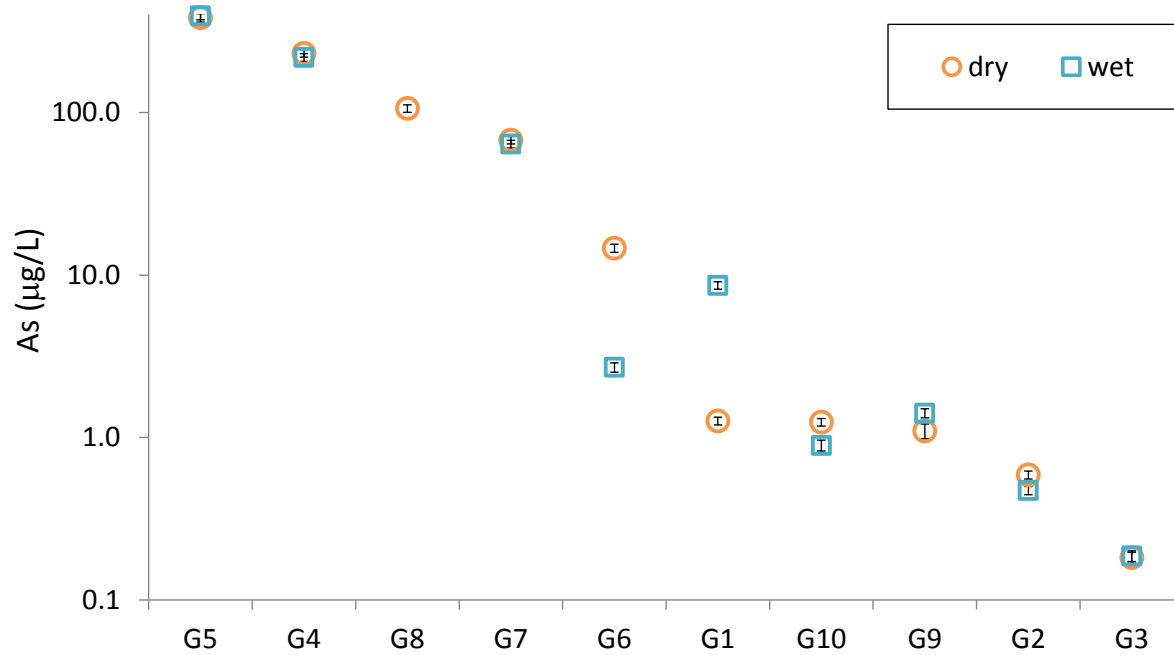


Figure 5. Dry and wet season concentrations for As (and total error bars) at each groundwater site.

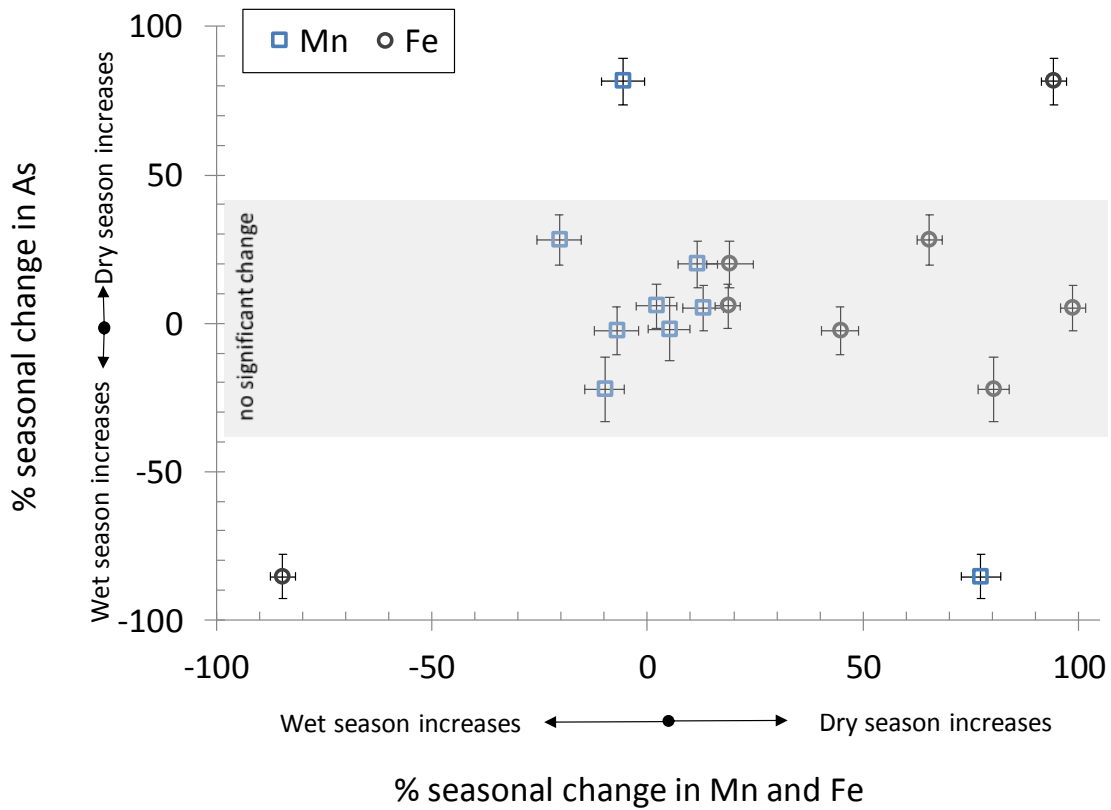


Figure 6. Percentage seasonal changes in As concentrations relative to percentage seasonal changes in Mn and Fe concentrations (with total error bars). The grey bands on both sides of the 0% change show where changes are significant for As concentrations.

Figure 7
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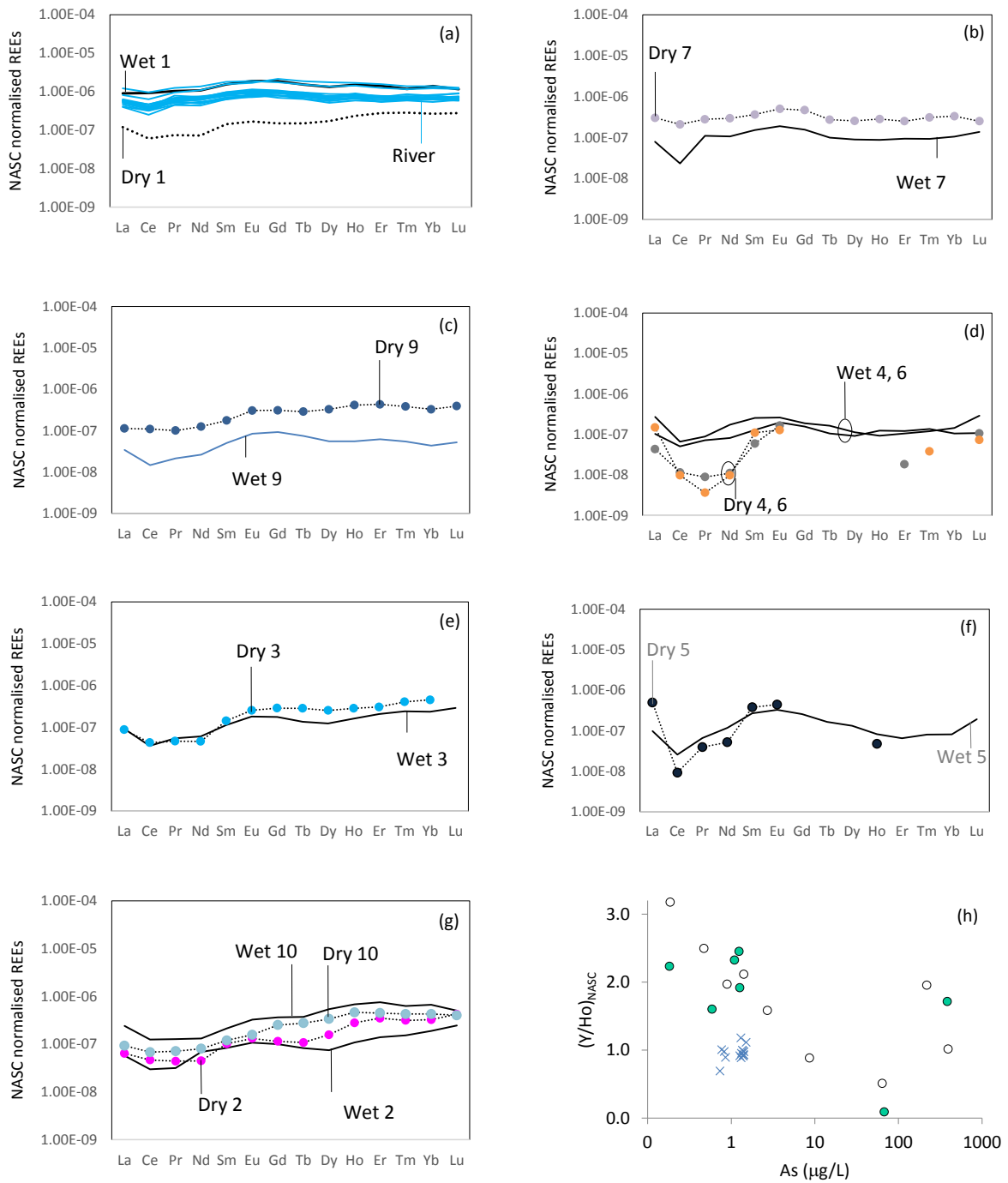


Figure 7. Rare earth element data normalised to NASC for (a) the Bassac River (dry and wet seasons) and groundwater at site G1, (b)-(g) groundwater sites G2-G10 during the dry and wet seasons, and (h) ratio of $(Y/Ho)_{NASC}$ relative to arsenic concentrations in river water (crosses), dry season groundwater (filled circles) and wet season groundwater (open circles).

Figure 8

[Click here to download Figure: Figure 8.docx](#)

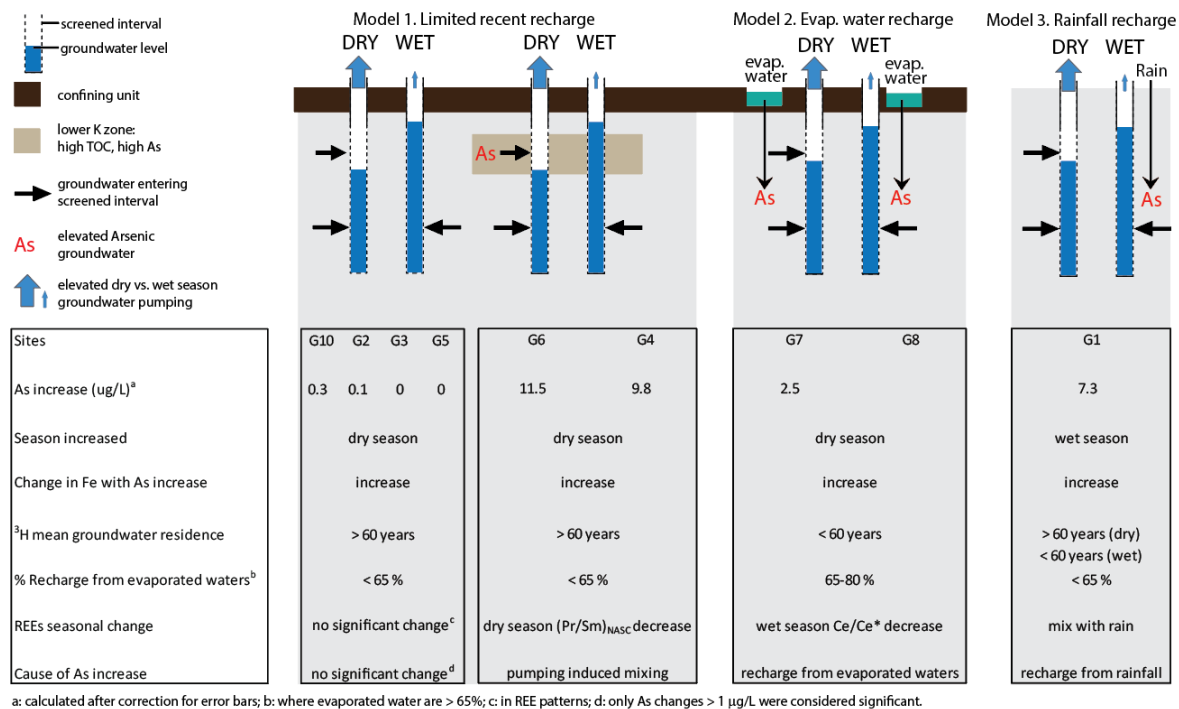


Figure 8. Simplified schematic for the models of As mobilisation proposed in this study. The seasonal changes in the groundwater elevations are unrepresentative of actual magnitudes. The dry season declines in groundwater levels were unable to be measured, and are based on anecdotal evidence and changes in seasonal groundwater elevations that were reported in a previous study by JICA (2002).