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1 **Global patterns and dynamics of climate-groundwater interactions**

2

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26

27

28 **Summary Paragraph**

29 Groundwater is the largest available store of global freshwater¹, upon which more than two
30 billion people rely². It is therefore important to quantify the spatiotemporal interactions
31 between groundwater and climate. However, current understanding of the global scale
32 sensitivity of groundwater systems to climate change^{3,4} – as well as the resulting variation in
33 feedbacks from groundwater to the climate system^{5,6} - is limited. Here, using groundwater
34 model results in combination with hydrologic datasets, we examine the dynamic timescales
35 of groundwater system responses to climate change. We show that nearly half of global
36 groundwater fluxes could equilibrate with recharge variations due to climate change on
37 human (~100 year) timescales, and that areas where water tables are most sensitive to
38 changes in recharge are also those that have the longest groundwater response times. In
39 particular, groundwater fluxes in arid regions are shown to be less responsive to climate
40 variability than in humid regions. Adaptation strategies must therefore account for the
41 hydraulic memory of groundwater systems which can buffer climate change impacts on water
42 resources in many regions, but may also lead to a long, but initially hidden, legacy of
43 anthropogenic and climatic impacts on river flows and groundwater dependent ecosystems.

44

45

46 **Text**

47 Groundwater flow systems exist in dynamic balance with the climate, connecting interacting
48 zones of recharge (i.e. the replenishment of water in the subsurface) and discharge (the loss
49 of groundwater from the subsurface), with multiple feedbacks. As climate varies, changes in
50 the quantity and location of natural groundwater recharge lead to changes in groundwater
51 storage, water table elevations and groundwater discharge¹. These changes in time and space
52 play a central role in controlling the exchange of moisture and energy across the Earth's land
53 surface^{5,6} and connect processes critical to, for example, hydro-ecology, as well as carbon and
54 nutrient cycling⁷. Climate-groundwater interactions may also have played a key role in the
55 evolution of our own and other species⁸ and continue to be critical in setting the availability
56 of water for abstraction by humans in coupled food-water-energy systems¹. Recent global
57 mapping of water table depths⁹ and the critical zone¹⁰ suggest where interactions of climate
58 and groundwater may be most tightly coupled. However, they do not resolve where
59 groundwater systems are most sensitive to changes in climate and vice versa, or the
60 timescales over which such changes may occur.

61 Here, we derive and combine global scale analytical groundwater model results and other
62 hydrologic data sets to provide the first global assessment of the sensitivity of groundwater
63 systems to changes in recharge in both space and time (Figure 1), and discuss their utility as
64 an emergent constraint in understanding and modelling groundwater interactions with climate
65 and other Earth systems at the global scale.

66 We have characterized the mode of groundwater-climate interactions as being either
67 principally bi-directional or uni-directional using an improved formulation of the water table
68 ratio (*WTR*)^{11,12} mapped globally at high resolution (Figure 1a, Figure S1-2). The *WTR* is a
69 measure of the relative fullness of the subsurface and thus the extent of the water table's
70 interactions with topography. Values of *WTR*>1 indicate a topographic control on water table

71 conditions broadly correlating to shallow (<10 metres below ground level, m bgl) water table
72 depths (*WTDs*) globally (see Methods and Figure S3). This is indicative of a prevalently bi-
73 directional mode of groundwater-climate interaction (Figure 1c) where the climate system
74 can both give to the groundwater system in the form of recharge, and receive moisture back
75 via evapotranspiration if *WTDs* are shallow enough.

76 The land surface in such regions rejects a proportion of the potential recharge, and
77 groundwater can have a limiting control on land-atmosphere energy exchanges⁵; a tight two-
78 way coupling between groundwater and surface water is also common. In contrast, in
79 ‘recharge controlled’ areas where $WTR < 1$, water tables are more disconnected from the
80 topography and, while groundwater may still receive recharge from the land-surface, the
81 extent of two-way interaction between climate and groundwater is limited and the mode of
82 interaction is predominantly uni-directional (Fig. 1c).

83 We find that regions where $WTR > 1$ cover around 46% of the Earth’s land area (see Methods,
84 Figure 1a,b) and contribute to the large, but until recently underestimated, extent of
85 groundwater-vegetation interactions globally^{10,13,14}. Consistent with previous regional
86 analyses and the form of the governing equation (see Methods), our results indicate that bi-
87 directional interactions are more likely to occur in areas with high humidity, subdued
88 topography and/or low permeability. In contrast, regions with $WTR < 1$ are more common in
89 drier climates or more mountainous topography¹¹.

90 In order to assess the large scale temporal sensitivity of climate-groundwater interactions we
91 have used an analytical groundwater solution to quantify groundwater response times (*GRTs*)
92 globally and at high resolution. *GRT* is a measure of the time it takes a groundwater system to
93 re-equilibrate to a change in hydraulic boundary conditions¹⁵. For example, the *GRT*
94 estimates the time to reach an equilibrium in baseflow to streams (or other boundaries) after a
95 change in recharge rate, potentially from climate or land use change. Our results indicate that

96 groundwater often has a very long hydraulic memory with a global median *GRT* of nearly
97 6000 yrs, or approximately 1200 yrs when hyper-arid regions, where recharge is <5 mm/y,
98 are excluded (Figure 1d,e). Only 25% of Earth's land surface area has response times of less
99 than 100 yrs (herein called 'human timescale'). However, this is equivalent to nearly 44% of
100 global groundwater recharge flux, calculated by aggregating contemporary recharge over the
101 land area where *GRT*<100 y, expressed as a proportion of the total global recharge. Around
102 21% by area have uni-directional climate-groundwater interactions and response times on
103 human timescales, mostly associated with high permeability geology suggesting a strong
104 lithological control (Figure 2a).

105 The remainder (4%) in areas with bi-directional climate-groundwater interactions are mostly
106 located in the humid, lowland, tropical regions with unconsolidated sediments (e.g. Amazon
107 and Congo Basins, Indonesia), low-lying coastal areas (e.g. Florida Everglades, Asian mega-
108 deltas) or in high latitude, low topography humid settings (e.g. northeastern Canada, parts of
109 northern Europe).

110 A powerful advantage of using analytical groundwater equations such as the *WTR* is that they
111 allow us to directly assess the spatial sensitivity of the mode of climate-groundwater
112 interactions. By taking the derivative of *WTR* with respect to recharge (Figure S4) we have a
113 measure of the sensitivity of the relative fullness of the subsurface to changes in recharge (see
114 Methods). Our results indicate that the mode of climate-groundwater interaction is very
115 insensitive to relative changes in recharge (Figure 2b, Figure S5), with only 5% of the Earth's
116 land surface switching mode for a 50% relative change in recharge rate. This represents a
117 large change in natural groundwater recharge in the context of projections for the coming
118 century¹⁶. However, when absolute recharge rate changes are considered, more sensitivity is
119 apparent and a pattern emerges (Figure S6-7) that indicates the strong inverse relationship
120 between the spatial and temporal sensitivity of groundwater systems to changes in recharge

121 that we observe (Figure 3b). At small, local scales our calculations may have relatively large
122 uncertainties, stemming from the uncertainties in global data sets used for the analysis
123 particularly for hydraulic conductivity (see Methods). However, at the larger scales
124 considered here, Monte Carlo Experiments (MCE) indicate that, once the variance in each
125 parameter is combined, the global estimates have relatively small standard deviations
126 (Figures 1-2, S2).

127 The global pattern of *GRT* (Figure 1d) indicates a propensity for longer hydraulic memory in
128 more arid areas. Despite the expected scatter due to geomorphological and lithological
129 heterogeneity, there is a power law relationship between median *GRT* and groundwater
130 recharge (*R*) such that $GRT \propto 1/R^y$ with $y \sim 2$ (Figure 3a). This discovery is not directly
131 expected from the form of the governing equations but is rather an emergent property of
132 groundwater system interactions with the Earth's land surface and climate system. The
133 principal control on the observed power law is the distribution of perennial streams
134 (Figure S8) to which the *GRT* is most sensitive, and which itself is strongly controlled by
135 climate (Figure S9-11). How to characterize, quantitatively, this climatic control on the
136 perennial stream distributions is a pertinent question for further hydro-geomorphological
137 research.

138 We should not therefore expect *GRTs* to be static nor consider them as 'time constants'
139 despite being mathematically equivalent to other diffusion processes. Rather, *GRTs* will
140 evolve in time as both climate and geology vary the geometry and hydraulic properties of
141 groundwater flow systems. This will occur over long but diverse timescales associated with
142 changing river geometries.

143 Despite its importance, most global climate, Earth system, land surface and global hydrology
144 models exclude groundwater or do not allow groundwater to flow between model grid cells¹⁸⁻
145 ²⁰. While our results suggest that the spatial distribution of the mode of climate-groundwater

146 interactions may be rather static over century long timescales, we have shown that nearly a
147 half of the world's groundwater flux is responsive on 100 y timescales. Hence in order to
148 capture the important mass and energy transfers correctly, which may affect regional
149 precipitation and temperature dynamics^{5,6}, lateral flow circulation of groundwater must be
150 incorporated into the next generation of global models rather than assuming within-grid-cell
151 hydrological closure of the water budget as is currently often assumed²¹⁻²³. Our *GRT*
152 calculations provide direct estimates of spin-up times to improve groundwater-enabled global
153 models, without having to use the currently employed methods of extrapolation²². Given the
154 long *GRTs* present over much of the Earth's land-surface, defining initial conditions with an
155 equilibrium water table calculated for present-day climate conveniently, but wrongly,
156 assumes stationarity in groundwater levels and fluxes. Since groundwater is known to be the
157 part of the hydrological system that takes longest to achieve equilibrium²⁴, new approaches
158 that incorporate the existence of long term transience should continue to be developed²⁵.

159 The global distribution of *GRTs* suggests that widespread, long-term transience in
160 groundwater systems persists in the present day due to climate variability since at least the
161 late Pleistocene in many semi-arid to arid regions (Figure 3a). This is consistent with
162 observations of larger than expected groundwater gradients, given the current low recharge,
163 that have been observed in present day arid zones²⁵. While groundwater residence time and
164 groundwater response time are fundamentally different concepts, we also note the
165 correspondence between high *GRT* and significant volumes of fossil-aged groundwater
166 storage in arid regions^{2,26}. The outcome of this result is that groundwater discharge to oases,
167 rivers or wetlands in otherwise dry landscapes will be particularly intransient in comparison
168 to climate change, in as much as climate controls the variations in groundwater recharge.
169 However, our results also indicate that groundwater response times tend to be greater in
170 regions where water tables are most sensitive to changes in recharge (Figure 3b). This

171 follows from the fact that both the groundwater response time and the derivative of the water
172 table ratio share a strong dependence on the square of the distance between perennial streams
173 (L , compare Equations 10 and 14).

174 Away from these more arid contexts, the responsiveness of groundwater systems has recently
175 been demonstrated to be as important as climate controls for the development of hydrological
176 drought²⁷. For example, low GRT systems tend to enhance the speed of propagation of
177 meteorological drought through to hydrological drought whereas higher GRT systems
178 attenuate climate signals to a greater extent but also show greater lags in recovery from
179 drought. Thus, even within relatively small geographic areas, geological variations can lead
180 to very different drought responses even under similar climate variability. By way of a
181 specific example, increasing lags between meteorological and hydrological drought indicators
182 have been observed between the two most significant aquifers in the UK²⁸ in a manner
183 consistent with what would be expected from our estimates of GRT (i.e. Cretaceous Chalk
184 limestone - GRT s of months to years, Permo-Triassic sandstone – GRT s of years to 100s
185 years, Figure 1d).

186 Our analysis therefore provides a new framework for understanding global water availability
187 changes under climate change. First, the discovery of a power law relating groundwater
188 recharge and GRT suggests that important areas of groundwater discharge in naturally water
189 scarce parts of the world are likely to be more resilient to climate fluctuations than humid
190 areas. However, where groundwater response times are higher, water tables also tend to be
191 most sensitive to changes in recharge in the long term. Hence, accounting appropriately for
192 groundwater-climate interactions within analyses of global water scarcity in the context of
193 climate change is thus of great importance when explicitly considering the contribution of
194 groundwater storage changes²⁹. Second, the long memory of groundwater systems in
195 drylands also means that abrupt (in geological terms) changes in recharge or widely

196 distributed groundwater abstraction will leave longer legacies. There may also be initially
197 ‘hidden’ impacts on the future of environmental flows required to sustain streams and
198 wetlands in these regions. It is critical therefore that climate change adaptation strategies
199 which shift reliance to groundwater¹ in preference to surface water should also take account
200 of lags in groundwater hydrology³⁰ and include appropriately long timescale planning
201 horizons for water resource decision making. Third, robust assessments of the impact of
202 climate change on hydrological drought require estimates of ‘groundwater responsiveness’²⁷.
203 The timescale of such responses can be directly informed by our results and improve the
204 decision making process with regard to adaptation strategies to changing drought frequencies
205 under climate change.

206

207 **Figure Captions**

208 **Figure 1. Global distributions of water table ratios (*WTR*) and groundwater response times (*GRT*) with**
209 **their conceptual interpretation as metrics of climate-groundwater interactions. (a)** Global map of
210 **$\log(WTR)$ with hyper-arid regions of recharge (R) < 5 mm/y shaded grey¹⁷. (b)** Frequency distribution of global
211 **values of $\log(WTR)$. (c)** Conceptual model for *WTR* as a metric for either bi-directional or uni-directional
212 **groundwater-climate interactions - *WTR* is dependent on R , terrain rise (d), distance between perennial streams**
213 **(L) and the saturated thickness of the aquifer (b). (d)** Global map of $\log(GRT)$. (e) Frequency distribution of
214 **global values of *GRT* - median 5727 yrs (standard deviation, $\sigma = 376$ yrs), or 1238 yrs ($\sigma = 92$ yrs) when hyper-**
215 **arid regions are excluded. (f)** Conceptual model of *GRT* as a metric of the temporal sensitivity of groundwater-
216 **climate interactions.**

217

218 **Figure 2. Global distributions of the temporal and spatial sensitivity of the mode of climate-groundwater**
219 **interactions. (a)** Temporal sensitivity: percentage of uni-directional and bi-directional groundwater systems, by
220 **area globally, that will re-equilibrate significantly to changes in recharge on the timescale of <100 y or >100 y.**
221 **(b)** Spatial sensitivity: percentage of the global area that would change mode from bi-directional to uni-
222 **directional climate-groundwater interactions, or vice versa, for a relative change of 50% in recharge, given an**
223 **unlimited amount of time. Mapped values use the baseline parameter set (see Methods). The median percentage**

224 coverage of Earth's landmass for each category from the Monte Carlo Experiments is labelled in the key with
225 standard deviations in percentage coverage shown in brackets. Grey areas represent contemporary recharge
226 <5 mm/y (ref¹⁷).

227

228 **Figure 3. Global quantitative inter-relationships between climate and the temporal (*GRT*) and spatial**
229 **(*WTR*) sensitivity of groundwater-climate interactions. (a)** Globally, median *GRT* values scale approximately
230 with the inverse of recharge (*R*) squared. Relationships between recharge and aridity index categories are shown
231 on the top axes as derived in Figure S12. Box extents are at 25-75% percentiles, with Tukey whiskers and
232 outliers. Histograms within each box represent median *GRT* values from each MCE realisation. **(b)** The
233 sensitivity of climate-groundwater interactions in time (*GRT*) and space (*dWTR/dR*) are log-correlated. Each
234 point uses median values for a geographic location from the MCE realisations. Inset plots are frequency
235 distributions of the slope and r^2 derived from linear regressions carried out for each realisation indicating
236 consistency in the relationship across the uncertainty range.

237

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320

321 **Author Contributions**

322 The idea for the paper was conceived by MOC and TG. Analyses were by all authors. The
323 manuscript was written by MOC with input from all authors.

324

325 **Data Availability statement**

326 Once the paper is accepted the main data outputs (i.e. *WTR* and *GRT* digital maps) will be
327 made freely available for download via an online data repository and a link included in the
328 final published version of the paper.

329

330

331 **METHODS**

332 **Derivation of Equations**

333 ***Governing groundwater flow equations***

334 The governing equations were formulated by considering an ideal homogeneous, horizontal
335 unconfined aquifer bounded at one end ($x = L/2$) by a stream assumed to be a constant head
336 boundary and at the other ($x = 0$) by a no-flow boundary representing a flow divide
337 (Figure S13). The one-dimensional (Boussinesq) equation of groundwater flow for such an
338 aquifer receiving homogeneous recharge can be given as follows:

339
$$\frac{\partial}{\partial x} \left(Kh \frac{\partial h}{\partial x} \right) = S \frac{\partial h}{\partial t} - R(t) \quad (1)$$

340 where K is hydraulic conductivity [LT^{-1}], S is storativity [-], $h(x,t)$ is hydraulic head [L], t is
341 time [T], x is distance [L] and $R(t)$ is groundwater recharge [LT^{-1}].

342 If changes in transmissivity due to fluctuations in groundwater heads are assumed to be
343 negligible, Equation (1) may be linearised as follows:

344
$$T \frac{\partial^2 h}{\partial x^2} = S \frac{\partial h}{\partial t} - R(t) \quad (2)$$

345 where T is transmissivity [L^2T^{-1}], and $T = KH$, with H the average saturated thickness [L].

346 The lateral boundary conditions are as follows:

347
$$\frac{\partial h(0,t)}{\partial x} = 0, h\left(\frac{L}{2}, t\right) = b \quad (3)$$

348 The parameter L is thus a characteristic length equivalent to the distance between perennial
349 streams which act as fixed head groundwater discharge boundaries.

350 ***Water table ratio (WTR) derivations***

351 For steady state flow, where $h(x,t)$ becomes $h(x)$, the solution to Equation 1 for the stated
 352 boundary conditions is:

$$353 \quad h(x) = \left(b^2 + \frac{R}{K} \left(\frac{L^2}{4} - x^2 \right) \right)^{0.5} \quad (4)$$

354 At the flow divide, $x = 0$, therefore:

$$355 \quad h(0) = \sqrt{b^2 + \frac{RL^2}{4K}} \quad (5)$$

356 For steady state flow, the solution to the linearised form, Equation 2, for the stated boundary
 357 conditions is:

$$358 \quad h(x) = \frac{R}{2T} \left(\frac{L^2}{4} - x^2 \right) + b \quad (6)$$

359 At the flow divide, $x = 0$, therefore:

$$360 \quad h(0) = \frac{RL^2}{8T} + b \quad (7)$$

361 The *WTR* is defined¹² as the ratio of the head at the flow divide above the fixed head
 362 boundary (i.e. $h_0 - b$) to the maximum terrain rise above the fixed head boundary, d [L].

363 This yields a new, non-linearised, form of the *WTR*, from Equation 5 as follows:

$$364 \quad WTR_{NL} = \frac{\sqrt{b^2 + \frac{RL^2}{4K}} - b}{d} \quad (8)$$

365 For the linearised form, from Equation 7, and as originally given by ref¹², the *WTR* is:

$$366 \quad WTR_L = \frac{RL^2}{8Td} = \frac{RL^2}{8KHd} \quad (9)$$

367 Equations 8 and 9 become equivalent for combinations of small L or R , or large K .

368 All maps and analysis presented in this paper use the non-linear form of the *WTR*
 369 (Equation 8) with the exception of Figure S1 where the two versions are compared, and

370 calculated using the L parameters derived using a minimum river discharge threshold of
371 $0.1 \text{ m}^3/\text{s}$. A comparison of global maps and frequency distributions for the linear and non-
372 linear forms are shown in Figure S1-2. The frequency distribution comparison (Figure S2)
373 shows that the new non-linear formulation has a narrower and more symmetric distribution
374 with a median closer to zero than the linearised form. This is indicative of its better physical
375 representation such that the extent of higher $WTRs$ is limited by the feedback between higher
376 water table elevation and concomitant increases in transmissivity inherent in the non-linear
377 Boussinesq equation (Equation 1).

378 The WTR is a measure of the relative fullness of the subsurface and thus the extent of the
379 water table's interactions with topography. We have therefore used the WTR to characterize
380 the dominant mode of groundwater-climate interactions as being either principally bi-
381 directional or uni-directional based on whether they are 'topographically controlled'
382 ($WTR > 1$) or 'recharge controlled' ($WTR < 1$), respectively. This is a reasonable approximation
383 since a global comparison with water table depths ($WTDs$) (Figure S3) indicates that $WTR > 1$
384 broadly correlates to shallow (<10 metres below ground level) water table conditions. This
385 condition is indicative of a prevalently bi-directional mode of groundwater-climate
386 interaction where the climate system can both give to the groundwater system in the form of
387 recharge, and receive moisture back where local variations in $WTDs$ enable
388 evapotranspiration to occur from groundwater directly. In contrast, areas with $WTR < 1$ show
389 increasingly large $WTDs$ well beyond plant rooting depths leading to predominantly uni-
390 directional climate-groundwater interactions where the groundwater system receives recharge
391 from the climate system but there is more limited potential for feedback in the other direction.
392 The sensitivity of the WTR to changing recharge is given by differentiating Equation 8 with
393 respect to R :

394
$$\frac{dWTR_{NL}}{dR} = \frac{L^2}{8Kd} \left(b^2 + \frac{RL^2}{4K} \right)^{-0.5} \quad (10)$$

395 This equation represents the sensitivity of the maximum head to recharge relative to the
 396 topography which can be understood as the sensitivity of the ‘fullness’ of the subsurface to
 397 changes in recharge.

398 Following from Equations 8, we calculate the recharge required for the WTR to equal 1 for
 399 every grid cell as:

400
$$R_{WTR=1} = \frac{4K}{L^2} (d^2 + 2db) \quad (11)$$

401 The difference between R and the values given in Equation 11 then gives an expression for
 402 the change in recharge (ΔR) needed to effect a change in the WTR across the transition
 403 between topography control (bi-directional climate-groundwater interactions) and recharge
 404 control (unidirectional climate-groundwater interactions) modes. In absolute terms this is:

405
$$\Delta R_{abs} = R - R_{WTR=1} \quad (12)$$

406 and in relative terms it becomes:

407
$$\Delta R_{rel} = \frac{R - R_{WTR=1}}{R} \quad (13)$$

408 ***Groundwater response time (GRT) definition***

409 The groundwater response time is, in general terms, a measure of the time it takes a
 410 groundwater system to respond significantly (as defined below) to a change in boundary
 411 conditions^{15,31-35} and is defined here as follows:

412
$$GRT = \frac{L^2 S}{\beta T} \quad (14)$$

413 where β is a dimensionless constant, T is transmissivity [$L^2 T^{-1}$], S is storativity [-] and L is
 414 the distance between perennial streams [L]. To illustrate why this equation defines a time of

415 response consider a groundwater mound such as that shown in Figure S13. Let the initial
416 shape of the mound (of maximum height A), due to some steady recharge, be given by:

$$417 \quad h(x, 0) = A \cdot \cos\left(\frac{\pi x}{L}\right) \quad (15)$$

418 If recharge suddenly ceases (i.e., a step change) then it can be shown, in the manner of ref³³,
419 that the solution to the linearised Equation 2 without recharge (i.e. $R(t)=0$) is:

$$420 \quad h(x, t) = h(x, 0) \cdot \exp\left(-\frac{t}{GRT}\right) \quad (16)$$

421 for β is equal to π^2 .

422 Hence, for this case, the GRT controls the timescale for the groundwater levels to decay
423 exponentially to reach 63% re-equilibrium after a change in boundary (recharge) conditions
424 (i.e., an “e-folding” timescale). This value for β was chosen in order to be consistent with
425 mathematically equivalent uses of ‘time constants’ (often denoted as τ), in other branches of
426 science.

427 As outlined by ref³⁴, comparing the timescale of a particular forcing to the GRT can be a
428 useful measure of the degree of transience a groundwater system will manifest in terms of
429 variations in lateral groundwater flow. However, there is an important difference to note in
430 the case of a step change in conditions, as used to define GRT in Equation 14, in comparison
431 with a periodic variation in the forcing recharge (of period P). For the step change case
432 outlined above, both heads and fluxes decay exponentially after the change in recharge.
433 However, in the periodic case, where $GRT \gg P$, variations in recharge lead to very stable
434 groundwater fluxes (including at the downstream lateral boundary) but large temporal
435 changes in groundwater head across much of the aquifer³⁵. Thus, it is important to distinguish
436 between the control of GRT on the degree of transience in either heads or fluxes, depending
437 on the nature of the boundary conditions.

438 **Spatial input data and manipulation**

439 ***Global mapping of the distance between perennial streams (L)***

440 The distance between perennial streams (*L*) was calculated using a globally consistent river
441 network provided by the HydroSHEDS database³⁶ which was derived from the 90 m digital
442 elevation model of the Shuttle Radar Topography Mission (SRTM). For this study, we
443 extracted the global river network from the HydroSHEDS drainage direction grid at 500 m
444 pixel resolution by defining streams as all pixels that exceed a long-term average natural
445 discharge threshold of 0.1 cubic meters per second, resulting in a total global river length of
446 29.4 million kilometers. Smaller rivers with flows below this threshold were excluded as they
447 are impaired by increasing uncertainties in the underpinning data. However, the sensitivities
448 of the most important results of this paper to the chosen threshold are considered in our
449 uncertainty analysis below. Estimates of long-term (1971-2000) discharge averages have
450 been derived through a geospatial downscaling procedure³⁷ from the 0.5° resolution runoff
451 and discharge layers of the global WaterGAP model (version 2.2, 2014) a well-documented
452 and validated integrated water balance model^{16,38}. Only perennial rivers were included in the
453 assessment; intermittent and ephemeral rivers were identified through statistical discharge
454 analysis (lowest month of long-term climatology is 0) and extensive manual corrections
455 against paper maps, atlases and auxiliary data, including the digital map repository of
456 National Geographic³⁹. *L* was calculated for every pixel of the landscape (Figure S8) by
457 identifying the shortest combined Euclidean (straight-line) distance between two river
458 locations at opposing sides of the pixel. Neighbourhood low pass filters (5x5 kernel size)
459 were applied to remove outlier pixels and speckling. All calculations were performed in
460 ESRI© ArcGIS environment using custom-made scripts.

461 ***Global mapping of the water table ratio (WTR), groundwater response times (GRT) and***
462 ***other expressions***

463 Global *WTR* maps were created from the above equations using: the recharge rate (R in m/y),
464 based on ref¹⁷, a minimum saturated thickness of the aquifer (b) set to 100 m (refs^{40,41}), the
465 distance between two perennial streams (L , in m, as described above), intrinsic permeability
466 values (m^2) reported in ref⁴⁰ were converted to hydraulic conductivity (m/s) by assuming
467 standard temperature and pressure (1×10^7 multiplication factor) and then converted to units
468 of m/y. The maximum terrain rise between rivers (d , in m) was based on the range of
469 elevations in the 250m GMTED2010 data set⁴².

470 The *GRT* was mapped using the same L data and hydraulic conductivity values as for the
471 *WTR* calculations. Transmissivity (T , m^2/y) was calculated by multiplying the hydraulic
472 conductivity with a fixed saturated thickness of 100 m (refs^{40,41}). It was assumed that
473 storativity (S) for unconfined aquifers is dominated by the specific yield and that this can be
474 approximated by mapped porosity values⁴⁵. Owing to the significant uncertainties in these
475 assumptions for calculating T and S values the parameters were subjected to a Monte Carlo
476 analysis as described below.

477 Each of the data sets was prepared to match a global equal-area projection with a grid size of
478 1 km x 1 km, and the calculations of the data sets were performed in ArcGIS. To avoid
479 mathematical problems, for zero values of d and R , 1 and 0.00001 were added, respectively.
480 For *WTR* estimates, regions where contemporary groundwater recharge was estimated as < 5
481 mm/y (ref¹⁷) were excluded from the analysis due to the increasingly large relative
482 uncertainties in recharge below this range, and the resulting unrealistic sensitivity of the
483 resulting *WTR* estimates. For deriving the frequency distributions and comparisons of
484 parameters from the range of derived geo-spatial data sets, point values were taken from each
485 raster of interest for 10,000 randomly distributed locations across the Earth's land-surface.
486 Global distributions of the parameters d , K and S are given in Figure S10 and relationships

487 between R and L ; d and WTR ; and R and WTR are explored in Figure S9, and Figures S11,
488 respectively. All areal calculations ignore the Antarctic landmass.

489 Although we have made best use of coherent available global datasets at high (1 km)
490 resolution for the calculations, our results are intended for appropriate large scale
491 interpretation, not detailed local analysis.

492 **Justification of the model assumptions**

493 Our calculations are based on mapped surface lithology only and, as such, they represent a
494 first estimate of the response of unconfined groundwater across the global land surface. The
495 more complex responses of regional or local confined aquifers, which may be locally
496 important to discerning groundwater-climate interactions, are not considered. However, such
497 confined aquifers only cover around 6-20% of the Earth's surface⁴³, are often located in more
498 arid parts of the world and are, by definition, inherently less connected to the land surface and
499 climate-related processes.

500 Using 1-D analytical solutions to the groundwater flow equations gives a powerful advantage
501 over the use of more complex models in enabling the sensitivity of the key parameters
502 controlling patterns and timescales of climate-groundwater interactions to be analysed
503 analytically. This, for example, allows us to sample the entire parameter space directly rather
504 than a restricted subset via a limited ensemble of more computationally expensive numerical
505 model runs. Equation 1 assumes the validity of the Dupuit-Forchheimer approximation
506 whereby the water table is assumed to be a true free surface governed by effective hydraulic
507 parameters and that water pressure in the direction normal to the flow is approximately
508 hydrostatic. This is a good approximation when the ratio of the lateral extent of the average
509 saturated depth is more than approximately 5 times its depth¹², i.e. $H/(L/2) < 0.2$ (see
510 Figure S13). Calculating the maximum saturated depth h_{max} as the smaller of $d+b$ or h_0 , and

511 approximating the average saturated depth as $(h_{max} + b)/2$, we find that the criterion
512 $H/(L/2) < 0.2$ is met in 96% of our global grid calculations. Locations which fail this test are
513 all in mountainous regions where Equation 1 cannot account accurately for steep hillslope
514 groundwater hydraulics and hence our results may be less reliable in such areas.

515 The *GRT* is a parameter which consistently appears in solutions to the groundwater flow
516 equations and has been used for decades³² as a robust estimate for the timescale of re-
517 equilibration of a groundwater system following a change in boundary conditions^{8,15,30-35,44-48}.
518 Thus it is an appropriate metric for long term transience which is currently impossible to
519 model in state of the art coupled groundwater-surface water models, which are limited to
520 short run times even for regional scale analyses due to their massive computational demands.
521 More realistic aquifer geometries and initial water table configurations lead to behaviours
522 which are more complex than the case of a simple exponential decay⁴⁶, and non-uniform flow
523 fields (strong convergence or divergence) can also lead to variations in *GRT* (refs^{44,47,48}). We
524 have therefore included these factors in an uncertainty analysis as outlined below.

525 While the models used here cannot represent the detailed process interactions in the way that
526 a distributed fully coupled 3-D model would, they have a strong theoretical basis and show
527 consistency with other large scale studies based on very different model assumptions and data
528 sets. Justification for the approach of using *WTR* as a proxy for the mode of climate-
529 groundwater interaction is given in at least four ways. First, at global scale, similarities of
530 *WTR* with shallow *WTD* globally⁹ are strong (Figure S3), given the very different model
531 assumptions and data sets employed in the two studies. Second, at a continental scale for the
532 contiguous US a recent study compared the results of a physically based, 3-D, fully coupled
533 surface water-groundwater model validated against water table depth data, against the *WTR*
534 metric⁴¹. The results show scatter as expected due to variations in the derivation of the
535 comparative characteristic length scales used in the comparison. However, general trends and

536 geographic patterns at a regional scale compare well between the WTD computed by the fully
537 coupled model and the calculated WTRs. Third, also at a continental scale for the contiguous
538 US, a systematic relationship has been shown between *WTR* and mean stream junction angles
539 which are indicative of a strong coupling between surface and subsurface⁴⁹. Lastly,
540 comparisons of *WTR* calculations against a more complex 3-D regional groundwater flow
541 model, has indicated that the *WTR* is a robust indicator of groundwater's connection to the
542 land surface as it is a strong predictor of the propensity for local versus regional flow
543 conditions⁵⁰. Our analyses thus allow us to make a robust first global scale estimate of the
544 sensitivity of climate-groundwater interactions, while enabling the range of uncertainty to be
545 fully and directly appreciated.

546 **Uncertainties and Monte Carlo experiments**

547 We ran 10,000 Monte Carlo experiments (MCE) at 10,000 randomly distributed locations
548 across the Earth's land-surface to investigate the range of uncertainty due to parameter
549 uncertainties as well as model structural simplifications.

550 Hydraulic conductivity (*K*) was allowed to vary log-normally within uncertainty ranges
551 defined in refs^{40,50}, this parameter having by far the highest uncertainty of any others used in
552 our calculations. Groundwater recharge (*R*) values were taken from ref¹⁷ but allowed to vary
553 through a normal distribution with a standard deviation of 22% of this baseline, chosen
554 according to the difference with a contrasting global recharge distribution^{52,53} commonly used
555 in other global hydrological calculations. Storativity (*S*) was sampled from a normal
556 distribution with standard deviations of 25% of the mapped value⁵³. Although the absolute
557 error in the DEM used is only 1-2 m, we allowed the maximum terrain rise (*d*) to vary
558 normally with a standard deviation of 10% to allow for uncertainties due to gridding. The
559 minimum saturated thickness of the aquifer (*b*) was allowed to vary log-normally around

560 100 m with a standard deviation of 0.3 orders of magnitude. Sampled distributions were cut
561 off at zero to stop meaningless negatives being included in the calculations.

562 Parameter uncertainty in the distance between perennial streams (L), calculated from the
563 variation in L for an order of magnitude change in discharge threshold used to define the
564 stream network (from 0.1 to 1 m³/s), gives a median uncertainty of a factor of 1.9. However,
565 there is also additional uncertainty to L due to the choice of the one-dimensional groundwater
566 flow solutions applied, which ignore non-uniform (i.e. convergent or divergent) flow fields
567 which are common in real catchments. In order to account for the maximum likely range of
568 possible uncertainty, we have compared the 1-D analytical solutions used here to cases of
569 radial flow which represent an extreme 2-D non-uniform flow end-member for natural
570 groundwater flow systems. By equating the distance between perennial streams (L) to be
571 equal to the radius of the flow domain for the equivalent radial solutions, we can estimate the
572 impact of this choice on both *WTR* and *GRT*. For *WTR*, by replacing Equation 6 with Eq
573 30.11 from ref⁵⁴, the average error is approximately a factor of 2. For the *GRT*, comparison of
574 recession timescales for 1-D and radial flow cases (e.g. Appendix A of ref⁴⁶) indicates a
575 similar level of uncertainty due to non-uniform flow as for the *WTR*. We therefore added a
576 log-normal variation in L with a standard deviation of 0.3 orders of magnitude to
577 accommodate the likely range of combined parameter and structural uncertainty.

578

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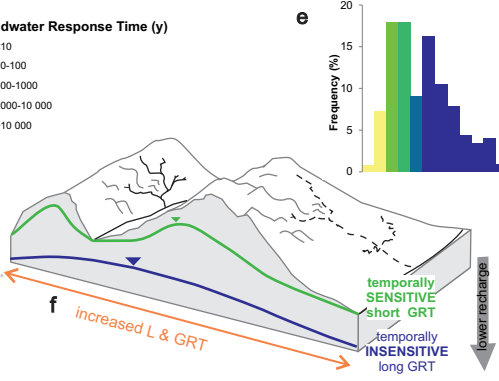
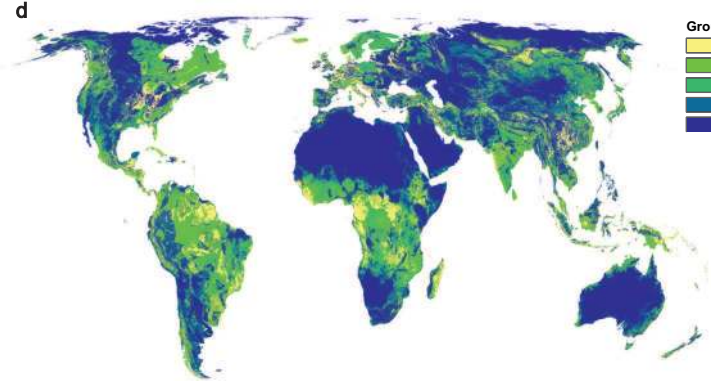
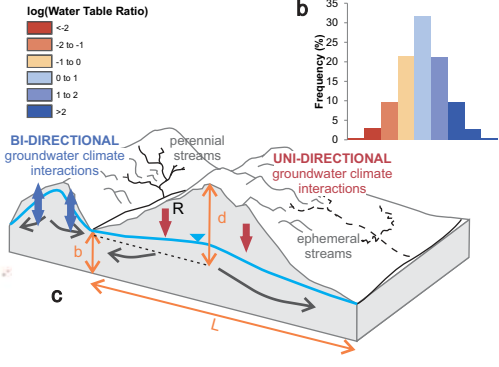
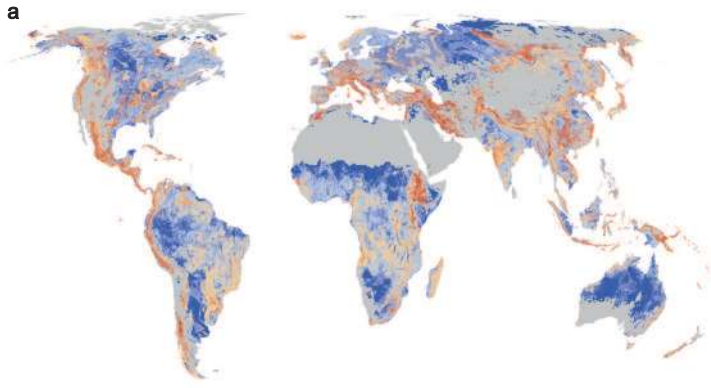
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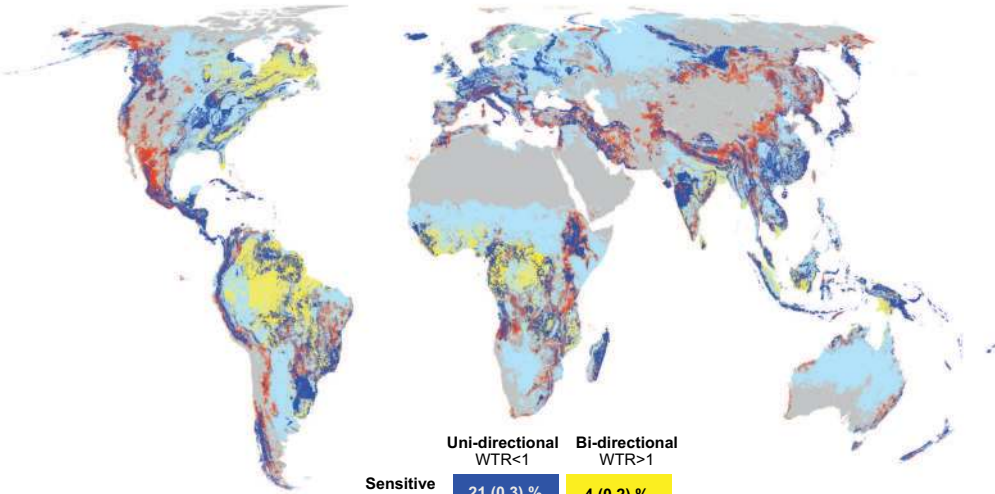
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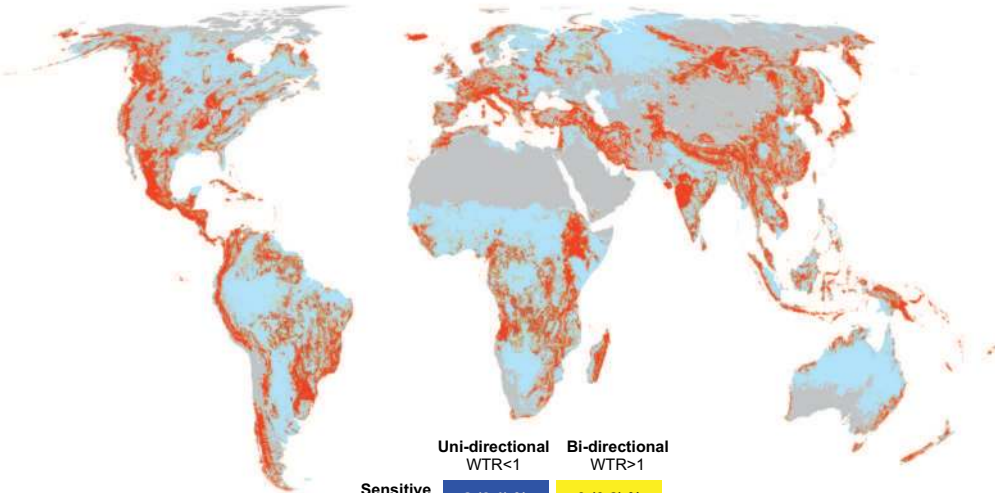


a



	Uni-directional WTR<1	Bi-directional WTR>1	
Sensitive GRT<100 y	21 (0.3) %	4 (0.2) %	R<5 mm/y
Insensitive GRT>100 y	7 (0.2) %	42 (0.4) %	26 (0.1) %

b



	Uni-directional WTR<1	Bi-directional WTR>1	
Sensitive $\Delta R < 50\%$	2 (0.1) %	3 (0.2) %	R<5 mm/y
Insensitive $\Delta R > 50\%$	26 (0.3) %	42 (0.3) %	26 (0.1) %

