



Revisiting hydraulic response times in regional multilayer groundwater systems: the role of aquitards and surface boundaries

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Abstract. Hydraulic response times describe how fast groundwater systems adjust to stresses, such as climatic changes or groundwater abstraction, making them relevant for sustainability, renewability, and vulnerability assessments. These are typically estimated for step changes in boundary conditions using analytical solutions derived from simplified, homogeneous, one-dimensional flow systems, depending on the hydraulic diffusivity and spatial extent of individual hydrogeological units. While unconfined systems are widely recognized to adjust via surface–groundwater interactions, confined systems are often treated as hydraulically isolated from surface boundaries assuming fully impermeable confining units. Here, we revisit these assumptions using a simplified analytical model that represents a multilayer groundwater system as an equivalent homogeneous medium with anisotropic diffusivities in the horizontal and vertical directions, to derive an analytical formulation for its dominant time constant. Different analytical approaches are evaluated against a synthetic numerical model of a regional multilayer groundwater basin subject to step reductions in recharge rates, considering a broad range of hydraulic diffusivities. Results show that one-dimensional formulations can overestimate response times of confined aquifers by several orders of magnitude when the role of aquitards and active surface boundaries is neglected. When inter-layer hydraulic coupling is strong, vertical diffusion through aquitards can significantly accelerate pressure propagation between deeper aquifers and surface boundaries. We introduce a dimensionless criterion comparing horizontal and vertical response timescales to determine whether a multilayer system responds as a hydraulically coupled continuum or as a set of weakly coupled layers. For weakly coupled systems, each layer responds independently and one-dimensional approaches remain appropriate. Conversely, in strongly coupled systems, response times emerge as basin-scale properties generally increasing with depth, and a two-dimensional formulation is needed. These findings delineate the hydrogeological contexts under which different analytical formulations provide reliable order-of-magnitude estimates of hydraulic response times in regional multilayer systems.



1 Introduction

Regional groundwater systems commonly occur within extensive sedimentary basins presenting a layered structure, with permeable units inter-bedded with low-permeability formations. Approximately 35% of the continental land surface is underlain by sedimentary basins that offer favourable conditions for groundwater exploitation, due to the presence of regional aquifers within multilayer systems (Richts et al., 2011). Figure 1 shows the global distribution of major groundwater basins and their mean modern recharge rates (BGR and UNESCO, 2015; Richts et al., 2011). The highlighted basins correspond to some systems that have been subject of numerous hydrogeological studies and that represent the diversity in spatial extent, climatic conditions, and geological settings.

In several of these basins, confined aquifers play an important role in water supply and are becoming increasingly strategic in the context of global water challenges, as pressures on surface water and shallow aquifers intensify. These systems can show long delayed responses to stresses, which entails challenges for their understanding and management (Currell et al., 2016; Kelly et al., 2013; Walton, 2011).

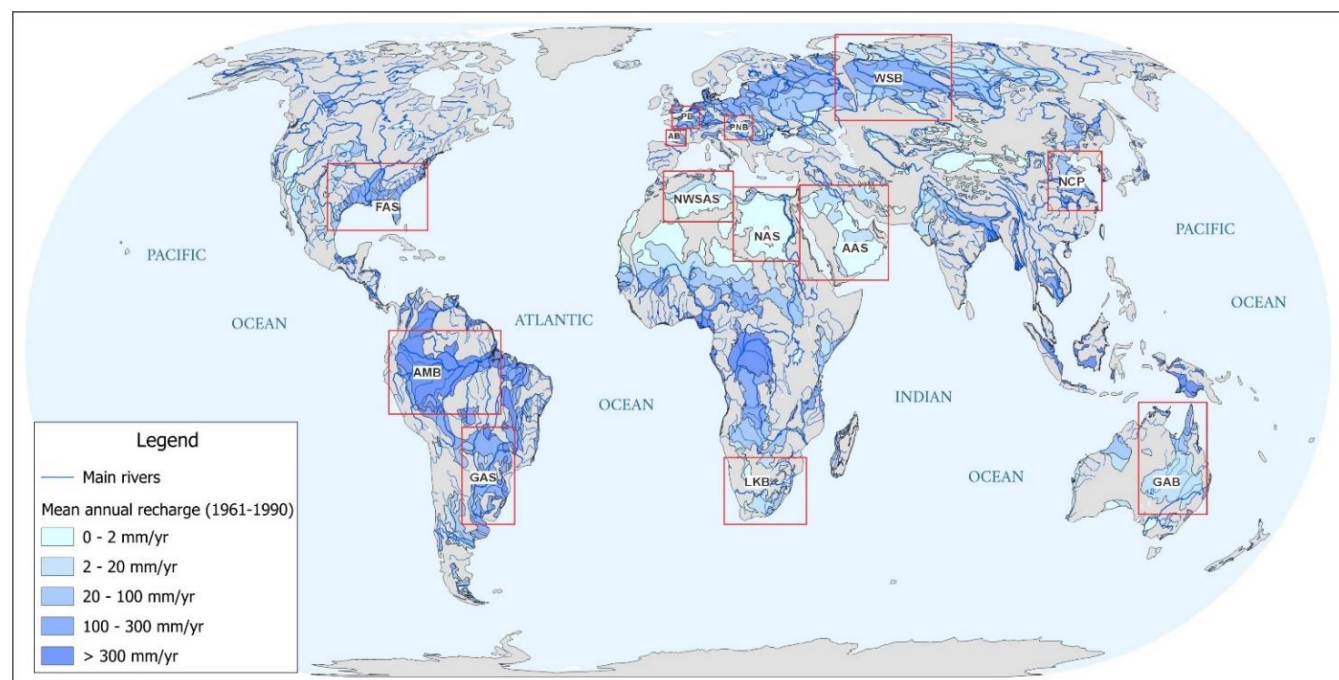


Figure 1. Global distribution of major groundwater basins and mean recharge rates. AAS: Arabian Aquifer System, AB: Aquitaine Basin, AMB: Amazon Basin, FAS: Floridan Aquifer System, GAS: Guarani Aquifer System, GAB: Great Artesian Basin, LKB: Lower Kalahari Basin, NAS: Nubian Aquifer System, NCP: North China Plains, NWSAS: North West Sahara Aquifer System, PB: Paris Basin, PNB: Pannonian Basin, WSB: West Siberian Basin. Adapted from the World Hydrogeological Map (BGR and UNESCO, 2015; Richts et al., 2011). Projection: Equal-area Eckert IV.

These challenges relate to how fast hydraulic perturbations propagate through the groundwater system and how this transient response is relevant in different contexts. One major field of research concerns sustainable and renewable groundwater use,



focusing on how groundwater systems respond to pumping and how much water can be abstracted without causing unacceptable long-term impacts (Bierkens and Wada, 2019; Bredehoeft, 2002; Bredehoeft and Durbin, 2009; Currell et al., 2016; Cuthbert et al., 2023; Erskine and Papaioannou, 1997; Gleeson et al., 2012, 2020; Pierce et al., 2013; Zhou, 2009). A second perspective addresses groundwater vulnerability, emphasizing the sensitivity of systems to natural or anthropogenic stresses, including recharge variations driven by extreme events, long-term climate change, climate variability, and land-use changes (Cuthbert et al., 2019; Houben et al., 2025; Manga, 1999; Schreiner-McGraw and Ajami, 2021).

Another line of research focuses on the influence of paleoclimatic conditions on present-day hydraulic behaviour and the extent to which groundwater systems may remain in disequilibrium with current climate conditions (Ferguson et al., 2026; Lemieux et al., 2008; Rousseau-Gueutin et al., 2013; Schulz et al., 2017; Schwartz et al., 2010; Vasseur et al., 2015). Some important paleoclimatic transitions are known to have occurred globally during the Pleistocene–Holocene transition (~ 12000 yr BP) (Rousseau-Gueutin et al., 2013; Tierney et al., 2017; Vasseur et al., 2015). At the global scale, this transition involved rising temperatures and increased precipitation, marking a shift from the colder, drier conditions near the Last Glacial Maximum to the warmer, more humid conditions of the Holocene (He and Clark, 2022). However, these changes did not occur uniformly across regions. For example, during the early Holocene (~ 11,000 to 5,000 yr BP), the Sahara region in North Africa experienced a humid phase, with precipitation rates estimated to have been up to ten times higher than in present-day arid conditions (Tierney et al., 2017). Although such climatic transitions were gradual, they are often approximated as step changes when considered over geological timescales (Ferguson et al., 2026; Rousseau-Gueutin et al., 2013; Vasseur et al., 2015).

Consequently, the question of how long it takes for a groundwater system to adjust to a stress has been extensively addressed in hydrogeological research through the concept of hydraulic response times. Intuitively, it represents the time required for a hydraulic perturbation to diffuse throughout a groundwater system. Such a perturbation arises from a change in boundary conditions that drives the system from one equilibrium state to another. With this, response times quantify how long a groundwater system takes to re-equilibrate and effectively reach a new steady state (Carr and Simpson, 2018; Rousseau-Gueutin et al., 2013; Vasseur et al., 2015).

Several approaches have been developed to estimate response times for different purposes. These approaches include the use of analytical solutions of the transient groundwater flow equation for simplified one-dimensional flow systems (Cuthbert et al., 2019; Erskine and Papaioannou, 1997; Rousseau-Gueutin et al., 2013; Vasseur et al., 2015; Walker et al., 2015); applying the mean action time theory (MAT), which provides response time estimates without explicitly solving the transient equation (Carr, 2017; Carr and Simpson, 2018; Jazaei et al., 2014, 2016; Simpson, 2018; Simpson et al., 2013); or using spectral analysis approaches over groundwater head or flow time series (Houben et al., 2022; Jiménez-Martínez et al., 2013; Liang and Zhang, 2015, 2013).



There is broad interest in developing simplified, closed-form analytical formulations to estimate quantities in hydrogeology, providing robust first-order estimates that describe the general behaviour of a system and offer insight into its functioning, particularly to inform further characterization, monitoring, or detailed modelling studies (Haitjema, 2006). Analytical solutions for hydraulic response times can provide robust order-of-magnitude estimates, provided the relevant processes at the studied scales are correctly accounted for.

One-dimensional analytical formulations to estimate hydraulic response times are often functions of the flow path length and hydraulic properties of individual hydrogeological units, therefore interpreted as intrinsic properties. This implies that they are independent of external factors, such as the magnitude of the stress or hydrological change (Currell et al., 2016). Some studies suggest that the response time to recharge variations is independent of the recharge rates (Rousseau-Gueutin et al., 2013; Simpson et al., 2013; Vasseur et al., 2015), or that the response time to pumping is independent of the pumping rate when assessed using relative changes in hydraulic heads or flows (Cuthbert et al., 2023; Jazaei et al., 2016). However, Cuthbert et al. (2019) showed that response times in unconfined systems depend on recharge rates as they influence water table depth and the distribution of perennial streams and the flow path lengths of the system. They noted that this dependence is not captured by the analytical solutions to the groundwater flow equation, but instead arises from the nature of the interactions between groundwater, land surface, and climate systems. This hints that the magnitude of hydrological changes or stresses acting on unconfined systems can influence their response times, as they can modify surface-groundwater interactions. In fact, the authors suggest that response times should not be considered static since they can evolve as climatic and geological changes shape the dynamics of the flow systems (Cuthbert et al., 2019, 2023).

Response times have been investigated in both unconfined and confined systems, as well as in mixed systems with confined and unconfined portions. For instance, previous studies of unconfined aquifers have reported response times ranging from several months to decades in humid regions with high surface drainage densities and short flow path lengths (Cuthbert et al., 2019; Houben et al., 2025; Manga, 1999), up to values on the order of $10^4 - 10^6$ years in arid and hyper-arid regions, where flow lengths are much larger due to poorly developed surface drainage networks (Cuthbert et al., 2019; Ferguson et al., 2026). The influence of surface drainage density on response times of unconfined systems has been previously discussed, as it controls the effective characteristic length of the flow system (Cuthbert et al., 2019).

There is an improved understanding of the role of surface boundaries in the response times of unconfined systems, that has led to detailed estimates (Cuthbert et al., 2019; Houben et al., 2025). In contrast, their role in controlling the response times of confined aquifer systems is less well understood. Current one-dimensional analytical formulations assume fully impermeable confining units, implying that equilibration with surface boundaries do not affect the response of confined aquifers (Rousseau-Gueutin et al., 2013; Vasseur et al., 2015). Estimates of response times for major regional confined and mixed aquifers, based



on such analytical approaches, indicate timescales on the order of $10^3 - 10^6$ years, independently of climatic context (Ferguson,
120 2026; Ferguson et al., 2026; Rousseau-Gueutin et al., 2013).

Finally, response times in aquitards have been considered in the context of delayed impacts of pumping in confined aquifers
(Burbey, 2001; Díaz-Nigenda et al., 2023; Helm, 1975; Konikow and Neuzil, 2007; Leake, 1990; Zhou et al., 2013). These
studies highlight the importance of accounting for the transient response of low-permeability units during the development of
125 confined aquifers within multilayer systems. Nevertheless, they have not been integrated in the assessment of response times
to large-scale climatic changes at the regional groundwater basin scale.

In this study, we examine the role of surface boundaries and vertical diffusion through aquitards in controlling hydraulic
response times in regional multilayer groundwater basins subject to large-scale step changes in boundary conditions. In the
130 next section, a conceptual framework of current analytical formulations used to estimate hydraulic response times is provided.
Following, the methodology, results, and discussion of our study are presented.

2 Conceptual Framework

The linear form of the three-dimensional groundwater flow equation in a homogeneous anisotropic porous medium is given
by Eq. (1), where D_j represents the hydraulic diffusivity in the j -th direction [$L^2 T^{-1}$], and R is a source/sink term representing
135 recharge or abstraction rates [$L T^{-1}$]. This form is valid for confined flow and for unconfined flow when water table changes
are small compared to the saturated thickness (Fetter, 2001; Freeze and Cherry, 1979), and is commonly referred to as the
hydraulic diffusion equation. The concept of hydraulic response time thus follows naturally from the description of transient
groundwater flow as a hydraulic diffusion process.

$$140 \quad \frac{\partial h}{\partial t} = D_x \frac{\partial^2 h}{\partial x^2} + D_y \frac{\partial^2 h}{\partial y^2} + D_z \frac{\partial^2 h}{\partial z^2} \pm R \quad (1)$$

In sedimentary basins, two main directions of anisotropy govern groundwater flow: parallel to stratigraphic bedding,
considered approximately horizontal (x and y -directions), and normal to it, considered approximately vertical (z -direction). For
confined aquifers, horizontal and vertical hydraulic diffusivity is defined as the ratio of the corresponding hydraulic
conductivity to specific storage (Reilly and Harbaugh, 2004; Rousseau-Gueutin et al., 2013). For unconfined systems,
145 horizontal diffusivity is given by the ratio of transmissivity to specific yield, while vertical diffusivity is governed by vertical
hydraulic conductivity and specific storage.



Establishing a working definition of response times requires understanding both the physical mechanisms of the hydraulic
150 diffusion process and their mathematical representation. It is essential to distinguish between the characteristic diffusion
timescale (τ_{ch}), the system time constants (τ_n), and the hydraulic response time (t_r). Although closely related and often used
interchangeably, these concepts capture different features of the transient behaviour. These approaches have been developed
from analytical solutions to the diffusion equation considering one-dimensional flow in simplified systems. For a homogeneous
porous medium with one-dimensional flow along the x -direction, and in the absence of sources or sinks, Eq. (1) can be written
155 as Eq. (2), where D is the hydraulic diffusivity. This formulation is valid for confined flow, as well as for unconfined flow
under Dupuit–Forchheimer assumptions with water table changes that are small compared to the saturated thickness.

$$\frac{\partial h}{\partial t} = D \frac{\partial^2 h}{\partial x^2} \quad (2)$$

160 Dimensional analysis of Eq. (2) (Domenico and Schwartz, 1998; Jost, 2024) reveals a natural timescale proportional to the
square of a characteristic length (L) divided by the diffusivity of the medium (D), known as characteristic diffusion timescale
(τ_{ch}), defined in Eq. (3). The characteristic length is a reference length that corresponds to the distance between system
boundaries along the flow direction (Domenico and Schwartz, 1998; Jost, 2024; Rousseau-Gueutin et al., 2013). This
characteristic timescale does not represent the actual time required for a system to re-equilibrate but instead provides a
165 fundamental temporal scale intrinsic to the diffusion process. It can be interpreted as a “measuring device” with the appropriate
time units to track the system as it undergoes a transient response.

$$\tau_{ch} = \frac{L^2}{D} \quad (3)$$

170 For a system with given initial and boundary conditions, the general form of the analytical solution to Eq. (2) can be expressed
as a series expansion of infinite spatial modes, each associated with a distinct exponential decay rate, as shown in Eq. (4) (Carr
and Simpson, 2018; Haberman, 2019; Rousseau-Gueutin et al., 2013; Vasseur et al., 2015). Here, $h(x, t)$ represents the
transient solution, and $h_{\infty}(x)$ corresponds to the final steady state solution. $\phi_n(x)$ are the eigenfunctions representing each
spatial mode, and λ_n are the respective eigenvalues [L^{-2}] driving their decay rates, both being dependent on the boundary
175 conditions. Additionally, C_n are coefficients determined by the initial conditions (Haberman, 2019; Zill and Cullen, 2009). The
transient solution can be expressed in terms of the residual diffusion, $r(x, t)$, which represents the deviation of the system from
the final equilibrium state (Carr and Simpson, 2018; Haberman, 2019; Zill and Cullen, 2009). The time constants for each
spatial mode, τ_n , can then be defined as the inverse of the product of the eigenvalues and the diffusivity, as given in Eq. (5).
They describe the temporal evolution of the residual diffusion, which tends to zero as the system approaches a new equilibrium.

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$$h(x, t) = h_{\infty}(x) + r(x, t) = h_{\infty}(x) + \sum_{n=1}^{\infty} C_n \phi_n(x) e^{-\lambda_n D t}, \quad n = 1, 2, 3, \dots \quad (4)$$



$$\tau_n = \frac{1}{\lambda_n D}, \quad n = 1, 2, 3, \dots \quad (5)$$

When solving the diffusion equation as a boundary-value problem using separation of variables, the time constants are determined by the domain extent, the type of boundary conditions, and the hydraulic properties of the medium (Haberman, 2019; Zill and Cullen, 2009). They are therefore independent of spatial location. However, the decay of each spatial mode is weighted by its amplitude and eigenfunction, both of which vary across position. This introduces a spatial structure in the response, primarily determined by the complexity of the head field and its representation in the series expansion. As a result, each location has an effective time constant depending on which modes contribute most strongly to the solution. This spatial structure in response times arises even in homogeneous cases. However, in real systems, heterogeneities will exert the leading control on the spatial structure of the response.

In a one-dimensional homogeneous flow system, it can be shown analytically that the time constants are proportional to the characteristic timescale, with α_n being a proportionality constant depending on the boundary conditions, as given by Eq. (6). In all cases, the fundamental mode with $n = 1$ represents the largest time constant and dominates the late response of the system (Carr and Simpson, 2018; Rousseau-Gueutin et al., 2013; Vasseur et al., 2015).

$$\tau_n = \alpha_n \frac{L^2}{D}, \quad n = 1, 2, 3, \dots \quad (6)$$

Dominant time constants have been formulated analytically for simplified cases. In the case of an aquifer of length L with one-dimensional horizontal flow, limited by two Dirichlet boundaries and subject to a step change in boundary conditions, the dominant time constant is given by Eq. (7), where D_h is the horizontal hydraulic diffusivity. This formulation is valid for confined flow, as well as for unconfined flow under Dupuit–Forchheimer assumptions.

$$\tau_{h,Aquifer} = \frac{1}{\pi^2} \frac{L^2}{D_h} \quad (7)$$

For an aquitard unit of thickness b undergoing one-dimensional vertical flow and bounded by aquifer layers at the top and bottom, the time constant following a step change in the boundary conditions can be expressed as in Eq. (899), where D_v is the vertical hydraulic diffusivity (Díaz-Nigenda et al., 2023; Zhou et al., 2013).

$$\tau_{v,Aquitard} = \frac{1}{\pi^2} \frac{b^2}{D_v} \quad (8)$$



Similarly, Rousseau-Gueutin et al. (2013) developed an analytical formulation for the dominant time constant of a mixed aquifer of total length L , composed of an unconfined portion of length L_u and a confined portion of length L_c . The formulation
215 assumes one-dimensional horizontal flow, equal transmissivity in both portions, a no-flow boundary at the unconfined end, and a Dirichlet boundary at the confined end. The system relaxes from a linear initial condition with the water table near the surface, to a no-recharge state. The analytical expression for the time constant is given in Eq. (910), where S_u is the storage coefficient of the unconfined portion and T is the transmissivity of the system. The dominant time constant for this case
220 depends only on the storage coefficient of the unconfined area, that typically is around three orders of magnitude higher than in confined systems.

$$\tau_{h,\text{Mixed aquifer}} = \frac{S_u L_u}{T} \left(L_c + \frac{L_u}{2} \right) \quad (9)$$

In theory, given the exponential nature of the decay, the system takes an infinite time to reach its final steady state (Carr and
225 Simpson, 2018; Rousseau-Gueutin et al., 2013), and the time constant represents an “ e -folding” scale (Cuthbert et al., 2019), controlling how long it takes for hydraulic heads to reach roughly 63% of re-equilibration after a step change in boundary conditions (Cuthbert et al., 2019; Jost, 2024; Rousseau-Gueutin et al., 2013). In other words, the time constant represents the time in which the residual diffusion has decayed by about 63%. In practice, the system can be considered as having effectively
230 reached a new equilibrium, or as having “responded”, when the residual diffusion becomes small enough. This illustrates that defining a hydraulic response time necessarily involves the introduction of a threshold to determine when the system can be considered re-equilibrated. A sufficiently small response threshold, $\delta \ll 1$, must be selected to represent the proportion of residual diffusion remaining at $t = t_{res}$ (Carr and Simpson, 2018). Therefore, the response time can then be expressed as a function of the chosen response threshold, as in Eq. (11), where τ is a given analytical formulation of the dominant time constant.

$$t_{r,\delta} = \ln\left(\frac{1}{\delta}\right) \tau \quad (10)$$

It can be shown that after 3τ , the residual diffusion has decayed by about 95% relative to the initial deviation, which means that there is 5% of residual diffusion remaining. This is a general threshold used in the literature, therefore response times are
240 commonly defined as $t_{r,5\%} = 3\tau$, also referred to as time to reach near steady state, and it is the working definition we adopt in this study (Jost, 2024; Rousseau-Gueutin et al., 2013; Schwartz et al., 2010).

In these derivations, the imposed stress is always represented as a step change in boundary conditions, and hydraulic properties are considered homogeneous within the hydrogeological unit. In reality, groundwater systems are subject to multiple stresses
245 of different types and temporal scales, and heterogeneities drive complex responses across spatial scales. The purpose of

simplified analytical formulations for estimating hydraulic response times is to provide a robust, first-order estimator of the overall behaviour of a system, based on representative global hydraulic parameters and geometry, particularly in cases where data and characterisation are sparse. Estimating response times from simplified conceptualizations provides a useful metric for assessing the transient dynamics of a system, and can inform further monitoring and modelling decisions aimed at assessing groundwater vulnerability, renewability, and sustainable use (Currell et al., 2016; Ferguson et al., 2026; Houben et al., 2025).

3 Methodology

3.1 A two-dimensional analytical formulation

Current analytical formulations for estimating response times in confined aquifers assume no-flow boundary conditions at the top and bottom of the unit, implying that their response to a stress is not influenced by the equilibration with surface boundaries and vertical diffusion through aquitards. An analytical formulation is derived to estimate response times in confined aquifers while accounting for these processes. For this, we treat confined aquifers as part of an interacting multilayer system.

Considering a regional groundwater system comprised by k layers, where $D_{h,i}$ and $D_{v,i}$ correspond to the horizontal and vertical diffusivity of the i -th layer, and b_i to its thickness. The k -th layer corresponds to the deepest confined aquifer unit. The system has length L and total thickness B , with x -coordinate parallel to the stratigraphic bedding, and z -coordinate perpendicular to it, as shown in Figure 2. $D_{h,eq}$ and $D_{v,eq}$ are the equivalent horizontal and vertical hydraulic diffusivities, respectively, obtained by layer averaging.

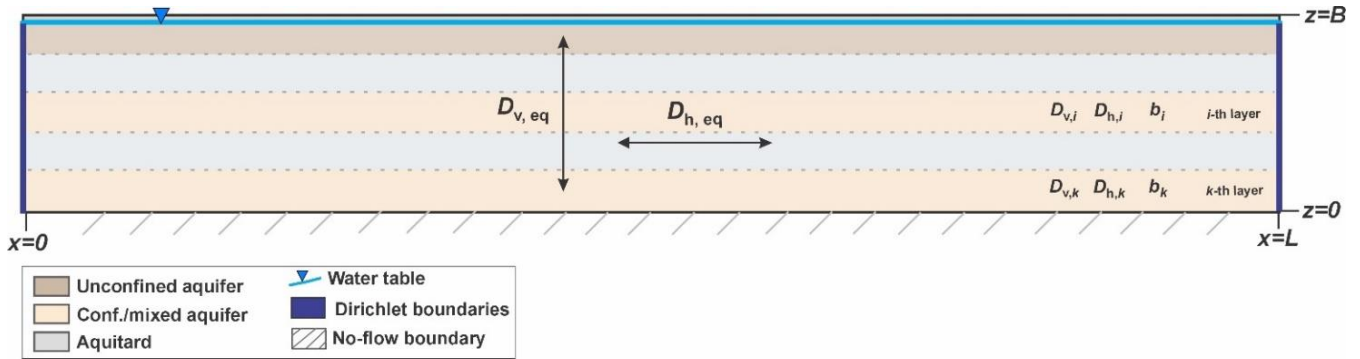


Figure 2. A homogenous equivalent of the multilayer system is assumed with two principal directions of anisotropy in the hydraulic diffusivity: bedding-parallel (approximately horizontal) and bedding-normal (approximately vertical).

The lateral boundaries are defined by Dirichlet conditions at $x = 0$ and $x = L$. The top boundary at $z = B$ is a prescribed water table consistent with the lateral boundaries, while the bottom boundary at $z = 0$ is a no-flow condition. The initial condition is a steady state head distribution satisfying the initial boundaries, $h_0(x, z)$. The transient state is induced by a step change in the lateral and top boundaries at $t = 0$, driving the system from the initial conditions to a final equilibrium state satisfying the



270 final boundaries, $h_{\infty}(x, z)$. This step change is used to represent a fast response of the water table in the unconfined areas to a step change in recharge rates.

We are interested in estimating the response time for the system to re-equilibrate after this change in boundary conditions, with focus on the deepest confined aquifer unit. For this, we approximate this multilayer system with an equivalent homogeneous
275 medium with anisotropic diffusivities in the horizontal and vertical directions. Hydraulic heads, $h(x, z, t)$, are then governed by Eq. (12).

$$\frac{\partial h}{\partial t} = D_{h,eq} \frac{\partial^2 h}{\partial x^2} + D_{v,eq} \frac{\partial^2 h}{\partial z^2}, \quad 0 < x < L, 0 < z < B, \quad t > 0 \quad (11)$$

280 We use this two-dimensional linear diffusion equation to solve for the time constants of the system using separation of variables and eigenfunction decomposition, following the approach described in Zill and Cullen (2009, p.467). The dominant time constant of the system is obtained for the fundamental mode and expressed in terms of system length, total thickness, and equivalent hydraulic diffusivities. The full mathematical derivation is presented in Appendix A. The dominant time constant of this two-dimensional homogeneous, anisotropic flow system is given by Eq. (12), where $\tau_{h,eq}$ and $\tau_{v,eq}$ can be interpreted
285 as equivalent horizontal and vertical time constants, as presented in Eq. (13) and (14), respectively. Therefore, the two-dimensional time constant corresponds to half the harmonic mean of the equivalent time constants. Since harmonic means are driven by minimum values, the response of the system is then controlled by the direction with the fastest pressure propagation.

$$\tau_{2D} = \frac{1}{\frac{1}{\tau_{h,eq}} + \frac{1}{\tau_{v,eq}}} \quad (12)$$

290 $\tau_{h,eq} = \frac{1}{\pi^2} \frac{L^2}{D_{h,eq}} \quad (13)$

$$\tau_{v,eq} = \frac{4}{\pi^2} \frac{B^2}{D_{v,eq}} \quad (14)$$

As discussed in the previous section, response times exhibit a spatial structure and, for this system, they are generally expected to increase with depth. The expression in Eq. (12) provides only the dominant time constant for a two-dimensional,
295 homogeneous, anisotropic flow system subject to a step change in boundary conditions. When applied to a multilayer system using equivalent diffusivities obtained through layer averaging, it therefore represents the longest response time in the system, which is expected to occur in the deepest layer.

The equivalent homogeneous anisotropic approximation remains valid when layers are hydraulically coupled through diffusive
300 aquitards. However, when diffusive hydraulic coupling becomes weak, the response resolves independently in each layer since



the governing equations are only weakly coupled through small leakage terms (Hantush, 1960; Neuman and Witherspoon, 1969). To assess the applicability of this approximation to represent hydraulic diffusion in a multilayer system, we propose a dimensionless criterion to characterize the degree of inter-layer diffusive hydraulic coupling by comparing two response timescales: (i) the equivalent horizontal time constant of the system, defined in Eq. (13), and (ii) the time constant of the unit with the highest vertical diffusive resistance [$T L^{-1}$], namely the ratio of layer thickness to its vertical diffusivity, corresponding to the slowest-responding aquitard within the multilayer system, which can be computed from Eq. (8). The ratio between these two timescales, r , is defined in Eq. (15).

$$r = \frac{\tau_{h,eq}}{\max_i(\tau_{v,Aquitard,i})} \quad (15)$$

When $r \geq 1$, vertical diffusion through aquitards is comparable to, or faster than the overall horizontal propagation of pressure changes. The multilayer system therefore behaves as a coupled continuum for hydraulic diffusion. In this regime, response times vary continuously with depth, and the longest response time is expected in the deepest unit. Conversely, when $r < 1$, vertical diffusion is much slower than the overall horizontal propagation of pressure changes. At least one aquitard layer effectively limits the vertical response of the system, such that pressure changes cannot propagate across its thickness within the timescale of horizontal adjustment. The multilayer system therefore behaves as a set of weakly coupled layers. In this regime, one-dimensional formulations are sufficient to estimate the time constants of the individual aquifer and aquitard units, respectively.

These two regimes provide the basis for defining two complementary response timescales: a *unit-scale response time*, which characterizes the response of the deepest confined aquifer within the multilayer system, and a *basin-scale response time*, which represents the longest response time expected anywhere within the multilayer system. When $r \geq 1$, the response of a target confined aquifer depends not only on its own properties but also on those of its overlying confining units and is therefore given by $3\tau_{2D}$. At the same time, this response time also represents the basin-scale response time because, in coupled systems, the longest response times are expected to occur in the deepest unit. When $r < 1$, the unit-scale response time of the deepest aquifer is instead given by the one-dimensional formulation given in Eq. (7), $3\tau_{h,Aquifer,k}$. The basin-scale response time is then controlled by the slowest-responding aquitard and as estimated from Eq. (8), $\max_i(3\tau_{v,Aquitard,i})$. The unit-scale and basin-scale response times can therefore be defined by the piecewise functions provided in Eq. (16) and eq. (17), respectively.

$$t_{r_{Unit, 5\%}} = \begin{cases} 3\tau_{2D}, & \text{if } r \geq 1 \\ 3\tau_{h,Aquifer}, & \text{if } r < 1 \end{cases} \quad (16)$$



$$t_{r_{Basin, 5\%}} = \begin{cases} 3\tau_{2D}, & \text{if } r \geq 1 \\ \max_i(3\tau_{v,Aquitard,i}), & \text{if } r < 1 \end{cases} \quad (17)$$

335 The proposed analytical framework is validated as an order-of-magnitude estimate of unit-scale and basin-scale response times through comparison with numerical simulations. A synthetic conceptual and numerical model of a regional multilayer groundwater system is developed for this purpose. We compare the results of the numerical simulations with different analytical solutions to delineate the different hydrogeological contexts where each formulation is more suitable. Additionally, different configurations of hydraulic properties, horizontal extent, thicknesses, and number of layers are also evaluated. The
 340 validation and applicability of the analytical framework is discussed in more detail in the Supplementary Material S2.

Finally, we provide an example application of the proposed two-dimensional formulation to the Aquitaine Basin, France, which reasonably satisfies the main analytical assumptions. For this, we compile data on hydrodynamic parameters, thickness, and horizontal extent of each hydrogeological unit within the Cenozoic sequence of the Aquitaine Basin. We apply a Monte
 345 Carlo approach to account for uncertainties in the input parameters and provide a distribution of the hydraulic response times for the Oligocene and Eocene confined aquifers. We compare our results with previously published estimates and discuss their suitability and implications in relation to independent studies. More information about this implementation is presented in the Supplementary Material S3.

3.2 A synthetic conceptual and numerical model

350 A conceptual model of a regional multilayer groundwater basin was developed and translated to a two-dimensional numerical flow model to investigate response times to large-scale step changes in recharge rates. To inform this model, a compilation of the characteristics of several major groundwater basins worldwide was carried out to identify the orders of magnitude of key parameters, including the hydraulic properties of the main hydrogeological units, their spatial extent, thickness, and recharge rates. This compilation is presented in the supplementary information S1.

355 The mean flow path lengths of major regional aquifers compiled by Rousseau-Gueutin et al. (2013) is about 600 km, ranging from the orders of 100 km, as in the Aquitaine Basin, to more than 1700 km in the most extensive systems, such as the Great Artesian Basin. Total basin thicknesses reported in the literature, many of which were compiled by Gun (2022) and Margat and Gun (2013), range from several hundred metres to more than 10 km when the full sedimentary cover is considered.
 360 Additionally, the mean thickness of individual hydrogeological units typically is in the order of tens to hundreds of metres.

The synthetic case in this study represents a generic regional multilayer groundwater basin composed of five hydrogeological units. The first unit consists of an unconfined aquifer overlying an interbedded sequence of formations with contrasting hydraulic conductivities (Figure 3). These are classified as aquifers or aquitards based on their hydraulic conductivity relative



365 to adjacent layers, reflecting their role in controlling vertical flows. Unit 1 corresponds to an unconfined aquifer while Units 2
 to 5 correspond to mixed units, with unconfined and confined portions. The total thickness of the basin (B) goes up to 900
 meters, with aquifers having a maximum thickness of 200 meters and aquitards of 150 meters. The units dip from the outcrop
 areas on the right toward the main lateral boundary of the system on the left. This boundary can be interpreted, in a general
 sense, as representing the exchanges of the groundwater system with the ocean, a major surface water body, or a regional
 structural limit, such as a fault or lateral contact (Q_{GHB}). The maximum flow length (L) from the outcrops to the main lateral
 370 boundary is 600 km. Each unit has an outcrop area to allow for direct inflows via recharge from precipitations (R). Leakage
 flows through aquitards is also considered (Q_{lk}), as well as the surface-groundwater interactions (Q_{surf}).

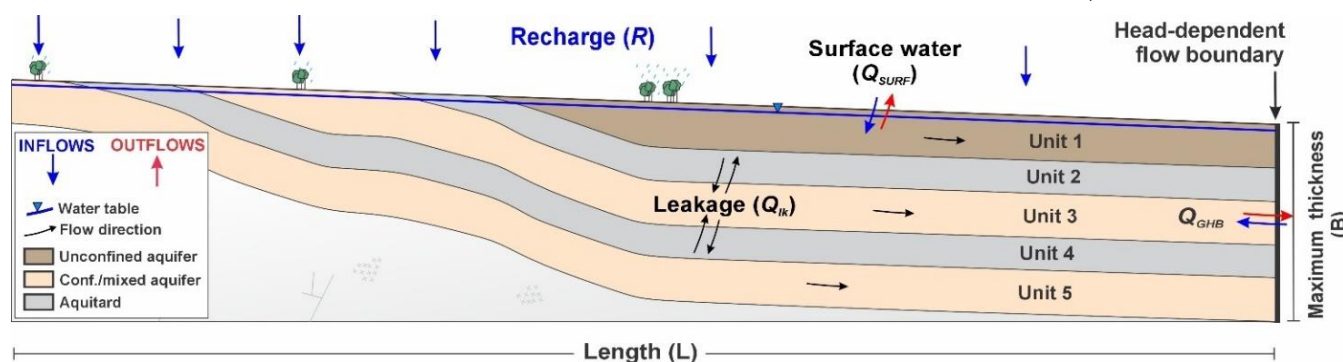


Figure 3. Conceptual model of the synthetic case showing hydrogeological units and considered processes.

375 We aim to determine how long the groundwater system takes to respond to a large-scale change in recharge rates over the long
 term. For this purpose, a 2D numerical flow model of the synthetic case was developed using MODFLOW 6 and Flopy (Bakker
 et al., 2018; Hughes et al., 2024; Langevin et al., 2017, 2024). Two principal directions of flow and pressure propagation are
 considered in our analysis: along the stratigraphic units (bedding-parallel), which is assumed to be approximately horizontal,
 and perpendicular to the units (bedding-normal), which is considered approximately vertical.

380 A structured grid with cells of 1 km in the horizontal direction was created. Each hydrogeological unit was subdivided into
 five model layers for a better representation of vertical hydraulic gradients. A smooth topographic gradient of 200 meters over
 the length of 600 km was used to represent a rather flat extensive basin. Surface water systems were accounted for using the
 drain package (DRN) all over the model domain with a fixed conductance of $1 \text{ m}^2 \text{ d}^{-1}$. The drain elevation is set to the
 385 topographic elevation and is aimed at representing the regional surface drainage base-level of the system (Cuthbert et al.,
 2023). The recharge package (RCHA) was used to simulate recharge from precipitation in all outcropping cells. A spatially
 uniform recharge rate of 200 mm yr^{-1} was applied to represent a humid system with shallow water tables that remain above
 the drainage base-level. The right lateral boundary of the system is simulated with the General Head Boundary package (GHB)
 using a conductance corresponding to the transmissivity of each model layer (Anderson et al., 2015). The reference head is set
 390 to the topographic surface at these boundaries. Finally, leakage flows are computed using zonal budgets per hydrogeological
 unit.



Each hydrogeological unit was assumed to be homogeneous and isotropic to enable direct comparison of estimated response times with analytical formulations. Cross-sections of the model grid and hydrogeological units are shown in Figure 4. A specific storage value of $1 \times 10^{-5} \text{ m}^{-1}$ was used to simulate compressible storage, and a specific yield of 0.15 was used for unconfined storage. The Storage Package (STO) was set to represent unconfined storage in the topmost active cells and to use a convertible storage formulation elsewhere in the model.

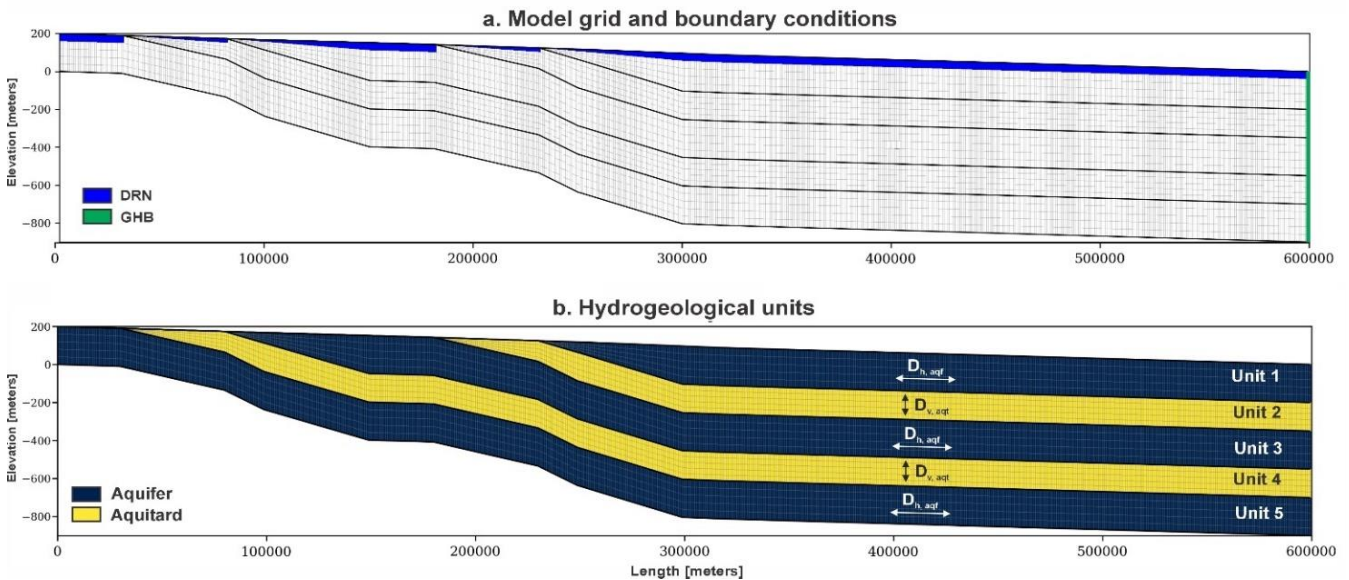


Figure 4. Numerical groundwater flow model. (a) Cross-section of the model grid and boundary conditions. (b) Hydrogeological units.

The Node Property Flow Package (NPF) was set to represent varying saturated thickness for unconfined cells. The influence of the hydraulic diffusivity of each unit on the response times was investigated through a sensitivity analysis. Hydraulic conductivities were varied while keeping storage parameters fixed, since conductivity tends to spread over more orders of magnitude as compared to storage coefficients. For aquifers, horizontal hydraulic conductivities ranged from 1×10^{-7} to $1 \times 10^{-3} \text{ m s}^{-1}$, while for aquitards, vertical hydraulic conductivities were varied from 1×10^{-13} to $1 \times 10^{-7} \text{ m s}^{-1}$. Within each simulation, the three aquifer units were assigned the same hydraulic conductivity and, likewise, the two aquitard units shared a common conductivity value. This led to a total of 117 hydraulic diffusivity scenarios.

For clarity, the terms aquifer horizontal diffusivity ($D_{h,aqf}$) and aquitard vertical diffusivity ($D_{v,aqt}$) are used throughout the text. These represent the diffusivity in the dominant direction of flow and pressure propagation for each layer type. Both are defined using the properties of confined cells and correspond to the ratio of hydraulic conductivity to specific storage. In each simulation, all aquifer layers share the same $D_{h,aqf}$, and all aquitard layers share the same $D_{v,aqt}$ (Figure 4). With the defined hydraulic parameter scenarios, $D_{h,aqf}$ ranges from 1×10^{-2} to $1 \times 10^2 \text{ m}^2 \text{ s}^{-1}$, and $D_{v,aqt}$ varies from 1×10^{-8} to $1 \times 10^{-2} \text{ m}^2 \text{ s}^{-1}$.



3.3 Numerical estimation of hydraulic response times

415 Hydraulic response times can be defined as given in Eq. (18) (Carr and Simpson, 2018), which can be evaluated with a numerical model by using the simulated hydraulic heads at each model cell X .

$$\frac{r(X, t_f)}{r(X, 0)} = \frac{h(X, t_f) - h_\infty(X)}{h(X, 0) - h_\infty(X)} = \delta \quad (18)$$

420 Equation (18) was solved for the synthetic case using a response threshold of $\delta = 5\%$, allowing comparison with analytical estimates of 3τ . Response times were calculated for two stress scenarios: a 100%, and a 50% step decrease in recharge rates. These changes are uniform across the model domain and aim to represent a large-scale climatic shift. The sensitivity analysis was performed to examine how response times depend on the diffusivity of both aquitards and aquifers, using the scenarios described in the previous section.

425 Figure 5 illustrates the followed approach. Starting from an initial steady-state (Figure 5a), a transient simulation was performed to track the evolution of hydraulic heads, $h(X, t)$, from the initial condition, $h(X, 0)$, toward a final state $h(X, t_f)$ (Figure 5b). A steady-state simulation was then conducted for each scenario to determine the final equilibrium state after the stress was applied, $h_\infty(X)$ (Figure 5c). This allowed the proportion of residual diffusion to be computed at each model cell and time step. The time discretization consisted of a single stress period with 200 time steps. A time step multiplier was used, starting with a first step of one day and increasing geometrically. This setup provided a total simulation time sufficient for the system to reach the prescribed response time threshold of 5% residual diffusion. For each model cell, the response time is determined as the time at which this threshold is reached. Cells in which the head differences between the initial steady state and the final steady state were less than 10^{-3} m were defined as “irresponsive” and excluded from the response time calculation
 430
 435 to avoid the influence of numerical noise in the results.

This approach allows for a detailed investigation of the spatial distribution of response times, which can be analysed statistically both at the unit and basin scales. The goal is to compare the numerical results with analytical estimates. Since analytical solutions rely on the dominant time constant which represents the long-term response within a system, an equivalent
 440 indicator is selected from the numerically derived spatial distribution of response times. For this, the 95th percentile of the simulated response time distribution within a region of interest is calculated using zonal statistics. This metric represents the time at which 95% of the region has responded, while avoiding the influence of highly localized effects or potential numerical noise associated with maximum values.

445 We apply available analytical formulations to estimate the *unit-scale response time* of the deepest confined aquifer (Unit 5), using the known geometry and hydraulic parameters of our synthetic model. We compare the analytical estimations with the

95th percentile of the simulated response times over the same unit. Since the aquifer is limited at its outcrop and on the right side by head-dependent flow boundaries, the unit is therefore approximated by a domain of length L bounded by Dirichlet conditions. For this reason, the formulation in Eq. (7) is used. In addition, the mixed aquifer formulation in Eq. (9) by Rousseau-Gueutin et al. (2013) is also evaluated, since the unit has unconfined and confined portions of known lengths. Finally, we also use the proposed two-dimensional unit-scale formulation presented in Eq. (16) to assess its performance to estimate the response times of the deepest confined aquifer within a multilayer system.

Similarly, the *basin-scale response time* is defined as the 95th percentile of the simulated response time spatial distribution across the entire model domain, and it is used to validate the proposed two-dimensional analytical formulation presented in Eq. (17).

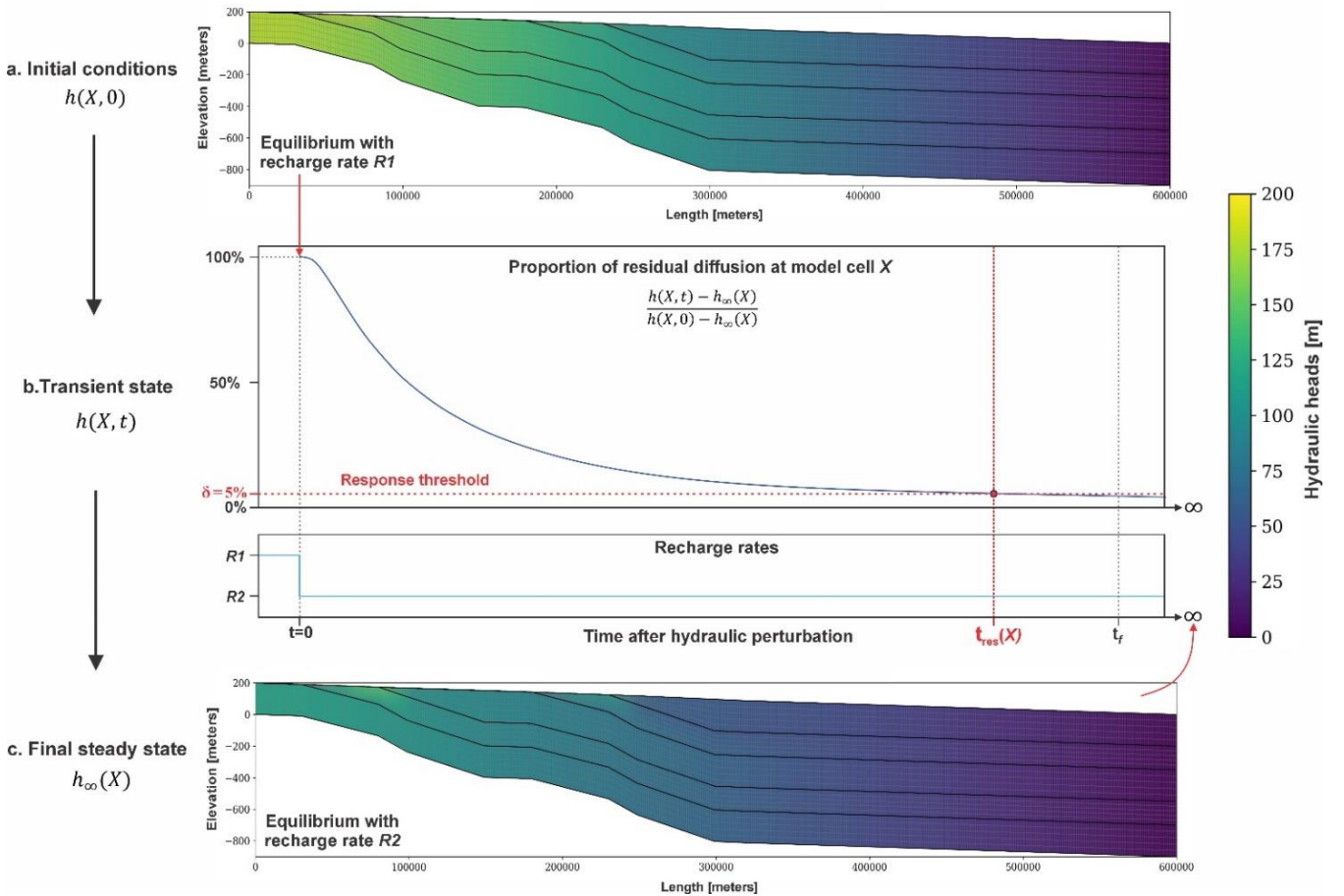


Figure 5. Estimation of response times using the numerical model. a) Initial conditions corresponding to the steady state heads in equilibrium with recharge rate $R1$. b) Transient simulation following a step change from recharge rate $R1$ to $R2$. The proportion of residual diffusion and hydraulic response times are determined at each model cell. c) Final steady state heads in equilibrium with recharge rate $R2$.



4 Results

4.1 Simulated response times to a 100% step decrease in recharge rates

For this case, the initial state corresponds to an equilibrium state with a recharge rate of 200 mm yr^{-1} that suddenly changes to
 465 a no-recharge condition. This case represents an extreme end member and follows the analytical and numerical model solved
 by Rousseau-Gueutin et al. (2013) to obtain Eq. (9). The 100% decrease in recharge leads to a hydrostatic condition and the
 long-term interruption of surface-groundwater interactions at the regional scale. Simulated water tables fall below the regional
 drainage base level, causing the disappearance of perennial streams as the drain boundaries do not allow inflows to the system.
 This configuration is intended to represent a large-scale transition from humid to arid conditions, similar to the shifts that have
 470 been discussed for some regions, such as the Sahara during the early Holocene (Rousseau-Gueutin et al., 2013; Tierney et al.,
 2017; Vasseur et al., 2015).

A zonal analysis for the deepest confined aquifer is performed. Figure 6a presents the contour map of the 95th percentile of the
 simulated response time distribution for the deepest aquifer (Unit 5) as a function of equivalent horizontal and vertical
 475 diffusivities. Under these conditions, simulated response times are in the order of 10^4 to 10^8 years, which agrees with previous
 estimations for systems in arid to hyper-arid regions (Cuthbert et al., 2019). These extremely long response times can be
 explained by two main mechanisms. First, in the absence of active surface boundaries, the effective characteristic length in the
 unconfined areas increases considerably. Therefore, the vertical modes of equilibration are not activated, and the simulated
 response times show no dependence on aquitard vertical diffusivity. The adjustment is instead governed by horizontal pressure
 480 propagation toward distant lateral boundaries. Second, desaturation in the unconfined areas becomes the dominant storage
 mechanism, so the long-term response is controlled primarily by specific yield, which is typically two to three orders of
 magnitude higher than the storage coefficients of confined units (Rousseau-Gueutin et al., 2013). Larger characteristic lengths
 and storage dominated by specific yield imply longer response times in the unconfined areas, after which the signal propagates
 toward the deeper parts of the basin. This behaviour is expected in arid regions with poorly-developed surface drainage
 485 networks, where water tables are flux-controlled and response times tend to be longer (Cuthbert et al., 2019; Haitjema and
 Mitchell-Bruker, 2005). Thus, at the regional scale we describe this type of response in the unconfined areas as *storage-driven*,
 controlled by desaturation and horizontal pressure propagation in the absence of surface boundaries.

We compare these numerical estimates with the analytical formulations for the deepest aquifer unit. As shown in Figure 6b,
 490 the one-dimensional formulation for a confined aquifer given in Eq. (7) underestimates response times in all cases by
 approximately two orders of magnitude, as shown by the mean absolute error of around 2.4 log years. This corresponds to the
 difference in orders of magnitude of the specific yield and the compressible storage coefficient used in our simulations. This
 discrepancy arises because this analytical formulation represents a fundamentally different process: it assumes full one-

dimensional confined flow with no explicit unconfined portion, thus expecting compressible storage as dominant mechanism, while the simulations are dominated by desaturation in the unconfined areas.

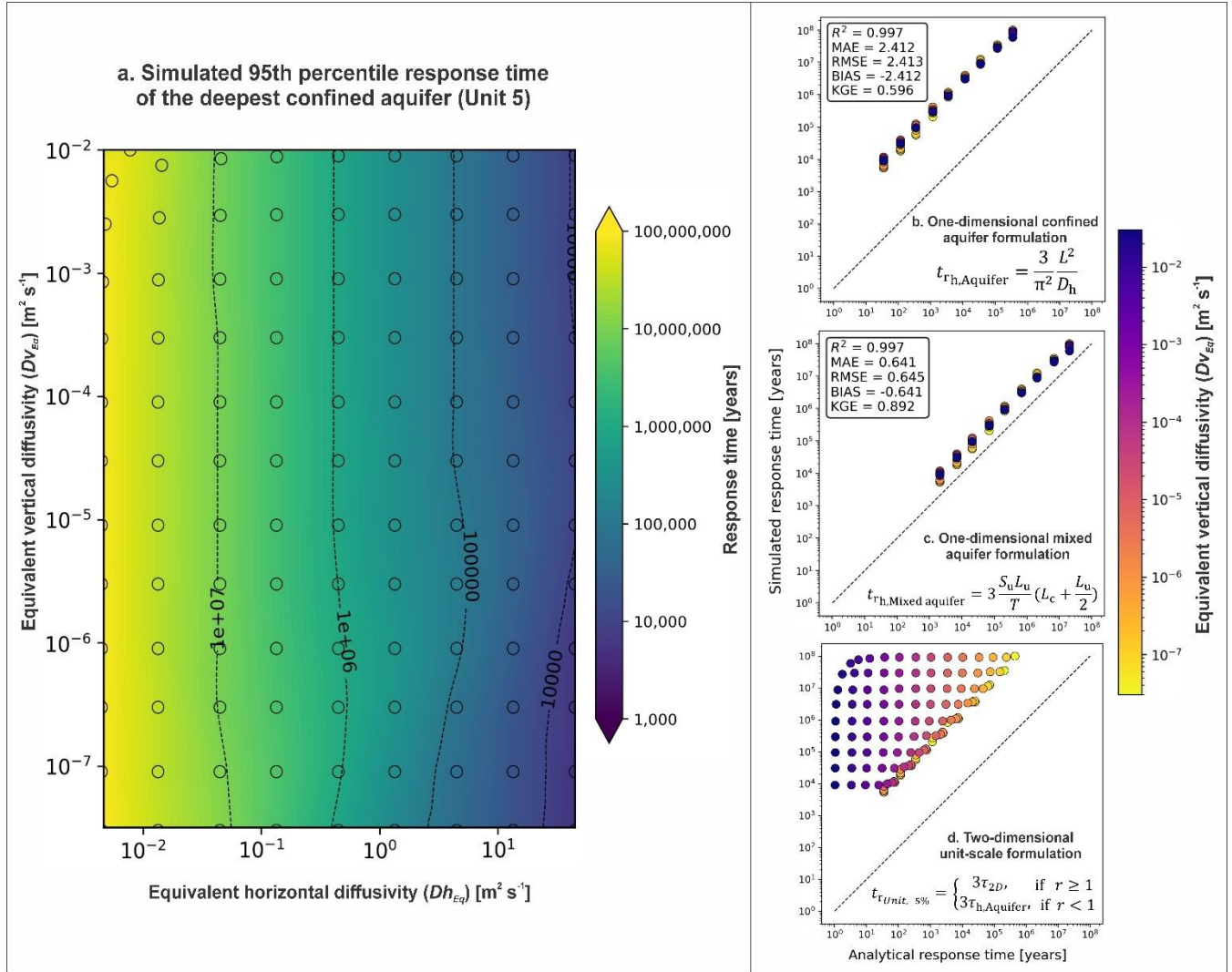


Figure 6. Simulated unit-scale response times to a 100% recharge decrease and scatter plots of simulated versus analytical estimates. (a) 95th percentile of the simulated response time distribution for Unit 5 as a function of the equivalent horizontal diffusivity ($D_{h,eq}$) and equivalent vertical diffusivity ($D_{v,eq}$). (b) Use of the one-dimensional confined aquifer formulation. (c) Use of the one-dimensional mixed aquifer formulation. (d) Use of the two-dimensional unit-scale formulation. R^2 : Coefficient of determination. MAE: Mean Absolute Error. RMSE: Root Mean Squared Error. KGE: Kling–Gupta efficiency.

In contrast, the one-dimensional formulation for a mixed aquifer presented in Eq. (9) provides response time estimates that differ by less than one order of magnitude from the simulated values, as shown by the mean absolute error of around 0.6 log-years (Figure 6c). The simulated response times remain generally higher, likely because the numerical model does not fully satisfy the analytical assumptions, such as strict one-dimensional flow. These results indicate that the analytical formulation for mixed aquifers by Rousseau-Gueutin et al. (2013) provides good order-of-magnitude estimates in this case, even when



compared to a system that includes additional processes, such as explicit aquitards and mixed boundary conditions. However, its applicability is restricted to regional systems characterized by deep water tables and the absence of active surface boundaries, as it is the case in arid-region basins. A misconception is that mixed aquifer systems inherently exhibit long response times due to the existence of an unconfined portion with higher storage coefficient (Rousseau-Gueutin et al., 2013). Nevertheless, our simulations show that these longer response times reflect the absence of active surface boundaries and a response dominated by unconfined storage, representing a particular hydrogeological context rather than general behaviour of mixed aquifers, as will be demonstrated in the following section.

Finally, the two-dimensional unit-scale formulation defined in Eq. (16) is used for comparison with the simulation results (Figure 6d). It provides the same results to the one-dimensional confined-flow formulation in cases where aquitard vertical diffusivity is low in relation to aquifer horizontal diffusivity, this is, for cases where $r < 1$. As aquitard diffusivity increases, the formulation increasingly underestimates the response times. This is due to the absence of active surface boundaries in the simulation, which prevents the activation of vertical diffusion modes and therefore does not allow vertical equilibration. This is expected since the response is primarily controlled by horizontal pressure propagation, and aquitard vertical diffusivity does not affect the simulated response times. This highlights a limitation of the two-dimensional formulation in this specific hydrogeological setting, as its vertical component does not contribute to the system response when surface boundaries are inactive.

4.2 Simulated response times to a 50% step decrease in recharge rates

For this case we implement a smaller recharge reduction to allow surface boundaries to remain active. The initial state corresponds to an equilibrium with a uniform recharge rate of 200 mm yr⁻¹ that decreases stepwise to 100 mm yr⁻¹. Cross-sections of the spatial distribution of response times for some hydraulic diffusivity scenarios are presented in Fig. 7. In all cases, aquifer horizontal diffusivity ($D_{h,aqf}$) is 1 m² s⁻¹, while the aquitard vertical diffusivity ($D_{v,aqt}$) varies to illustrate its impact on system response.

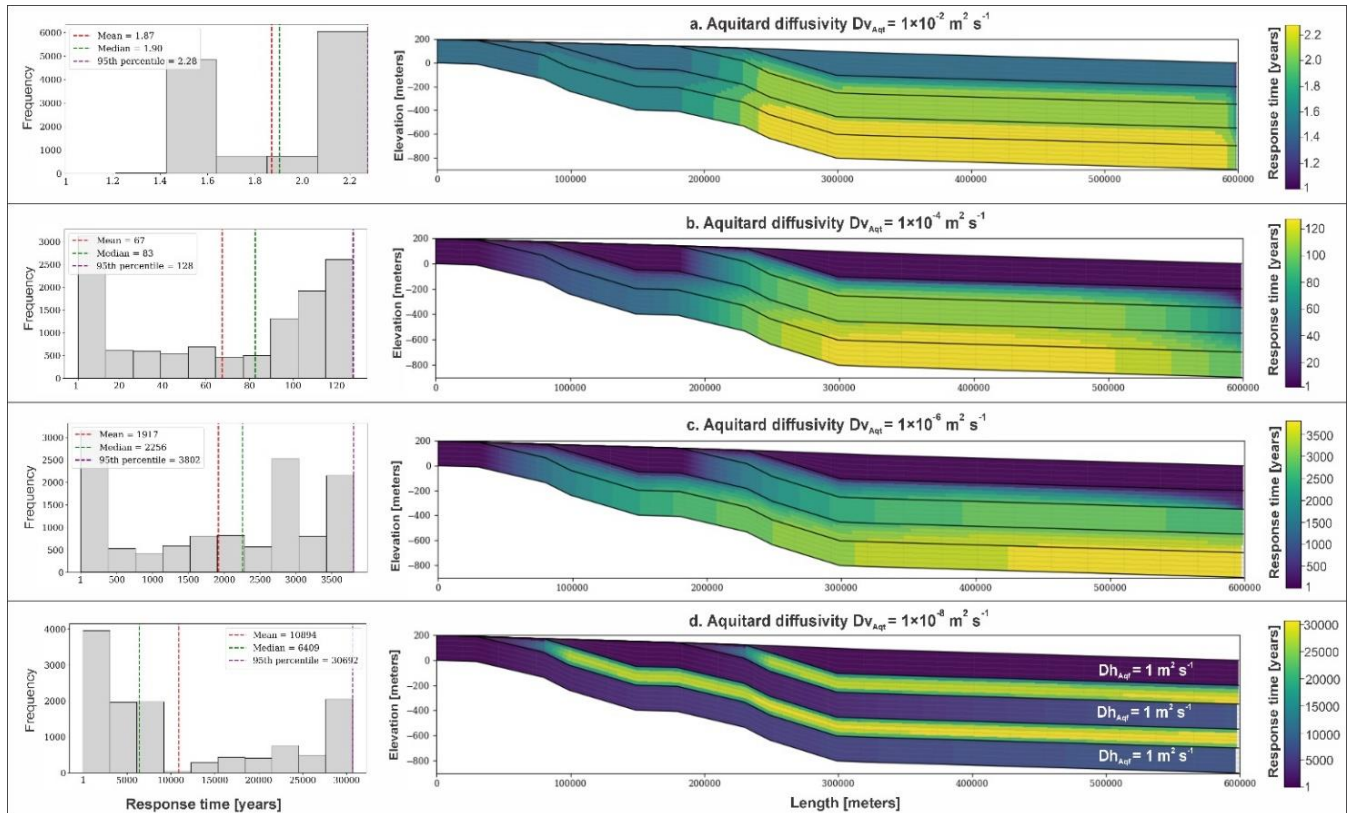
Across all hydraulic diffusivity scenarios, response times in the unconfined areas remain around one year. The simulated 50% decrease in recharge rates does not interrupt surface-groundwater interactions, with water tables remaining predominantly above the regional drainage level. Consequently, the response in unconfined areas is largely governed by the properties of the surface boundaries, which remain unchanged across all simulations. This is representative of humid settings where water tables are shallow, mainly topography-controlled, and response times are short due to well-developed surface drainage networks and short flow path lengths (Cuthbert et al., 2019; Haitjema and Mitchell-Bruker, 2005). In contrast to the previous case, we describe this type of response in the unconfined areas as *boundary-driven*, controlled by the degree of hydraulic coupling of the groundwater system to the surface boundaries, which depends on their conductance and spatial density.



540 The main differences occur in the deep parts of the basin, where aquitard vertical diffusivity controls the response structure of the system. In Figure 7a, aquitards have the highest diffusivity of $1 \times 10^{-2} \text{ m}^2 \text{ s}^{-1}$, corresponding to a strong hydraulic diffusive coupling between units. Under these conditions, response times in the deeper parts of the basin increase only slightly, reaching about two years. When aquitard diffusivity is decreased to $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$, hydraulic coupling weakens and response times in the deeper part of the basin increase up to the order of 100 years, while the unconfined areas remain unaffected, as shown in

545 Figure 7b. With a further decrease to $1 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$, the longest response times in deeper zones rise to around 3500 years, as depicted in Figure 7c. Finally, with the lowest aquitard diffusivity of $1 \times 10^{-8} \text{ m}^2 \text{ s}^{-1}$, the spatial distribution of the response becomes strongly stratified. Response times reach around 3000 to 7000 years in the confined aquifer units and up to about 30,000 years in the aquitard units, indicating that the hydraulic coupling is weak enough so that hydrogeological units respond independently from one another, as shown in Figure 7d. At the basin scale, response times present a bimodal distribution,

550 showing the contribution of short responses in unconfined areas and longer ones in the deeper parts of the basin, as shown in the histograms of Figure 7.



555 **Figure 7. Response time cross-section and histograms for a step decrease of 50% in recharge rates. All cases have the same aquifer hydraulic diffusivity ($D_{h,aqt} = 1 \text{ m}^2 \text{ s}^{-1}$). Each case uses its own color ramp scale to better represent the spatial distribution of response times as aquitard vertical diffusivity varies. (a) $D_{v,aqt} = 1 \times 10^{-2} \text{ m}^2 \text{ s}^{-1}$ (b) $D_{v,aqt} = 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ (c) $D_{v,aqt} = 1 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ (d) $D_{v,aqt} = 1 \times 10^{-8} \text{ m}^2 \text{ s}^{-1}$.**

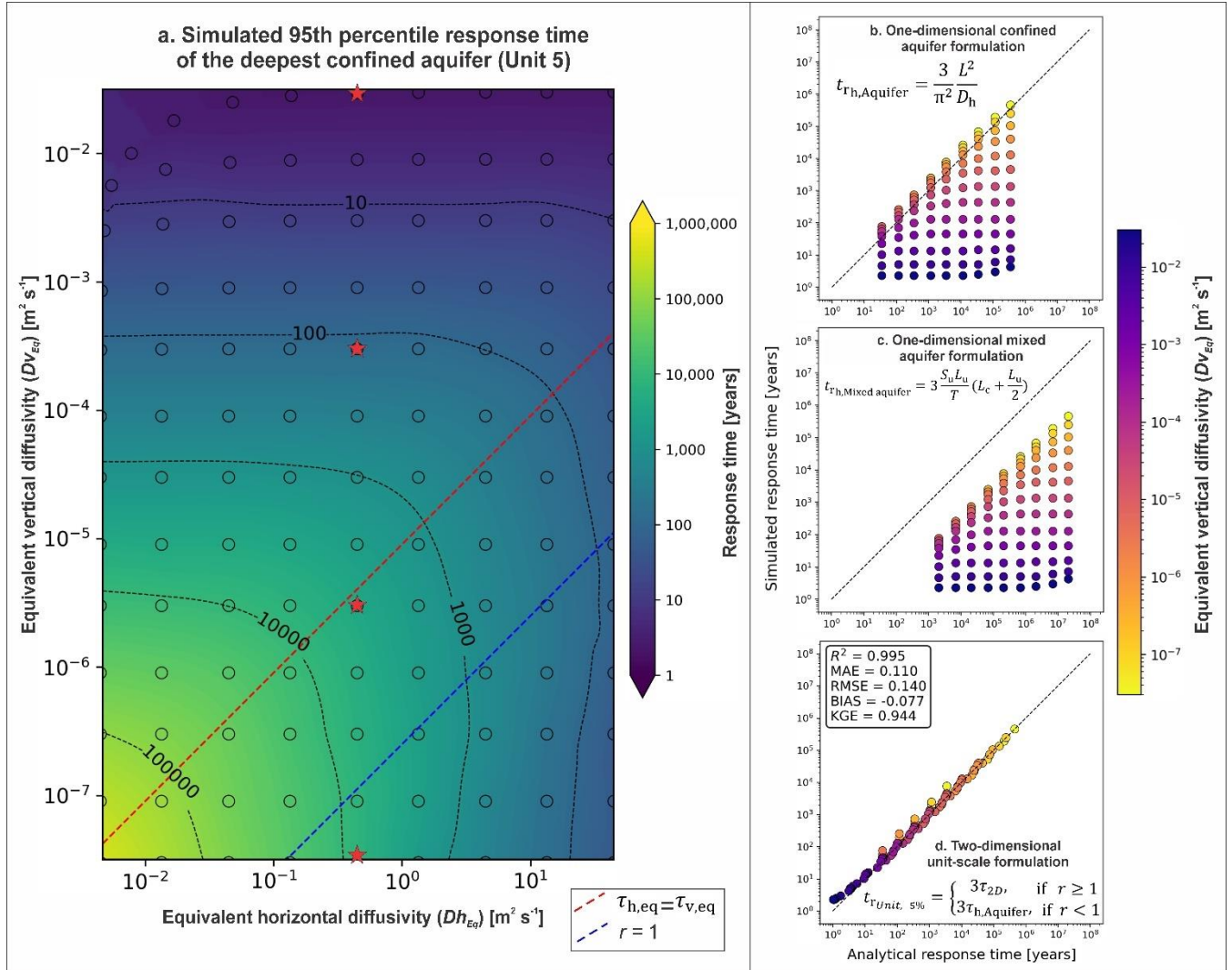


A contour map of the 95th percentile of the simulated response times for the deepest aquifer (Unit 5) is presented in Figure 8a as a function of the equivalent horizontal and vertical diffusivities of the system. Two regimes can be distinguished, reflecting the interplay between pressure propagation along the vertical and horizontal directions. The red dashed line in the contour map represents the cases in which the equivalent vertical and horizontal time constants are equal. Above this threshold, response times are controlled by the equivalent vertical diffusivity. In this regime, aquitard vertical diffusivity is sufficiently high for hydraulic perturbations to propagate more rapidly across the units than along them. Below this threshold, response times are controlled by the equivalent horizontal diffusivity, implying that when aquitard diffusivity is sufficiently low, pressure propagates faster along the stratigraphic bedding of the aquifer. Additionally, the blue dashed line indicates the threshold for the validity of the equivalent homogeneous anisotropic approximation ($r = 1$). Above this threshold, the hydrogeological units are well-coupled, and the longest response times are expected within the deepest unit of the multilayer system (e.g., Figure 7a to Figure 7c). Below this threshold, hydraulic coupling between hydrogeological units is weak and the longest response times are expected to be found within the slowest-responding aquitard unit (e.g., Figure 7d).

We compare the simulated response times within Unit 5 with those from analytical formulations. As shown in Figure 8b, the one-dimensional formulation for confined aquifers given in Eq. (7) provides good order-of-magnitude estimates in cases where aquitard vertical diffusivity is low in relation to aquifer horizontal diffusivity, this is, for cases where $r < 1$. In such cases, the one-dimensional flow assumption remains valid. As aquitard vertical diffusivity increases the analytical formulation increasingly overestimates response times, by up to four orders of magnitude in some cases. Under one-dimensional flow, the characteristic length L corresponds to the distance between system boundaries along the flow direction. For a confined or mixed aquifer overlain by a fully impermeable cover, this assumption is reasonable, and L can be approximated as the distance between the outcrop and a lateral boundary along the regional flow path. However, as aquitards become more diffusive, this assumption no longer holds. Higher vertical diffusivity enhances the transmission of pressure changes across stratigraphic units, strengthening the hydraulic coupling between deep aquifer units and surface boundaries, leading to shorter response times than predicted by purely one-dimensional flow assumptions.

Additionally, Figure 8c shows that the mixed-aquifer formulation given in Eq. (9) overestimates response times in all scenarios, by at least one order of magnitude and by up to seven orders of magnitude in cases with high aquitard vertical diffusivity. This overestimation arises from the nature of the processes represented in the analytical formulation. Rousseau-Gueutin et al. (2013) derived the solution for a mixed system relaxing from an initial state with shallow water tables in the unconfined portion to a no-recharge condition. In their conceptualization, flow is strictly one-dimensional, with fully impermeable confining units and no surface boundaries along the system extent. As a result, the system response is dominated by horizontal propagation and unconfined storage: the stress is imposed in the unconfined portion, where head declines cause desaturation, and the resulting signal propagates laterally toward the confined portion of the aquifer. Consequently, the analytical solution primarily reflects a storage regime controlled by specific yield. In contrast, the simulations presented here represent a system in which unconfined

response is rapid due to strong surface-groundwater coupling. The contribution of specific yield to the overall response is small and the dominant storage mechanism is the compressible storage in the deepest parts of the basin.



595 **Figure 8. (a) 95th percentile of the simulated unit-scale response time spatial distribution for Unit 5 as a function of the equivalent horizontal diffusivity ($D_{h,eq}$) and equivalent vertical diffusivity ($D_{v,eq}$). The red stars represent the configurations for the cases presented in Fig. 5. (b) Use of the one-dimensional confined aquifer formulation. (c) Use of the one-dimensional mixed aquifer formulation. (d) Use of the two-dimensional unit-scale formulation. R^2 : Coefficient of determination. MAE: Mean Absolute Error. RMSE: Root Mean Squared Error. KGE: Kling–Gupta efficiency.**

600 Finally, as presented in Figure 8d, the proposed two-dimensional unit-scale formulation given in Eq. (16) is used to correct for the effect of vertical diffusion in the cases where $r \geq 1$, showing a good agreement with the simulated response times for the deepest confined aquifer across all cases. The structure of the simulated response times implies that the adjustment is controlled by the direction of fastest pressure propagation. The harmonic mean form of this analytical formulation is consistent with this

observation, since it will be driven by the minimum between the equivalent vertical and horizontal time constants. The mean absolute error of 0.11 log years indicates an average deviation by a factor of 1.3 from the simulations. This shows that the two-dimensional formulation is useful to estimate the response times of the deepest unit within a multilayer system by accounting for the equivalent hydraulic diffusivity of the overlying confining units in cases where $r \geq 1$.

From the simulations, it is also possible to calculate the *basin-scale response time* using zonal statistics. For this, the 95th percentile of the simulated response time distribution over the whole basin is shown in Figure 9a as a function of the equivalent vertical and horizontal diffusivities. Figure 8a and Figure 9a show that unit-scale and basin-scale response times are equivalent when the multilayer system is well coupled ($r \geq 1$). In these cases, response times vary smoothly and continuously throughout the system, generally increasing with depth, such that the longest response times occur within the deepest confined aquifer. Consequently, the maximum response time at the basin scale is also representative of the response time of the deepest aquifer. However, in weakly coupled systems ($r < 1$), the longest basin-scale response times are instead found within aquitard layers.

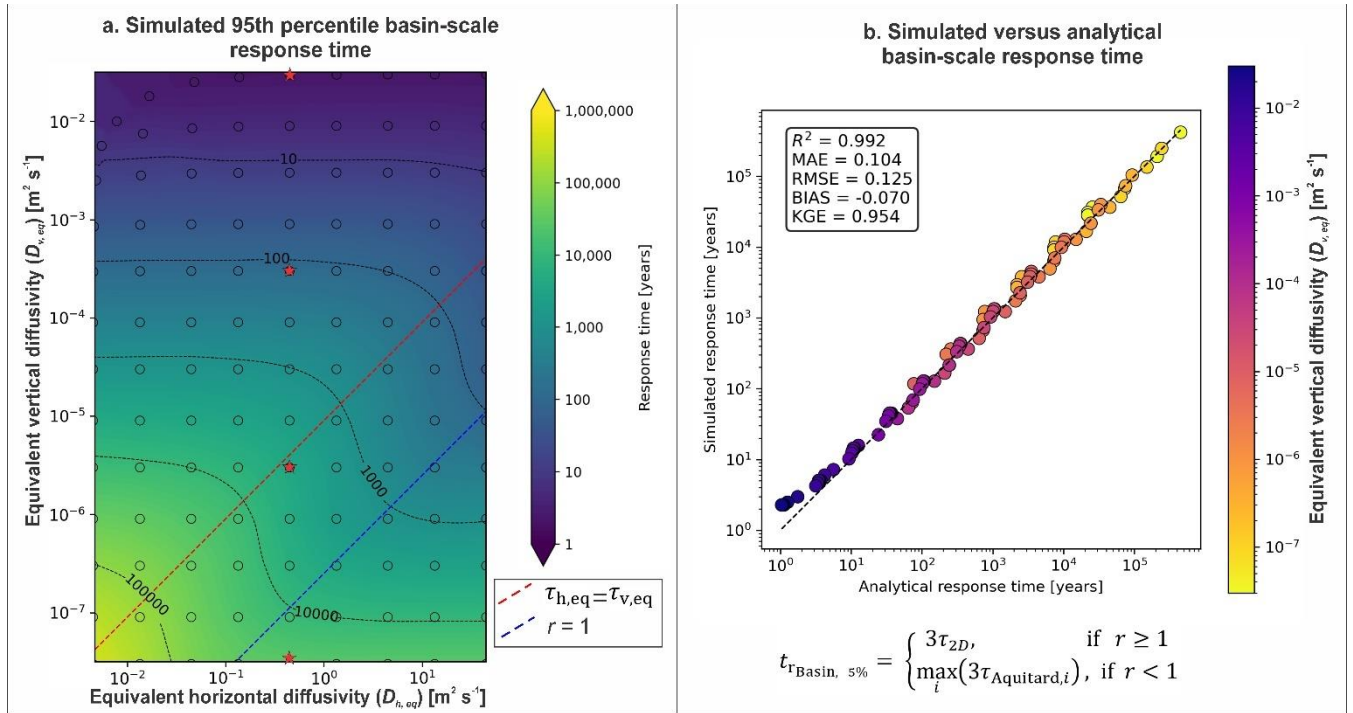


Figure 9. a) Contour maps of the simulated 95th percentile basin-scale response times as a function of the equivalent horizontal and ($D_{h,eq}$) and vertical diffusivity ($D_{v,eq}$). The red stars represent the example cases presented in Fig. 5. **(b)** Scatter plot of analytical versus simulated 95th percentile basin-scale response times. R^2 : Coefficient of determination. MAE: Mean Absolute Error. RMSE: Root Mean Squared Error. KGE: Kling–Gupta efficiency.

We compare the simulated 95th percentile basin-scale response times with the analytical estimates using the piecewise formulation from Eq. (17). Overall, there is a good agreement as illustrated in Figure 9b, with a mean absolute error of around 0.1 log years, indicating an average deviation by a factor of 1.2 from the simulations. This supports the robustness of this



analytical approach as an order-of-magnitude estimator despite deviations from analytical assumptions in geometry, structure, and boundary conditions in the numerical model. Further validation with variations of the numerical model is presented in the
 625 Supplementary Material S2.

5 Discussion

At the basin scale, two contrasting hydrogeological contexts can be identified, for which different analytical approaches are more appropriate. On the one hand, in arid regions characterized by sparse drainage networks and generally deep groundwater tables, the response of unconfined areas can be described as predominantly *storage-driven*. This behaviour is characterized by
 630 the slow horizontal propagation of hydraulic perturbations toward distant boundaries, with the long-term response controlled primarily by the specific yield of the unconfined areas. Under these conditions, the formulation for mixed aquifers proposed by Rousseau-Gueutin et al. (2013) provides good order-of-magnitude estimates of response times. Importantly, its applicability is largely independent of aquitard vertical hydraulic diffusivity because the limited extent of surface boundaries prevents the activation of vertical modes of equilibration.

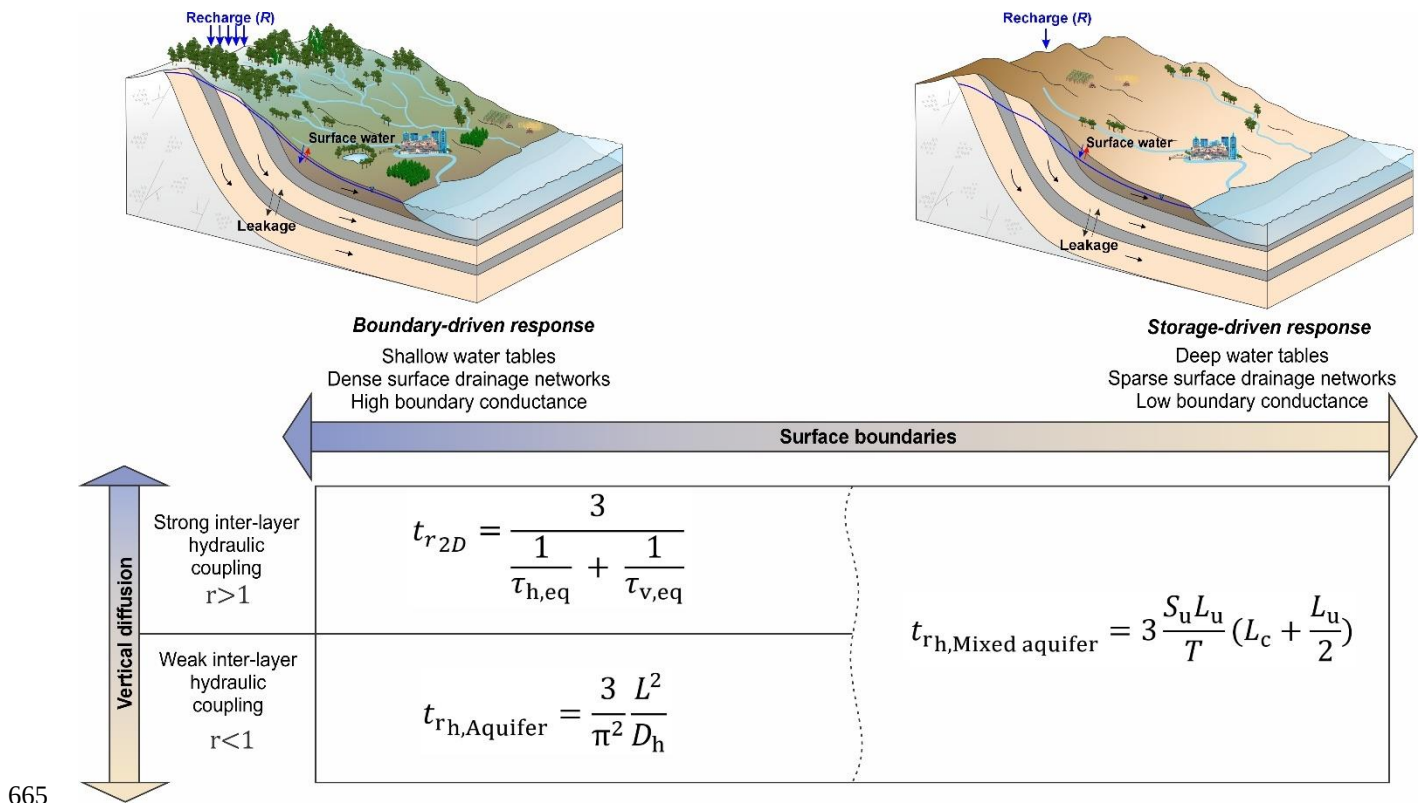
635 In contrast, systems in humid regions, characterized by shallow water tables and well-developed surface drainage networks, exhibit a predominantly *boundary-driven* response in unconfined areas, where short flow paths to surface boundaries allow rapid hydraulic adjustments. Consequently, the long-term response is controlled primarily by compressible storage in the deeper confined portions of the system. Under these conditions, aquitard vertical diffusivity plays a crucial role, as the presence
 640 of active surface boundaries enables the activation of vertical modes of equilibration. Aquitard layers therefore act as a coupling mechanism within the multilayer system, controlling the extent to which vertical diffusion influences the response of confined aquifers. To quantify this behaviour, we introduced a dimensionless parameter, r , which distinguishes between multilayer systems behaving as a set of weakly coupled layers ($r < 1$) from systems behaving as a coupled continuum for pressure propagation ($r \geq 1$). This criterion identifies whether one-dimensional formulations are sufficient to estimate the response
 645 time of individual layers or whether vertical diffusion provides significant hydraulic coupling between layers, in which case the two-dimensional approach is more suitable.

These described regimes can be viewed as end-members of a continuum. The response in unconfined areas is influenced by drainage density, landscape hydrology, vegetation uptake, and the strength of surface–groundwater interactions, among others.
 650 The transmission of these signals to deep parts of the basin depends on the inter-layer hydraulic coupling, primarily controlled by vertical diffusion. As a result, the regional adjustment of multilayer systems to large-scale stresses reflects a spectrum of behaviours, ranging from *storage-driven* to *boundary-driven* unconfined responses and their propagation in depth through vertical and horizontal diffusion. In this study, surface boundaries are addressed qualitatively through these two end-member



regimes, broadly representative of humid and arid conditions, respectively. A quantitative treatment requires further
 655 simulations and analysis to determine a relevant threshold differentiating the two regimes.

This distinction between unit-scale and basin-scale response times is relevant for many practical applications. Particularly, for
 sustainable groundwater management (Currell et al., 2016), the unit-scale response time of a specific aquifer is often of more
 interest, especially when the response times of aquitards are substantially longer than typical planning horizons, such that their
 660 delayed response may have little practical influence on the aquifer over the timescales relevant for groundwater management.
 Conversely, for applications concerned with the long-term propagation of perturbations through the geological system, such
 as nuclear waste disposal, the longest basin-scale response times are of greater relevance, particularly when they are associated
 with aquitard layers (Currell et al., 2016; Jost et al., 2007; Neuzil, 2015). The delineation of the most suitable analytical
 formulation to estimate *unit-scale response times* for each hydrogeological context is summarized in Figure 10.



665 **Figure 10. Delineation of hydrogeological contexts in which different analytical formulations are more suitable to estimate hydraulic response of confined aquifer units within multilayer systems.**

For instance, previous published estimations of response times of confined aquifers in the Aquitaine and Paris basins are in
 the order of 10^3 - 10^6 and 10^3 - 10^8 years, respectively (Ferguson et al., 2026; Rousseau-Gueutin et al., 2013). However, response
 670 times estimated from detailed mapping of flow-path lengths and hydrodynamic parameters in unconfined aquifers of these



basins indicate dominant response timescales ranging from months to decades (Cuthbert et al., 2019). Both basins also receive substantial annual recharge, in the order of 200 to 300 mm (Döll and Flörke, 2005; Margat and Gun, 2013; Richts et al., 2011). These characteristics may indicate a predominantly fast, *boundary-driven* response regime in unconfined systems with shallow water tables and well-developed surface water systems. Moreover, recent studies have shown that groundwater levels in
 675 underlying confined aquifers within these sedimentary basins are strongly influenced by annual to decadal hydroclimatic variations (Chidepudi et al., 2025a, b). Similarly, long-term numerical simulations of the Paris Basin suggest that most aquifer units reached new steady-states following climatic and geomorphological disturbances over the past five million years, with only deeper aquifers and very low-permeability units exhibiting localized disequilibrium (Jost, 2005; Jost et al., 2007). These findings suggest that the response times of confined aquifers in such systems may be substantially shorter than previously
 680 estimated, as they could be strongly influenced by vertical diffusion through aquitards and surface boundaries. An example calculation of response times for the Aquitaine Basin, using our proposed analytical framework for boundary driven systems, is presented in the Supplementary Material S3.

Conversely, systems such as the Nubian Aquifer System, the North-Western Sahara Aquifer System, the Great Artesian Basin,
 685 and the Arabian Aquifer System exhibit storage-driven unconfined response regimes, with response times exceeding 10^4 years in the unconfined areas (Cuthbert et al., 2019). These long timescales could reflect the persistence of fossil hydraulic gradients inherited from Late Pleistocene to Early Holocene climatic shifts (Bourdon, 1977; Lloyd and Farag, 1978; Schulz et al., 2017; Tierney et al., 2017). They are also consistent with current estimates of response times for confined and mixed aquifers in these basins (Ferguson et al., 2026; Rousseau-Gueutin et al., 2013).

690 There are ongoing questions as to whether hydraulic gradients observed in present-day arid regions reflect paleoclimatic remnants or are sustained by modern recharge. In our simulations, a zero-recharge condition is used as an end-member scenario representing a long-term climatic shift rather than the absence of present-day recharge. Under these conditions, the system ultimately evolves toward hydrostatic equilibrium over timescales exceeding 10 kyr. Considering that the last transition from
 695 humid to arid conditions is known to have occurred during the Late Pleistocene–Early Holocene (ca. 11–5 kyr BP) (Tierney et al., 2017), these simulated response times support the interpretation that observed hydraulic gradients may reflect the persistence of fossil signals (Schulz et al., 2017), superimposed on limited modern recharge. This is consistent with Rousseau-Gueutin et al. (2013), who showed that large arid-region groundwater basins may not reach hydraulic equilibrium between climatic transitions, as response times can exceed the timescales of the climate shifts. In this sense, the magnitude of
 700 hydrological change relative to given baseline conditions can influence system response when perturbations are large enough to modify the distribution of surface boundaries and dominant storage mechanisms at the regional scale over the long term.

We argue that some groundwater systems traditionally regarded as highly inertial can exhibit more rapid responses than previously assumed, particularly regional confined and mixed aquifers within basins typical of humid settings where inter-



705 layer hydraulic coupling is strong. Confined and mixed aquifers may respond much faster as vertical diffusion through
aquitards enhances the propagation of pressure changes between deep units and surface boundaries. Although vertical
diffusivities in aquitards are typically several orders of magnitude lower than horizontal diffusivities in aquifers, the distance
over which pressure propagates vertically can be much shorter than the horizontal extent of a regional groundwater basin. In
any case, shorter response times do not imply negligible transient dynamics. In fact, these systems may be more sensitive to
710 projected climate change and groundwater abstraction within relevant human and management timescales. In this sense,
hydraulic response times remain relevant for vulnerability, renewability, and sustainability assessments (Currell et al., 2016;
Cuthbert et al., 2023; Gleeson et al., 2012; Pierce et al., 2013).

Finally, both the numerical and analytical approaches used in this study assume intra-layer homogeneity. The effect of
715 heterogeneities and arising connectivity mechanisms, such as preferential flow paths, on the propagation of pressure changes
deserves further study, as well as determining the appropriate upscaling approach of hydraulic diffusivities for the estimation
of response times using simplified analytical solutions (Knudby and Carrera, 2006; Renard and Allard, 2013; Sanchez-Vila et
al., 2006; Trinchero et al., 2008). The use of simple closed-form analytical formulations to estimate response times is relevant
for obtaining first-order estimates, especially when data and characterization are sparse, which is common in many systems.
720 These estimates can inform further monitoring or modelling decisions (Haitjema, 2006), and help understanding the broad
behaviour and dynamics of the system. Detailed assessments on heterogeneities, complex boundary conditions, structures, and
uncertainties in the dynamics of a system would demand detailed modelling approaches supported with extensive observational
datasets.

6 Conclusions

725 This study shows that response times of regional multilayer groundwater systems are strongly controlled by the
hydrogeological context governing the propagation of pressure changes. Two contrasting regimes emerge at the basin scale.
In arid systems characterized by deep water tables and poorly developed surface drainage networks, the response of unconfined
areas is predominantly *storage-driven*, with hydraulic perturbations propagating slowly toward distant lateral boundaries.
Under these conditions, the long-term response is controlled primarily by unconfined storage, and the mixed-aquifer
730 formulation of Rousseau-Gueutin et al. (2013) provides good order-of-magnitude estimates. Its applicability is largely
insensitive to aquitard vertical diffusivity because the limited extent of active surface boundaries restricts vertical modes of
equilibration.

In contrast, humid systems characterized by shallow water tables and well-developed surface drainage networks exhibit a
735 predominantly fast *boundary-driven* response in unconfined areas, due to the interaction with surface boundaries. In these
settings, aquitard vertical diffusivity becomes a key control because vertical diffusion can couple hydrogeological units and



activate vertical modes of equilibration. Consequently, the one-dimensional confined-flow formulation (Domenico and Schwartz, 1998; Rousseau-Gueutin et al., 2013) may substantially overestimate response times when vertical hydraulic coupling is significant.

740

To identify the conditions under which coupling becomes important, we introduced the dimensionless parameter, r , which compares horizontal and vertical time constants. For $r < 1$, the multilayer system behaves as a set of weakly coupled layers, and one-dimensional formulations remain appropriate for estimating the response time of individual aquifers. For $r \geq 1$, the system behaves as a coupled continuum in which response times emerge as basin-scale properties. For this, we developed a two-dimensional analytical formulation representing the multilayer system as an equivalent homogeneous anisotropic medium. The dominant response time can be expressed as half the harmonic mean of the equivalent horizontal and vertical response times, being controlled by the direction of fastest equilibration. Numerical simulations covering a range of hydraulic properties, numbers of layers, basin extensions, and layer thicknesses show good agreement between analytical and simulated response times, supporting the formulation as a robust estimator of hydraulic response times in this context.

745

750 7 Appendix A – Derivation of the analytical formulation

The hydraulic heads, $h(x, z, t)$, of the two-dimensional, homogeneous, anisotropic flow system presented in Sect. 3.1 are governed by Eq. (A1), with the initial and boundary conditions summarized in Eq. (A2) to Eq. (A6).

$$\frac{\partial h}{\partial t} = D_x \frac{\partial^2 h}{\partial x^2} + D_z \frac{\partial^2 h}{\partial z^2}, \quad 0 < x < L, 0 < z < B, \quad t > 0 \quad (\text{A1})$$

755

$$h(x, z, 0) = h_0(x, z) \quad (\text{A2})$$

$$h(0, z, t) = \begin{cases} h_0(0, z), & t = 0 \\ h_f(0, z), & t > 0 \end{cases} \quad (\text{A3})$$

$$h(L, z, t) = \begin{cases} h_0(L, z), & t = 0 \\ h_f(L, z), & t > 0 \end{cases} \quad (\text{A4})$$

$$h(x, B, t) = \begin{cases} h_0(x, B), & t = 0 \\ h_f(x), & t > 0 \end{cases} \quad (\text{A5})$$

$$\frac{\partial h}{\partial z}(x, 0, t) = 0 \quad (\text{A6})$$

760

The solution can be decomposed into a final steady state solution, $h_\infty(x, z)$, that satisfies the final boundary conditions, and a transient perturbation $r(x, z, t)$ defined for $t > 0$, representing the residual diffusion, as given in Eq. (A7). Solving for the residual diffusion we obtain the second-order linear partial differential equation (PDE) given in Eq. (A8). The initial conditions can be rewritten as given in Eq. (A9), representing the difference between initial and final equilibrium states. For $t > 0$, the



765 boundary conditions become homogeneous since both $h_{\infty}(x, z)$ and $h(x, z, t)$ satisfy the final boundary conditions after the step change, as presented in Eq. (A10) to Eq. (A13).

$$h(x, z, t) = h_{\infty}(x, z) + r(x, z, t) \quad (\text{A7})$$

$$\frac{\partial r}{\partial t} = D_x \frac{\partial^2 r}{\partial x^2} + D_z \frac{\partial^2 r}{\partial z^2}, \quad 0 < x < L, 0 < z < B, \quad t > 0 \quad (\text{A8})$$

770 $r(x, z, 0) = h_0(x, z) - h_{\infty}(x, z) \quad (\text{A9})$

$$r(0, z, t) = 0 \quad (\text{A10})$$

$$r(L, z, t) = 0 \quad (\text{A11})$$

$$r(x, B, t) = 0 \quad (\text{A12})$$

$$\frac{\partial r}{\partial z}(x, 0, t) = 0 \quad (\text{A13})$$

775

We can apply eigenfunction expansion via separation of variables, assuming a solution of the form given in Eq. (A14). Substituting the derivatives in Eq. (A8) and introducing the separation constants $-\lambda$, ($\lambda > 0$) and $-\mu$, ($\mu > 0$), we obtain three ordinary differential equations (ODE), given in Eq. (A15).

780 $r(x, z, t) = X(x)Z(z)T(t) \quad (\text{A14})$

$$\frac{1}{X(x)} \frac{d^2 X}{dx^2} = -\lambda, \quad \frac{1}{Z(z)} \frac{d^2 Z}{dz^2} = -\mu, \quad \frac{1}{T(t)} \frac{dT}{dt} = -(D_x \lambda + D_z \mu) \quad (\text{A15})$$

Solving the spatial ODEs and applying boundary conditions, the respective eigenvalues and eigenfunctions can be found, as given in Eq. (A16) and (A17). The temporal ODE has a general solution of the form given in Eq. (A18), where C is an integration constant depending on the initial conditions.

785

$$X_n(x) = \sin\left(\frac{n\pi}{L}x\right), \quad \lambda_n = \left(\frac{n\pi}{L}\right)^2, \quad n = 1, 2, 3, \dots \quad (\text{A16})$$

$$Z_m(z) = \cos\left(\frac{(2m-1)\pi}{2B}z\right), \quad \mu_m = \left(\frac{(2m-1)\pi}{2B}\right)^2, \quad m = 1, 2, 3, \dots \quad (\text{A17})$$

$$T_{n,m}(t) = C e^{-(D_x \lambda_n + D_z \mu_m)t} \quad (\text{A18})$$

790

Each product $r_{mn}(x, z, t) = X_n(x) Z_m(z) T_{mn}(t)$ satisfies the PDE and the boundary conditions for the residual diffusion, and the general solution can be expressed as their superposition, given in Eq. (A19), where $C_{m,n}$ are coefficients that can be determined from the initial conditions via double Fourier series expansion.

795 $h(x, z, t) = h_{\infty}(x, z) + \sum_{m=1}^{\infty} \sum_{n=1}^{\infty} C_{m,n} \sin\left(\frac{n\pi}{L}x\right) \cos\left(\frac{(2m-1)\pi}{2B}z\right) e^{-(D_x \lambda_n + D_z \mu_m)t} \quad n, m = 1, 2, 3 \dots \quad (\text{A19})$



From this, the time constants of the system can be expressed as in Eq. (A20), where $\tau_{h,n}$ and $\tau_{v,m}$ represent the equivalent time constants of the horizontal and vertical modes, respectively. The dominant time constants controlling long-term response correspond to the fundamental modes with $n = m = 1$.

800

$$\tau_{n,m} = \frac{1}{\frac{1}{\tau_{h,n}} + \frac{1}{\tau_{v,m}}} \quad (\text{A20})$$

$$\tau_{h,n} = \frac{1}{(n\pi)^2} \frac{L^2}{D_x} \quad (\text{A21})$$

$$\tau_{v,m} = \frac{4}{((2m-1)\pi)^2} \frac{B^2}{D_z} \quad (\text{A22})$$

805 This analytical formulation is applied to a multilayer system by assuming equivalent diffusivities in the vertical and horizontal directions through layer averaging. The equivalent vertical diffusivity can be computed as the thickness-weighted harmonic mean (Hickson et al., 2009a, b), and the equivalent horizontal diffusivity as the thickness-weighted arithmetic mean of the k individual layers, as given in Eq. (A23) and (A24).

$$810 \quad D_x = D_{h,eq} = \frac{1}{B} \sum_1^k D_{h,i} b_i \quad (\text{A23})$$

$$D_z = D_{v,eq} = \frac{B}{\sum_1^k \frac{b_i}{D_{v,i}}} \quad (\text{A24})$$

The hydraulic diffusivity for a confined unit will be given by the ratio of the horizontal or vertical hydraulic conductivity to specific storage, as given by Eq. (A25) and (A26). For unconfined aquifers, horizontal hydraulic diffusivity is typically derived from transmissivity and specific yield. Since transmissivity in this setting depends on the saturated thickness, it is generally approximated as the product of hydraulic conductivity and a representative thickness, b . When head fluctuations are small relative to the total saturated thickness, b can simply be taken as the average aquifer thickness. Alternatively, b can be defined through a steady-state approximation that accounts for recharge, conductivity, and aquifer geometry, as proposed by Erskine and Papaioannou (1997). Vertical diffusivity is governed by the vertical hydraulic conductivity and specific storage, as given in Eq. (A27) and (A28).

820

$$D_{h,conf} = \frac{k_h}{S_s} \quad (\text{A25})$$

$$D_{v,conf} = \frac{k_v}{S_s} \quad (\text{A26})$$

$$D_{h,unc} \approx \frac{k_h b}{S_y} \quad (\text{A27})$$



$$825 \quad D_{v,unc} = \frac{k_v}{S_s} \quad (A28)$$

8 Code availability

The code containing Python scripts of model setup files as well as pre- and postprocessing tools are made available through the GitHub repositories: <https://github.com/femarivera/ConfinedLab/tree/v0.1.1> and <https://github.com/femarivera/ConfinedLab-mlibs/tree/v0.1.1> (Marin-Rivera, 2026b, a).

830 9 Supplementary information

The supplement is available online.

10 Author contributions

CFMR, AP, JG contributed to the conceptualization of the work. CFMR developed, runed, and postprocessed models. CFMR, AP, JG, GF performed analysis of the results. CFMR drafted first manuscript. AP, JG, GF reviewed and edited final
 835 manuscript.

11 Competing interests

The authors declare that they have no conflict of interest.

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