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ISSN: 0976-3376

ASIAN JOURNAL OF SCIENCE AND TECHNOLOGY

Asian Journal of Science and Technology
Vol. 17, Issue, 04, pp. 14218-14224, April, 2026

RESEARCH ARTICLE

NUMERICAL RESOLUTION OF DIFFUSIVITY EQUATION IN SATURATED POROUS MEDIA

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ARTICLE INFO

Article History:
Received 17th January, 2026
Received in revised form
26th February, 2026
Accepted 02nd March, 2026
Published online 30th April, 2026

Key words:
Aquifer, pumping test, Hydraulic head, Transmissivity, Storage coefficient, Diffusivity.

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ABSTRACT

Mobilizable fresh water is a scarce and precious resource, generally stored in groundwater aquifers. Therefore, for effective management, it is essential to understand and simulate groundwater flow in saturated porous media. In our work, we presented the general principles of groundwater hydraulics and established the diffusivity equation. A numerical solution based on discretization using the finite difference method was developed. The results of the numerical simulation of a pumping test, performed using this mathematical model and represented by curves and tables showing the evolution of the head at a given distance from the well, were considered equivalent to those of an experimental setup. This allowed us to verify, in steady state, the radius of influence of the pumping test in the porous medium and, in transient state, its hydrodynamic parameters. The results obtained, compared with those of the numerical model, showed satisfactory agreement. This work opens up interesting perspectives for the modeling of aquifer systems and the management of groundwater resources.

Citation: ***Fallou SARR, Vieux Boukhaly TRAORE, Omar Ngor THIAM and Mamadou Lamine SOW.* 2026.** "Numerical Resolution of Diffusivity Equation in Saturated Porous Media", *Asian Journal of Science and Technology*, 17, (04), 14218-14224.

INTRODUCTION

Water is the source of all life and covers more than 70% of the Earth's surface. It is estimated that the oceans (non-potable waters) contain approximately 95 to 98% of this water, with the remainder distributed among other major reservoirs and constituting the world's only freshwater reserve. Of this, more than half is stored in glaciers and remains difficult to access for drinking water needs. Lakes and rivers, which represent less than 1% of available freshwater, generally require treatment to be made potable, while groundwater is more abundant (approximately 30%) and often directly drinkable. This groundwater resource, which is difficult to observe, requires an understanding of the geological environments through which the water flows. Hydrogeology thus provides the basis for modeling groundwater flow, which is essential for effective water resource management. It is defined by two different but complementary approaches [1], [2], [3], [4], [5], [6], [7], [8]: (i) a specialization within geology that deals with the properties of geological environments, particularly their capacity to contain and allow groundwater to flow. Hydrogeology is therefore, in this case, a discipline focused on geological environments, which are considered the container of groundwater [9]; (ii) a specialization within hydrology that deals with groundwater while taking geological conditions into account. It is therefore, in this case, a discipline focused on the behavior of groundwater, which is considered the CONTENT of aquifer geological environments [10], [11]. The hydrogeologist must first explain what is happening in the subsoil, and then deduce an estimate of one or more quantities [12]. In terms of forecasting, the quantity of water is one of the main problems that hydrogeologists must be aware of [13], [14], [15]. However, to present a quantitative approach to groundwater flow, it is necessary to understand the hydrological conditions [16], [17], [18]. Whatever the ultimate goal, the best way to make the most of all available data relating to an aquifer is to combine this data with the appropriate physical laws (expressed as equations) to form a mathematical model based here on solving the diffusivity equation [9], [19], [20]. Therefore, for the development of this model, it is necessary to retain the essentials of hydrogeology and groundwater hydraulics, adopting assumptions and approximations that will allow us to simplify real-world phenomena that are complex by nature. This work is organized into two phases, one theoretical and the other practical. The objective of this study is, firstly, to review the general concepts of groundwater hydrodynamics and, secondly, to use numerical results to determine the hydrodynamic parameters of a groundwater aquifer.

METHODS AND MATERIALS

Problem position: A porous medium is a material composed of a solid matrix and voids, called pores. These voids are often filled with water, air, or other fluids. In our study, we are interested in water-saturated media, that is, media in which all the voids are filled with water. In hydrogeology, the analysis of porosity is not exclusively geometric but also refers to the water contained within the porous medium, its physical bonds with the solid matrix, and its potential movements. Indeed, the total porosity of a rock is not a sufficient condition for water flow. It is the interconnection of the pores, as well as the fluid-solid interactions, that define the fluid circulation. In this study, we focus on aquifer porous media. These are permeable geological formations whose pores or fissures communicate and are large enough for water to circulate under the influence of gravity. An aquifer thus constitutes a reservoir of groundwater. Three types of aquifers are distinguished: unconfined aquifers, confined aquifers, and semi-confined aquifers. We will choose a confined aquifer. It can be defined as a groundwater table trapped within a permeable geological formation, between two impermeable formations (Figure 1).

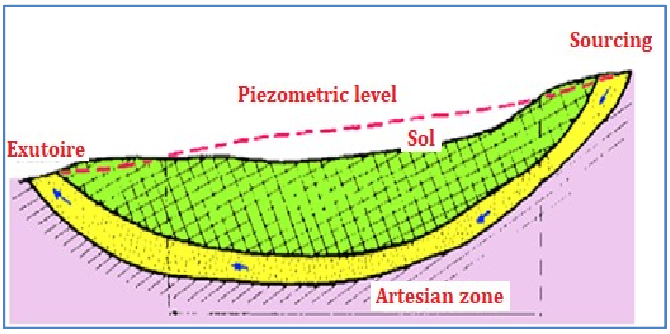

Figure 1. Confined surface aquifer

Here, the water table is under pressure, and when a well is drilled, the water rises to a higher equilibrium level. Sometimes, the water table even erupts from the ground; this is the phenomenon of artesian aquifers. The formation shown is a saturated aquifer throughout its thickness; it is bounded above by an impermeable (clay) or semi-permeable layer. The piezometric level, different from the water table level and always above the base of the upper impermeable layer, is theoretical until a borehole or piezometer reaches the aquifer through its top. Such a borehole is called an artesian well, and if the water rises to the surface (piezometer level above ground level), it is called a gushing artesian well. It flows naturally without pumping. Thus, the total energy of the aquifer is expressed by its hydraulic head. It is the sum of the water height, pressure, and velocity. At point M, the hydraulic head is defined by the equation (1):

$$h_M = \frac{U^2{}_M}{2g} + \frac{P_M}{\rho g} + Z_M \tag{1}$$

The flow velocities U_M in soils are such that $U_M < 10\ cm.s^{-1}$. Therefore $\frac{U^2{}_M}{2g} < 0{,}5mm$.

The value of the velocity load is then negligible compared to that of the other loads $\left(\frac{U^2{}_M}{2g} \ll \frac{P_M}{\rho g} + Z_M\right)$.

This leads us to equation (2):

$$h_M = \frac{P_M}{\rho g} + Z_M \tag{2}$$

The hydraulic gradient represents the difference in piezometric levels between two pointsis given by relation (3):

$$\vec{\imath} = -\overrightarrow{grad}h = \begin{cases} -\frac{\partial h}{\partial x} \\ -\frac{\partial h}{\partial y} \\ -\frac{\partial h}{\partial z} \end{cases} \tag{3}$$

$\vec{\imath}$ indicates the direction and intensity of the flow (water flows from areas of higher pressure to areas of lower pressure). If $\vec{\imath}$ is constant, the flow is said to be uniform (a very common assumption in groundwater flow). In the case of a uniform flow, $\vec{\imath}$ is constant and its value is given by the equation (4).This is illustrated by the (Figure 2)

$$\vec{\imath} = \frac{\Delta h}{\Delta l} = \frac{h_A - h_B}{L} \tag{4}$$

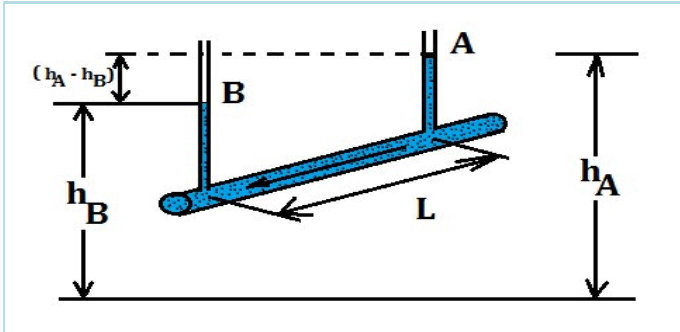

Figure 2. Hydraulic gradient of a uniform flow

Hydraulic permeability, or conductivity, is the ability of a reservoir to allow water to pass through it under the influence of a hydraulic gradient. It expresses the resistance of the medium to the flow of water passing through it. It can be measured by two parameters: (i) the permeability coefficient K, which represents the speed at which water (or a fluid with a viscosity of 1 centipoise) passes through a unit cross-section perpendicular to the direction of flow in a medium under a unit hydraulic gradient at 20°C; (ii) the intrinsic permeability coefficient ki, which is the parameter that reflects the ability of the solid skeleton to transfer a reference fluid. Hydraulic conductivity is therefore a characteristic of the matrix and the fluid, but not an inherent and intrinsic property of the reservoir. These two parameters are related by the following equation (5):

$$K = k_i \frac{\rho g}{\mu} \tag{5}$$

Darcy's law is given by equation (6):

$$\vec{U} = K\vec{\imath} = -K\overrightarrow{grad}h = -\frac{k_i}{\mu}\left(\overrightarrow{grad}p + \rho g\overrightarrow{grad}z\right) \tag{6}$$

The Darcy permeameter experiment is performed by observing unidirectional flow. The intrinsic permeability coefficient is an isotropic property of the porous medium, independent of spatial direction. In many cases, the properties of the medium are not isotropic: this is clearly the case, at least, between the permeability in the x, y direction and that in the z direction of stratified clay layers. Darcy's law can therefore be written as relation (7).

$$\vec{U} = \overline{\overline{K}}\vec{\imath} = -\overline{\overline{K}}\overrightarrow{grad}h \tag{7}$$

The permeability coefficient K is defined by the relation (8) :

$$\overline{\overline{K}} = \begin{bmatrix} K_{xx} & 0 & 0 \\ 0 & K_{yy} & 0 \\ 0 & 0 & K_{zz} \end{bmatrix} \quad (8)$$

The apparent speed can then be expressed along the x, y and z directions according to the relation (9):

$$\begin{cases} U_x = -K_{xx}\frac{\partial h}{\partial x} \\ U_y = -K_{yy}\frac{\partial h}{\partial y} \\ U_z = -K_{zz}\frac{\partial h}{\partial z} \end{cases} \quad (9)$$

The flow rate Q is then given by the relation (10):

$$Q = U.A = k.A.i \quad (10)$$

Darcy's law is valid under four conditions: continuity, isotropy, and homogeneity of the reservoir, and laminar flow. Laminar flow is maintained when the Reynolds number satisfies the relation (11):

$$R_e = \frac{\vartheta d}{\mu\rho} < 2000 \quad (11)$$

A physical property (e.g., porosity) of an aquifer is homogeneous when its spatial distribution is uniform. In other words, the value is the same throughout all the reverse enrichment zones. The anisotropic nature of a transfer property (e.g., permeability) is obtained when its value depends on the direction considered in space. Aquifers are generally anisotropic and heterogeneous, but in practice, this anisotropy and heterogeneity are often neglected. We will therefore consider the aquifer as isotropic. Furthermore, since we always use large volumes of soil and the characteristics are only the averages of point values within it, the heterogeneities cancel each other out and are greatly reduced. The result, as a whole, can therefore be applied to a homogeneous aquifer. We will thus consider an aquifer as a homogeneous and isotropic medium. This assumption will be accepted throughout our study. Transmissivity is the measure of the amount of water that can be transmitted horizontally through the total saturated thickness of the rock per unit width under the effect of a hydraulic gradient equal to one. It is determined numerically by the product of the layer's permeability and its thickness given by equation (12).

$$T = K\,.\,e \quad (12)$$

It can be obtained (THEIS method) using formula (13):

$$T = \frac{Q}{4\pi} * \frac{W(u)}{s} \quad (13)$$

And the error in transmissivity is given by the relation (14):

$$\Delta T = T\left(\frac{\Delta V}{V} + \frac{\Delta t}{t} + \frac{\Delta W(u)}{W(u)} + \frac{\Delta s}{s}\right) \quad (14)$$

Porosity is a static characteristic of soils; it is independent of the movement of water that may be present within them. For a given sample of a formation, whether an aquifer or not, porosity is expressed as the ratio between the volume of voids (or pores) and the total volume of the sample given by equation (15):

$$n = \frac{V_v}{V_t} \quad (15)$$

Natural soil is composed of grains of varying sizes. To measure this porosity, the volume of voids must be measured, which is equivalent to estimating the water volume of an aquifer. Two categories of water are distinguished: (i) gravity water: this water obeys the laws of gravity, can be extracted from the ground, circulates in aquifers, and feeds wells and springs; (ii) capillary water: this water cannot be mobilized except by steaming and remains bound to the soil grains by surface tension, molecular adhesion, and adsorption; it therefore does not participate in groundwater circulation. However, a reservoir is never completely devoid of capillary water. In hydrogeology, effective porosity is preferred to the more theoretical total porosity. Effective porosity is thus expressed as equation (16):

$$n_e = \frac{V_e}{V_t} \quad (16)$$

It is expressed as a percentage of the volume of water flowing by gravity. The actual velocity V_e of water flow through the soil pores is thus defined as a function of Ugiven by equation (17):

$$V_r = \frac{U}{n} \quad (17)$$

A drained site provides effective porosity (gravity water). Retention water provides residual porosity (capillary water). The sum of these two porosities gives the total porosity. The storage coefficient, denoted S, can be physically defined as the quantity of water released (gravity water) from a vertical prism with a base of 1 m² and the height of the aquifer under a unit change in hydraulic head (Δh=1). The specific storage coefficient, Ss, corresponds to the proportion of water recovered, relative to a unit height of the saturated formation. S is expressed as a percentage and is measured by pumping tests expressed by the equation (18).

$$S = e\,.\,S_s \quad (18)$$

The storage coefficient is measured in the field using pumping tests. The storage coefficient can be obtained using theformula (19):

$$S = \frac{4\,T\,t\,u}{r^2} \quad (19)$$

Its error ΔS is given by the relation (20):

$$\Delta S = S\left(\frac{\Delta T}{T} + \frac{\Delta u}{u} + 2 * \frac{\Delta r}{r} + \frac{\Delta t}{t}\right) \quad (20)$$

Diffusivity governs the propagation of influence within the aquifer. It is denoted by αn and is obtained numerically bythe relation (21):

$$\alpha_n = \frac{T}{S} \quad (21)$$

Mathematical formulation: The mathematical formulation is based on Partial Differential Equations and the concepts of discretization and meshing:

Partial differential equations: The partial differential equations are divided into a continuity equation (relation 22), Darcy's law assuming negligible water compressibility (relation 23), otherwise (relation 24) and the isothermal equation of state of water (relation 25).

$$div(\rho\vec{U}) + \frac{\partial}{\partial t}(\rho n) + \rho q = 0 \quad (22)$$

$$\vec{U} = -\overline{\overline{K}}\vec{\nabla}h \quad (23)$$

$$\vec{U} = -\frac{\overline{\overline{k_i}}}{\mu}(\vec{\nabla}p + \rho g\vec{\nabla}z) \quad (24)$$

$$\rho = \rho_0 \exp(\beta(p - p_0)) \quad (25)$$

For a confined aquifer, the diffusivity equation is written according to the relation (26):

$$\Delta h = \frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} = \frac{q}{T} + \frac{S}{T}\frac{\partial h}{\partial t} \quad (26)$$

Concepts of discretization and meshing: The term mesh is often used to refer to the arrangement of discretization points in space. A mesh can be fixed or moving over time in space, but we will only consider fixed meshes. Our equation has more than one spatial direction, and there are several options for discretizing it. We will only use structured meshes (discretization is done in rows and columns). In two spatial dimensions, a computation point is identified by two spatial indices i and j and a temporal index n (Figure 3).

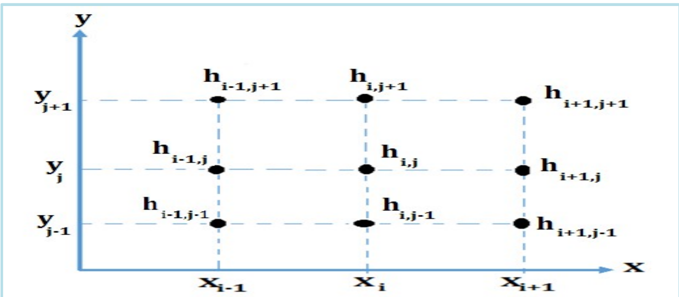

Figure 3. Discretization of two spatial dimensions

We can therefore note that for a discretization of a domain D, the load will be calculated at a finite number of points. The chosen mesh is a grid with steps Δx and Δy. The interior of the domain is numbered from i=1 to i_max and from j=1 to j_max. Each point is identified by a pair of indices (i, j) where the first index denotes the row number and the second the column number.

Boundary and initial conditions: The aquifer under consideration has an initial hydraulic head of 50 meters. The values of its hydrodynamic parameters are: (i) the transmissivity T=10-1 m².s-1; (ii) the storage coefficient S=10-2; For the spatial and temporal discretization, we will take: (i) the spatial step along the x-axis is Δx=100 meters; (ii) The spatial step along the y-axis is Δy = 100 meters. (iii) The time step is Δt = 10 seconds. The boundary conditions of the domain are of the Dirichlet and Neumann type. The radius of influence R (or action) due to pumping is such that:

200 m < R < 300 m. The pumping time is t = 1000 s, and we will record hydraulic heads from the medium for 1700 s, starting from the beginning of pumping. The pumping well is at node (4,4), i.e., the central node.

Numerical formulation

Resolution methods: In transient regimes, solving the diffusivity equation requires distinguishing between the temporal and spatial aspects of the problem. For the numerical solution of the diffusivity equation in 2D, we used the finite difference method. Using the explicit scheme, we derived the expression for the hydraulic head at time n+1 (unknown) based on expressions for heads at time n (already known).

Resolution of equation: We considered a square mesh, i.e., Δx=Δy. This condition applied in the diffusivity equation leads to relation (27):

$$h_{i,j}^{n+1} = \left(1 - 4\frac{T}{S}\frac{\Delta t}{\Delta x^2}\right)h_{i,j}^n + \frac{T}{S}\frac{\Delta t}{\Delta x^2}\left(h_{i-1,j}^n + h_{i+1,j}^n + h_{i,j-1}^n + h_{i,j+1}^n\right) \quad (27)$$

Elaborate calculation code: The circulation code comprises : (i) an initial phase for reading the data (mesh, time step, hydrodynamic parameters of the aquifer, verification of the stability condition, etc.); (ii) three calculation loops within which a calculation module is introduced to determine the different values of h_(i,j)^n. The first loop represents time, and the other two are space loops. The time and space steps will be chosen to respect the stability criterion of the explicit scheme given by the relation(28) [19]:

$$\frac{T}{S}\Delta t\left\{\frac{1}{\Delta x^2} + \frac{1}{\Delta y^2}\right\} \leq \frac{1}{2} \quad (28)$$

Resolution tool: The flowchart below describes the calculation code developed in Fortran for the numerical solution of the two-dimensional flow problem in transient regime (Figure 4).

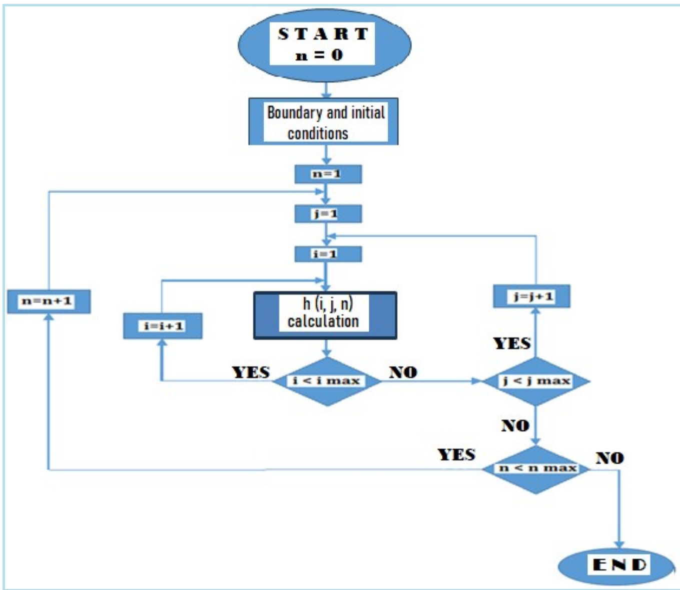

Figure 4. Flowchart of the digital code

RESULTS AND DISCUSSION

The Figure 5 illustrates the 3D representation of the aquifer during and after the pumping test through the hydraulic heads h(4, j) where j = {2, 3, 4}. The variation of the heads during and after a pumping test at constant flow rate is represented in tabular (Table) and graphical (Figure 6). Regarding the evolution of the water level in the porous medium, we observe in detail various phenomena which, given their effects on the water level behavior in the well, can be grouped into three categories: (i) the phase of lowering of piezometric levels in the aquifer induced by pumping. Water extraction from the aquifer modifies the transmissivity at the well and, consequently, the hydraulic head. This disturbs the flow near the well because the hydraulic gradient is neither zero nor constant. The water then flows in the aquifer towards areas of lower pressure, i.e., towards the well; (ii) a phase of pressure stabilization after a prolonged pumping test: this is the steady state.

values close to their initial values. Consequently, the hydraulic head in the aquifer stabilizes.

Validation: Figures 7 and 8 respectively represent the drawdown curve of the charges as a function of their distance from the well and the superposition of the THEIS characteristic curve and the experimental curve.This section aims to determine the radius of influence (steady state), the transmissivity, and the storage coefficient (transient state). We will use the Dupuit and Theis methods, respectively, for steady state and transient state.

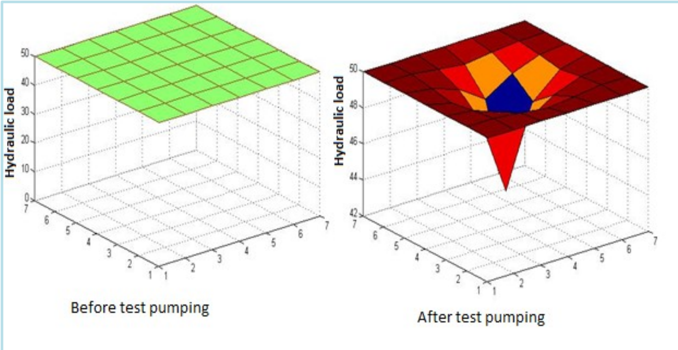

Figure 5. 3D representation of the aquifer

Table 1. Variation of loads h (4, j) during and after a test pumping of flow rate Q = 2m^{3}·s^{-1}

Q en L.s^{-1}	2000						0				
t en s	0	10	20	30	630	1000	1010	1020	1200	1500	1700
h(4,4)	50	48	46,8	46	41,99	41,95	43,81	44,93	48,82	49,77	49,92
h(4,3)	50	50	49,8	49,56	46,65	46,61	46,61	46,79	48,95	49,8	49,93
h(4,2)	50	50	50	49 ,98	48,75	48,73	48.73	48,73	49,71	49,88	49,96

Table 2. Variation of loads h (4, j) during and after a test pumping of flow rate Q = 1m^{3}.s^{-1}

Q en l/s	1000						0				
t en s	0	10	20	30	630	1000	1010	1020	1200	1500	1700
h(4,4)	50	49	48,4	48	48,86	45,83	46,8	47,37	49,38	49,88	49,96
h(4,3)	50	50	49,9	49,78	48,27	48,24	48,24	48,34	49,47	49,89	49,96
h(4,2)	50	50	50	49,99	49,35	49,34	49,34	49,34	49,69	49,94	49,98

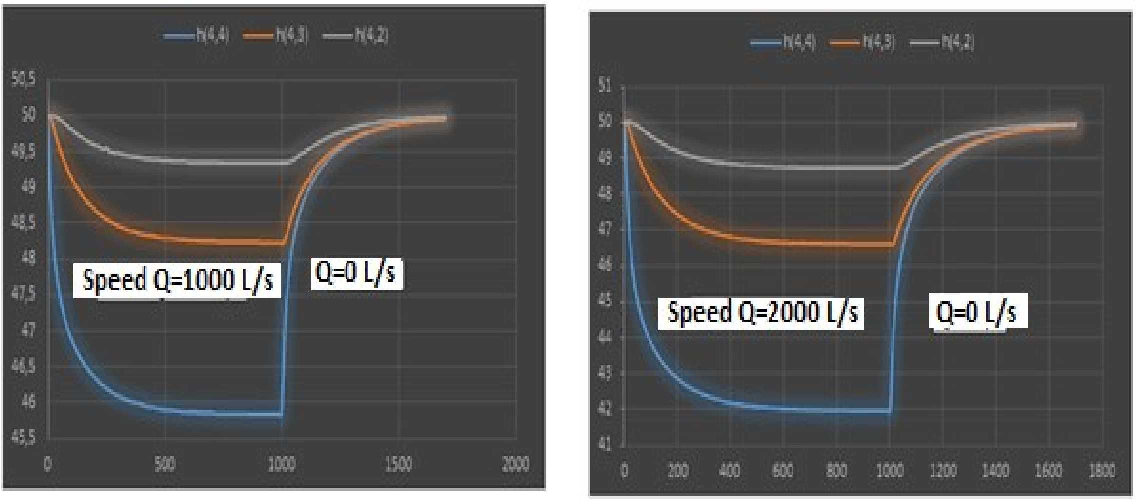

Figure 6. Evolution curves of water levels h (4, j) according to pumping flow rate

The amount of water extracted is equal to the amount of water supplied by the flows near the well. The hydraulic gradient (around the well) becomes constant but not zero in the area of influence. (iii) When pumping ceases, the mass of water supplied to the well by the flows causes a rise in the water level in the aquifer. This rise reduces the hydraulic gradient of the aquifer. The hydrodynamic parameters at the well return to

The first method allowed us to graphically determine the value of the radius of influence, which is equal to R = 280 m. The uncertainty in this radius was calculated and is equal to ΔR = ±20 m. The theoretical value of the radius of influence will therefore be presented as follows: R = (280 ± 20) m. In reality, the drawdown curve as a function of the distance to the well is not a straight line but has a conical shape. The value found is therefore approximate. With the second method, we observed that the point of intersection of the two curves has the following coordinates: (i) on the THEIS curve: W(u) = 1.1 and 1/u = 4.6; (ii) On the experimental curve: s = 1.7 and t = 100 s. The pumping rate is 2 m^3/s and the distance between the measurement node and the pumping well is 100 m. The transmissivity calculation gives T = 0.1 m^2/s and the error in the transmissivity is $\Delta T = \pm 0.02$ m^2/s. Therefore, the transmissivity is given by the following value: T = (0.1 ± 0.02) m^2/s. The storage coefficient is S = 0.01 and its uncertainty is $\Delta S = \pm 0.003$. Therefore, the storage coefficient is given by the following value: S = (1 ± 0.3) %

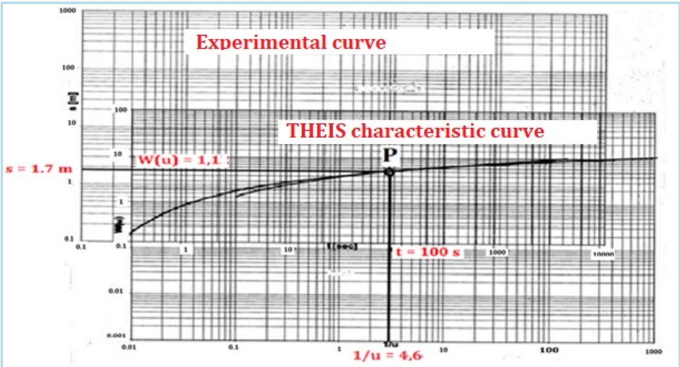

Figure 7. Drawdown curve of the loads as a function of their distance from the well

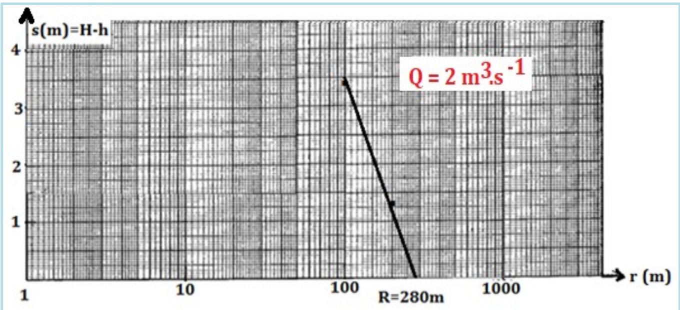

Figure 8. Superposition of the THEIS characteristic curve and the experimental curve

CONCLUSION

In this work, we reviewed hydrogeology and groundwater hydraulics, specifically the storage and transfer properties of a porous medium. We also discussed several methods for interpreting pumping tests used to determine the hydrodynamic parameters of an aquifer. Following the theoretical study, a numerical code, based on the discretization of the diffusivity equation using the finite difference method, was developed. The scheme used for this code is self-explanatory. A numerical simulation with hydrodynamic parameters of an arbitrary aquifer allowed us to plot the evolution of several hydraulic heads in the porous medium under consideration. We then considered these results as experimental to try to determine: (i) the radius of influence due to the pumping test using the Dupuit method in steady state. The result obtained for the radius of influence, and its approximate accuracy, determined by the Dupuit method in steady state, is satisfactory, yielding a value between 200 and 300 meters, as indicated by the numerical data; (ii) the transmissivity and storage coefficient of the porous medium in transient state using the Theis method. The values obtained for the transmissivity and storage coefficient, as well as their accuracy, are very close to the values used in the numerical simulation. The results obtained are satisfactory because they are consistent, within the same order of magnitude, with the values used in the numerical simulation. In addition to developing a numerical code to simulate flows in a porous medium, mastering the methods for determining the hydrodynamic parameters of a porous medium provides a foundation for modeling groundwater flow.

Acknowledgements: We would like to thank Professor Sambou for his valuable collaboration in collecting the parameters and information necessary for this study.

Conflict of interest: The authors declared that there are no conflicts of interest related to this manuscript

Source of Funding: No funding was received for this work. Everything was done on a modest personal funding.

Authors' Contribution: This work was carried out in close collaboration with all authors. Authors FS and VBT designed the study, defined the analysis protocol, and wrote the manuscript. Authors ONT and MLS reviewed and validated the document.

Nomenclature

Latin	Grec
A, Section traversée par l'écoulement, (m^2)	α, Specific compressibility coefficient of the porous medium, ($kg^{-1}.m.s^2$)
Cv, Coefficient de consolidation, (-)	αn, Diffusivity, ($m^2.s^{-1}$)
e, Epaisseur d'une nappe, (m)	β, Compressibility of water, (m^{-1})
g, Accélération de la pesanteur, (m^2.s-1)	ρ, Dynamic viscosity of the liquid, ($kg.m^{-1}.s^{-1}$)
h, Hauteur piézométrique, (m)	ρ, Density, ($kg.m^{-3}$)
$\vec{i}$, Vecteur gradient hydraulique, (-)	τ, Elevation of the bedrock of a free aquifer, (m)
K, Perméabilité moyenne des éléments de l'aquifère, (m.s-1)	
ki, Perméabilité induite, $(m)^2$	
$\bar{\bar{K}}$,$\bar{\bar{k}}$, Tenseurs de perméabilité, (--)	
n, Total porosity of a porous medium, (%)	
ne, Effective porosity, (%)	
P, Pressure, (Pa)	
q, Flow rate in the aquifer per unit area, (m/s)	
Q, Pumping rate of a borehole, (m^3/s)	
R, Radius of influence or action due to pumping, (m)	
r, Distance to the borehole, (m)	
Re, Reynolds number, (-)	
s, Drawdown due to pumping, (m)	
S, Storage coefficient of a porous medium, (-)	
Ss, Specific storage coefficient, (m/s)	
t, Time, (s)	
T, Transmissivity, (m^2/s)	
U, Darcy velocity or flow velocity, (m/s)	
Vr, Actual flow velocity between pores, (m/s)	
u, THEIS variable, (-)	
W, THEIS function, (-)	
x, y, Space variables, (m)	
Δx, Δy, Space step, (m)	
Δt, Time step, (s)	

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