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A Procedural Schedule For Groundwater Flow In Porous Media

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ABSTRACT: The classical Darcy,'s law is generalized by regarding groundwater flow as a function of the hydraulic head; which is' a quantity of primary interest. This generalized law and the law of conservation of mass are then used to derive the generalized form of the groundwater flow equation. Analytical solution of this groundwater flow equation for which a fractal dimension for the flow is assumed. Equation of unsteady flow in a leaky aquifer is discussed. Prediction of groundwater flow with illustrations of contouring the water table map helps to predict the direction of flow. [New York Science Journal. 2009;2(2):9-19]. (ISSN: 1554-0200).

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INTRODUCTION

A problem that arises naturally in groundwater investigation is to choose an appropriate geometry for the geological system in which the flow occurs. For example, one can use a model based on unsteady state radial flow to simulate the flow in porous media with a very large pore i1uid density (Black *et aI*, 1986). This is in particular the case with the delineation of freshwater aquifer in the Coastal area of Lagos State (lkoyi, Lekki, Apapa and Victoria Island), characterized by the presence of boreholes drilled in these area that serve as the main drawdown in pumping wells. Attempts to fit in analytical solution of the groundwater flow equation with a one dimensional flow and fit a Conventional radial flow model to the observed drawdown at early times underestimates and later times over estimates[1– 4].

The derivation of a generalized groundwater flow equation from the law of mass Conservation and energy balance is usually an indication that the theory is not implemented correctly or does not fit the observations. To investigate the possibility on the Lagos coastal areas. A generalized equation of groundwater flow in three-dimensional equation is expressed as[6-10]

$$\frac{\partial}{\partial x}\left(K\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K\frac{\partial h}{\partial z}\right)=S_s\frac{\partial h}{\partial t}$$

(1)

A fractal one-dimensional groundwater flow equation is assumed as an hypothetical case of a closed aquifer for which the flow is essentially horizontal direction and independent of y and z- axis.

$$\frac{\partial}{\partial x}\left(K\frac{\partial h}{\partial x}\right)=S_s\frac{\partial}{\partial t}$$

$$\nabla.[k\nabla h] = S_s \; \partial_1 \; \mathrm{h} \tag{2}$$

Where S_S the specific storativity
Where K the hydraulic conductivity tensor of the aquifer
Where h(x,t) the hydraulic head with x and t the usual spatial and time coordinate
Where ∇ the gradient operator
Where ∂_1 the time derivative

The model showed that the dominant flow in these aquifers is essentially horizontal and linear and not vertical and radial as commonly assumed. However, more recent investigations (Cloot and Botha, 2006) suggest that the flow is also influenced by the geometry of the bedding parallel fractures, a feature that Equation (3) cannot account for. It is therefore possible that equation may not be application flow in fractured rock other than a porous media[I 1- 15].

In an attempt to circumvent this problem, we introduced a conventional geometry of the aquifer, which assumed a fractal one dimensional flow. (see fig. I). Although this model has been applied with reasonable success in the analysis of the hydraulic head from borehole in the Lagos' Coastal Area[16].

As a review of the derivation of Equation (2) will show [see Bear, 1972], Darcy's Law is used as a keystone in the derivation of Equation (2)[5].

$$\mathrm{q}(\mathrm{x},1) = -k\nabla h \quad \text{... (3)}$$

This law proposed by Darcy early in the 19^{th} century, is relying on experimental results obtained from the flow of water through a one-dimensional sand column, the geometry of which differs completely from that of a fracture [17]. There is therefore a possibility that the Darcy's law not be valid for flow in fractured rock formation but is only a very crude idealization of reality. Nevertheless, the relative success achieved by (Cloot and Botha, 2006) to describe many of the properties of Karoo aquifer on the campus of the university of free State, suggests also that the basic principle underlying this law may be correct: the observed draw down is to be related to either a variation in the hydraulic conductivity of the aquifer or a change in the hydraulic head. Any new form of the law should therefore be reduced to the classical form under a more common condition. Because K is essentially determined by the permeability of the porous medium and not the flow pattern, the gradient term in Equation (3) is the most likely cause for the deviation between the observed and the theoretical drawdown observed in the Karoo formation. In this work, the possibility is further investigated for a flow symmetry form of Equation (2) by creating an artificial vertical fault that divides the aquifer into two compartments of length L, on the left and L_1 on the right. The fault gauge is sufficiently low in hydraulic conductivity that acts as a flow barrier. Thus, the left compartment is hydraulically isolated from the right compartment. Initially, the hydraulic head is h_1 in the left compartment and h_2 in the right compartment. Assume that at time t = 0, the fault is ruptured by an eat1hquake, so that the two compartments are now hydraulically connected. The earthquake would deform the aquifer causing changes in hydraulic head. The question we want to answer is, what happens to the hydraulic head distribution in the aquifer after the fault rupture? [l9]

Therefore, when the fault ruptures, we expect groundwater to flow from compartment with higher head to compartment with lower head, in other word, flow would occur essentially in horizontal x-direction. Analytically, if we set the original length of the x-axis at the left hand boundary, then the How domain is for one-dimensional flow in a homogenous aquifer, the governing in equation (4).

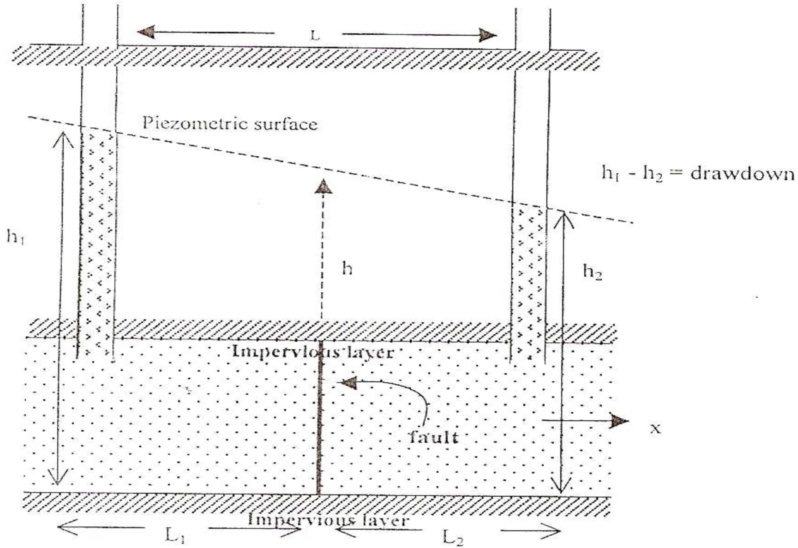

Fig.: Flow through a confined aquifer

The boundary condition are $\frac{\partial^2 h}{\partial x^2} = \frac{S_s}{K}\frac{\partial h}{\partial t}$ (4)

$$0 < x < L_1 + L_2$$

$$\frac{\partial h}{\partial x} = 0 \text{ at } x = 0$$

$$\frac{\partial h}{\partial x} = 0 \text{ at } x = L_1 + L_2$$

The initial conditions are

$$h(x,0) = h_1 \quad \text{for } 0 \le x \le L_1$$

$$h(x,0) = h_1 \quad \text{for } L_1 \le x \le L_1 + L_2$$

Separation of variables is employed and the solution is assumed as:

h(x,t) = f(x) . g(t) (5)

where:

$$a = \frac{S_s}{K} \text{ and } f(0) = 0 \text{ for } K_1 = \frac{n\pi}{L}$$

Using these in solution of (5), we have,

$$h(x,0) = A_0 + Cos\frac{n\pi}{L}x\lambda\frac{n^2 n^2 Kt}{S_s L^2}$$

Initial condition, we have,

$$h(x,0) = A_0 + \sum_{n=1}^{\infty} A_n Cos\frac{n\pi}{L}x\lambda\frac{n^2 n^2 Kt}{S_s L^2}$$

(6)

Substitute for A_0 and A_n from Fourier integral, we have

$$A_0 = \frac{1}{L}\int_0^L f(x)dx \qquad \text{and}$$

$$A_0 = \frac{2}{L}\int_0^L f(x)\,\mathrm{Cos}\,\frac{\mathrm{n}\,\pi}{\mathrm{L}}dx$$

$$h(x,t) = \frac{1}{L}\int_0^L f(x)dx + \frac{2}{L}\sum_{n=1}^{\infty}\lambda\frac{n^2n^2Kt}{S_sL^2}\,\mathrm{Cos}\,\frac{\mathrm{n}\,\pi x}{\mathrm{L}}\int_0^L f(x)\,\mathrm{Cos}\,\frac{\mathrm{n}\,\pi x}{\mathrm{L}}dx$$

(7)

Let x be a dummy variable of integration. To find the solution to the flow equation, we replace L in Equation (7) by L_1+L_2 and in addition, we replace f(x) by h(x,0) as defined. The integral inside the summation on the right hand side of equation (7) and substituting the preceding integral, we have.

$$h(x,t) = \frac{h_1L_1 + h_2L_2}{L_1 + L_2} + \frac{2}{L_1 + L_2}\sum_{n=1}^{\infty}\lambda\frac{n^2n^2Kt}{S_sL_1 + L_2}Cos\frac{n\pi x}{L_1 + L_2}\frac{(h_1 - h_2)(L_1 + L_2)}{n\pi}\,\mathrm{Sin}\frac{n\pi L_1}{L_2 + L_2}$$

$$h(x,t) = \frac{h_1L_1 + h_2L_2}{L_1 + L_2} + \frac{2(h_1 - h_2)}{\pi}\sum_{n=1}^{\infty}\frac{1}{n}\lambda\frac{n^2n^2Kt}{S_s(L_1 + L_2)^2}Cos\frac{n\pi x}{L_1 + L_2}\ \mathrm{Sin}\frac{n\pi L_1}{L_2 + L_2}$$

….(8)

This solution can be expressed in dimensionless form as:

$$h_D(x_D, t_D) = L_D + \frac{2}{\pi}\sum_{n=1}^{\infty}\frac{1}{n}\lambda^{-n^2\pi^2t_D}Cos(n\pi x_D)Sin(n\pi L_D)\ldots$$

(9)

Where:

The dimensionless distance- $$x_D = \frac{x}{L_1 + L_2}$$

The dimensionless time - $$t_D = \frac{Kt}{(L_1 + L_2)^2 S_s}$$

The dimensionless - $$L_D = \frac{L_1}{L_1 + L_2}$$

One advantage of a closed form analytical solution is that it allows us to examine the behaviour of the flow system. There are several interesting features in this solution. The first term on the' right hand side is the steady state part of the solution. It gives the head in the aquifer when *t* is very large (see fig. 3). The second term on the right hand side is the transient part of the problem. Because *t* appears in the argument of the exponential function, the second term tends to zero as *t* becomes large. Furthermore, note that the second term goes to zero at a faster rate if K/Ss (hydraulic diffusivity) is large. Thus, the hydraulic diffusivity is a quantity that controls the rate of hydraulic head.

Unsteady Flow in a Leaky Aquifer

The generalized groundwater flow equation in a leaky aquifer is of the form,

$$S_s \frac{\partial h}{\partial t} = K\nabla^2 h - G$$

(10)

where $G = \frac{e}{T}$, and $\mathrm{e} = \mathrm{K}^1 \frac{h_0 - h}{b^1}$ can determined form Darcy's law.

Under this condition, the above equation becomes a radial flow,

$$T\left(\frac{\partial^2 S}{\partial r^2} + \frac{1}{r}\frac{\partial S}{\partial r}\right) - S\frac{\partial}{\partial t} - q = 0$$

where

$$q = \frac{K^1}{b^1} S = \frac{S}{C}$$

(11)

Using the same approach as the solution for the confined (Theis) solution, we obtained the leaky partial different equation:

$$u\left(\frac{\partial^2 S}{\partial u^2} + \frac{\partial S}{\partial u}\right) + \frac{\partial S}{\partial u} - \frac{r^2}{4uL^2} S = 0$$

…(12)

and from separation of variables, we obtain an appropriate solution

$$S = \frac{Q}{4\pi T} W\left(u, \frac{r}{B}\right)$$

…(13)

The quantity $\frac{r}{B}$ is given by $\frac{r}{\sqrt{T/\left(K^1/b^1\right)}}$

Which holds as long as $U < 0.01$

Where W $\left(u, \frac{r}{B}\right)$ is dimensionless form from a logarithm plot chart.

The plot of S = *hi - h;* versus *t* at various observation wells, since drawdown is the hydraulic heads measured the level of the water table in wells relative to the piezometric surface (see fig. I). The change in water table in the pumping well or in observation well nearby is referred to as drawdown (see fig. 5).

APPLICATION

Set of drawdown data was analyzed in order to validate the new method. The examples were obtained from borehole drilled along the coastal area of Lagos State. The boreholes belong to companies operating along the coastal area. The examples were to illustrate the application using equations developed in this case.

Example 1

A well is located in an aquifer with a conductivity of 15 meters per day and a storativity of 0.005. The aquifer is 20 meters thick and is pumped at a rate of 2725 cubic meters per day. What is the drawdown at a distance of 7 meters from the well after one day of pumping?

- Hydraulic conductivity = 15 metres per day
- Storativity = 0.005

- Aquifer thickness = 20 metres
- Pumping rate = 2,725 cubic metres per day
- Distance from the well = 7 metres

T=Kb = m/day x 20m = 300 m^2/day

$$u = \frac{r^2 S}{4Tt} = \frac{(7m)^2 \times 0.005}{4 \times 300m^2 / day \times 1day} = 0.0002$$

From the table of W(u) and u, if u = 2 x10^{-4}, w(u) = 7.94:

$$h_1 - h_2 = \frac{Q}{4\pi T} W(u) = \frac{2727m^3 / day \times 7.94}{4 \times \pi \times 300m^2 / day} = 5.73m$$

The draw is 5.73 meters after one day.

Example 2

A well in a confined aquifer was pumped at a rate of 220 gallons per minute for about 8 hours. The aquifer was 18 feet thick. Time drawdown data for an observation well 824 feet away are given in table 2. Find T, K, and S.

W(u) = 1

I/u = 1

$h_1 - h_2$

$t/r^2 = 6.06 \times 10^{-6}$

Radial Diameter d = 20ft

Pumping Rate = 220 gallons per day for 8 hours

Aquifer thickness = 18 feet

Transmissivity:

$$T = \frac{114.6QW(u)}{h_0 - h} = \frac{114.6 \times 220 \times 1.0}{2.4} = 10{,}500 \ \ galonsgpd \,\mathrm{l\,ft}$$

Hydraulic Conductivity:

$$K = \frac{T}{b} = \frac{10{,}500}{18} = 580 gpd / \mathrm{ft}^2$$

Storativity

$$S = \frac{Tu}{2693} \times t / r^2 = \frac{10{,}500 \times 1}{2693} \times 6.06 \times 10^{-6} = 0.00002 \ldots. \quad \text{(Theis method)}$$

$$S = \frac{Qr^{1-n}}{4\pi Td} = \frac{220}{4\pi(10{,}500)(20)} = 0.000018 \quad \ldots \quad \text{(Cloot \& Botha method)}$$

$$S = \frac{QW(u)}{4\pi Td} = \frac{220 \times 1.0}{4 \times \pi \times 10{,}500} = 0.0016 \quad \ldots. \quad \ldots..$$

(Observation)

(See fig 4 & 5)

Example 3

An aquifer 10 meters thick is penetrated by a well It is overlain by a semipervious layer I meter thick with a K" of I 0-^{5} centimeter per second. There is no storage in the leaky confining layer. The aquifer has a K of] 0.^{2} centimeter per second and an S of 0.0005. If a well pumps at 500 cubic meters per day, compute values of drawdown at I, 5, 10, 50, 100, 500, and 1000 meters. (see Table 2)

Aquifer Thickness = 10 metres
Storativity = 0.0005
Pumping rate = 500 cubic metres per day
Various depths = I, 5, 10, 50, 100,500 & 1,000 metres
r = distance to the observation wells
t = time since pumping begin
K = 10^{-2} cm/sec x 60 sec/min x 1440 min/day x 10^{-2} m/cm = 8.64 m/day
K' = $10-^{5}$ cm/sec x 60 sec/min x 1440 min/day x $10-^{2}$ m/cm= 8.64x 10^{-3} m/day
b'=-lm
b’=l0m
T = Kb = 86.4 m^2/day
B = $(Tb'/K')^{1/2}$
= (86.4 m^2/day x 1 m /8.64 x $1O.^{3}$/day$)^{1/2}$
$(10^4)^{1/2}$
B = 100

$$u=\frac{r^2 s}{4Tt}=\frac{r^2\times 0.0005}{4\times 86.4\times 1}=1.44\times 10^{-6}r^2$$

$$\frac{r}{B}=\frac{r}{100}=10^{-2}r$$

$$h_1-h_2=\frac{2.6Q}{4\pi T}W\left(u,\frac{r}{B}\right)=1.06W\left(u,\frac{r}{B}\right)$$ (our observation) see fig. 5

As u = 1.44x $10^{-6}r^2$, we can find the value of u for each r-value

4.1 Discussion of Results:

All the three examples of drawdown data show that the new method underlying this law and the observed drawdown variations in hydraulic conductivity of the aquifer is correct. Each of the analytical solution describes the response to pumping in a very idealized representation of aquifer configurations. In the real world, aquifers are heterogeneous and isotropic: They usually vary in thickness; and they certainly do not extend to infinity. Where they are bounded, it is not by straight-line boundaries that provide perfect confinement. Aquifers are created by complex geologic processes that head to irregular stratigraphy and trendouts of both aquifers and aquitards. The Predictions that can be carried out with the analytical solution presented in this paper must be viewed as best estimates. In general, hydraulic head solutions are most applicable when the unit of study is a well.

They are less applicable on a large scale, where the unit of study is an entire aquifer.
The graphical method of solution starts with the construction of reversed type curve of W(u) against I/u on logarithm paper (see fig. 4). Data from observation well located at different distances from the pumping wells were used. If there is only one observation well, then it is sufficient to plot hi - h_2 as a function of *t* (table I). .

Using "Contouring the Water Table Map", we noticed that the contours form V's with the river and its tributaries. That's because the river is a "gaining" river. It is receiving recharge from the aquifer. The contours show that ground water is moving down the sides of the valley and into the river channel. The opposite of a gaining stream is a "losing" stream. It arises when the water table at the stream channel is lower than the stream's elevation or stage, and stream water flows downward through the channel to the water table. This is very common in dryer regions of the Southwest. In the case of a losing stream, the V will point downstream, instead of upstream. (see fig. 6)

When making a water table map, it is important that your well and stream elevations are accurate. All elevations should be referenced to a standard datum, such as mean Sea level. This means that all elevations are either above or below the standard datum (e.g., 50 feet above mean sea level datum). It's also very important to measure all of the water table elevations within a short period of time, such as one day, so that you have a "snapshot" of what's going on (Adeosun *et al* 2006). Because the water table rises and falls over time, you would be more accurate if readings are made before these changes occur.

Understanding how ground water flows is important when you want to know where to drill a well or a water supply, to estimate a well's recharge area, or to predict the direction of contamination is likely to take once it reaches the water table. Water table contouring can help groundwater developer to do all these things. Hence, groundwater flow through the subsurface is the whole essence of this paper and called for further investigation.

Table 1: Drawdown Table

Time After Pumping Started (min)	T/r^2	Drawdown (ft)
3	4.46 x 10^{-6}	0.3
5	7.46 X 10^{-6}	0.7
8	U8 x $l0^{-5}$	1.3
12	1.77 x 10^{-5}	2.1
20	2.95 x 10^{-5}	3.2
24	3.53 x 10^{-5}	3.6
30	4.42 x 10^{-5}	4.1
38	5.57 x 10^{-5}	4.7
47	6.94 x 10^{-5}	5.1
50	7.41 x$l0^{-5}$	5.3
60	8.85 x $l0^{-5}$	5.7
70	1.03 x 10^{-4}	6.1
80	1.18 x 10^{-4}	6.3
90	1.33 x 10^{-4}	6.7
100	1.47 X 10^{-4}	7.0
130	1.92 x 10^{-4}	7.5
160	2.36 x 10^{-4}	8.3
200	2.95 x 10^{-4}	8.5
260	3.83x$l0^{-4}$	9.2
320	4.72 x 10^{-4}	9.7
380	5.62 x 10^{-4}	10.2
500	7.35 x 10^{-4}	10.9

Table 2: Field Data

R	U		W
1m	1.44 x 10^{-6}	0.01	9.44
5m	3.6 x $l0^{-5}$	0.05	6.23
10 m	1.44 x 10^{-4}	0.1	4.83
50 m	3.6 x 10^{-3}	0.5	1.85
100m	1.44 x 10^{-2}	1	0.824
500 m	6 x 10^{-1}	5	0.007
1000 m	1.44	10	00001

From the computed values of W(u, r/u) at each observation point, the drawdown can be computed from h_0 - h = 1.06 W(u, r/b)

Table 3

R	h_1 - h_2
1m	9.44m
5m	6.23m
10 m	4.83m
50 m	1.85m
100m	0.824m
500 m	0.007m
1000 m	00001m

CONCLUSION

It has been clearly demonstrated that the study of flow in porous media was recognized in detailed through the physical behaviour of subsurface water and their interactions with the solid matrix (flow of groundwater was delineated through the presence of boreholes drilled along the coastal area of Lagos State for characterizing the flow in the subsurface aquifer. The classical Darcy's law governed the flow in porous media by regarding ground water flow as a function of the hydraulic head. A complete statement of this flow problem required specifying the extent of the flow domain, the governing equation, spatial distribution of properties, for example, hydraulic conductivity and specific storativity, boundary conditions and initial conditions. Analytical solution of this flow equation for which a fractal dimension was assumed to yield a closed form solution that could be written on paper and also be examined to understand the behaviour of the flow system in a typical limited homogeneous flow domain with relatively simple geometry.

The problem of solving fluid flow through porous media has proved analytically intractable and the problem of understanding flow and storage in aquifers is very complex. It was ·recognized that flow through' such a medium is very significantly influenced by the porous media characteristics such is porosity and permeability. A limitation of this work is the estimation of permeability (hydraulic conductivity) of the medium which can not be examined and investigated without being to the field, even if examined, permeability estimation has proved to be complex and this concept has limited the free flow of fluid within the porous medium. Therefore, this work is hoped to complement the study of flow through porous media that might have been done in other parts of the world and contributes to the unveiling knowledge of the applicability of flow in porous media. Prediction of this flow shows several interesting qualitative features such as graphs and contouring of water table map, which held to predict change in drawdown in pumping well and the direction of flow. The method becomes more accurate and easy to handle with little or no variations in the observed drawdown and water table flow prediction.

NOMENCLATURE

S = Storativity
T = Transimissivity (L^2T^1)
h_1 -h_2 = drawdown (L)
Q = pumping rate (L^3T^1)
t = time, (time since pumping began)
r = radial distance from pumped well (L)
e = leakage rate
B = leakage factor (Lb^{-1}) = thickness of leaky layer (L)
b^l = thickness of leaky layer (L)
K^1= vertical hydraulic conductivity of leaky layer (LT^{-1})
(X, y)= rectilinear coordinator
S_s = specific storage

h = head (L)
q = specific discharge (M^3d^{-1} per m^2)
K = hydraulic conductivity of aquifer (md^{-1})
h_2, h_1 = hydraulic heads measured along flow path
L = distance between head measurements (m)
W = width of cross - sectional flow (m)
D = height of cross-sectional flow (m)
W(u)= dimensionless form from chart
G = leakage factor
U = flow velocity

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