TECHNICAL REPORT TR (BR) 122

PARAMETER DETERMINATION IN A SEMIPERVIOUS STREAM - AQUIFER SYSTEM

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1994-95

PREFACE

Conjunctive use management of surface water and groundwater in a river basin has received considerable importance in the recent past. Stream-aquifer interaction studies are imperative for better management decision. Many numerical and analytical models dealing with stream aquifer interaction (Cooper and Rorabaugh, 1963; Hornberger et al., 1970; Morel- Seytoux, 1979; Mishra, 1987; etc.) have been developed in the past.

Most of the stream-aquifer interaction models assume the stream to be a fully penetrating one, though, in most of the cases, the stream penetrates the aquifers partially and has low hydraulic conductivity sediments deposited in its bottom. An extra resistance to flow is offered by these fine deposites.

Application of models of stream-aquifer interaction requires the values of aquifer parameters as a priori. These parameters, in case of a semi-pervious stream, are aquifer diffusivity and stream resistance. 'Stream resistance' is the resistance to flow offered by semi-perviousness and partial penetration of the stream.

It is possible to determine these parameters, by analyzing the response of the aquifer to the fluctuation in the stream stage. Parameters thus determined are more appropriate to be used for quantifying stream-aquifer interaction as compared to using of parameters obtained from pump tests. This approach does not require the energy for the excitation of the system. This approach has an other added advantage over the pump test that stream resistance can be determined.

In this report, the direct problem (prediction of head in the aquifer when parameters are known) has been solved using the concept of 'retardation coefficient'. The discrete pulse kernels have been employed to determine the aquifer response to the time-varying stream-stage. For the inverse problem, a non linear optimization technique (Marquardt, 1963) has been used and a model has been developed for the estimation of parameters in a semi-pervious stream aquifer system. The results of the model application on the published data of glacial-outwash aquifer near Cortland, NewYork (Reynolds, 1987) show that this model is applicable for similar field problems involving semi-pervious stream.

The present study has been carried out by Shri S.K. Singh, Scientist 'C' as per the work programme of the Ground Water Modelling and Conjunctive Use Division.

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ABSTRACT

The study of stream-aquifer interaction is a prerequisite for conjunctive use management. Many numerical and analytical models for stream aquifer interaction have been developed in the past, these include models developed by Cooper and Rorabaugh (1963), Hornberger et al.(1970), Morel- Seytoux and Daly(1975), Mishra(1987), etc.

Most of the investigators have dealt with fully penetrating streams. However, streams are generally partially penetrating with a semi-pervious bed of low hydraulic conductivity. Consequently, an extra resistance to flow is offered by a partially penetrating stream due to curved flow lines near the stream and low hydraulic conductivity of the fine sediments deposited at the bottom of the stream. Only a few researchers (Halek and Svec, 1979; Hall and Moench, 1972; etc.) have considered this aspect.

Application of models of stream-aquifer interaction requires the aquifer parameters be known a priori. These parameters are aquifer-diffusivity and stream resistance. In this report first the direct problem (direct problem is the prediction of water table induced due to stream-stage variation with known parameters) and then the inverse problem (inverse problem is the determination of parameters with known variations in stream-stage and induced water table) have been solved.

The direct problem has been solved using the concept of 'retardation coefficient' as proposed by Hantush(1965) for modelling a semi-pervious stream. Discrete pulse kernel coefficients have been employed to determine water table induced due to stream-stage variation. For the inverse problem, a non

linear optimization technique (Marquardt, 1963) has been used and a model has been developed for the determination of parameters of a semi-pervious stream and aquifer system.

Applicability of the model has been verified with published data (Reynolds, 1987) and it has been observed that the present method predicts aquifer response more correctly as compared to that used by Reynolds. This shows that the present model is applicable to similar field problems where the stream is partially penetrating and has semi-perviousness at its bed and bank.

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1.0 INTRODUCTION

Considerable emphasis has been given on the conjunctive use management of water resources in recent past. One of the important aspects of conjunctive use management is the study of stream-aquifer interaction. Many numerical and analytical models for stream aquifer interaction have been developed in the past. Valuable contributions to the understanding of stream-aquifer relationship have been made by Cooper and Rorabaugh(1963), Hornberger et al.(1970), Morel-Seytoux(1979), Morel-Seytoux and Daly(1975), Mishra(1987), Singh(1994), etc.

In most of the stream-aquifer interaction studies, stream has been assumed to completely penetrate the entire depth of aquifer. In nature, generally a stream penetrates an aquifer partially except for big rivers that have huge discharges. The bed of a partially penetrating stream may be comprised of fine sediments deposited by the stream. In such case, the bed of the partially penetrating stream has a lower hydraulic conductivity than that of the aquifer material. Thus, an extra resistance to flow is offered by a partially penetrating stream due to curved flow lines near the stream and low hydraulic conductivity of the bottom of the stream. The theories developed for fully penetrating streams can not be applied in true sense for the stream that partially penetrates the aquifer and for a fully penetrating stream that has clogged banks. Very few investigators (e.g., Halek and Svec, 1979; Hall and Moench, 1972; Morel-Seytoux and Daly, 1975, Mishra, 1987, Singh 1994, etc.) have analyzed the interaction of a partially penetrating stream and an aquifer.

Application of models of stream-aquifer interaction requires the aquifer parameters be known a priori. These parameters are,

i) aquifer diffusivity (ratio of the transmissivity and storage coefficient) in case of a fully penetrating river;

ii) and aquifer diffusivity and river resistance, in case of a partially penetrating or a semi-pervious stream. River resistance is the resistance to flow offered by a partially penetrating stream.

It is possible to determine above parameters, i.e., diffusivity of flood-plain aquifer, and stream resistance of a partially penetrating aquifer by analyzing the response of the aquifer to the fluctuation in the stream stage. Parameters determined in this way,are more appropriate to be used for quantifying stream-aquifer interaction as compared to using of parameters obtained from pump tests. This approach does not require the energy for the excitation of the system . This approach has an other added advantage over the pump test that stream resistance can be determined.

The problem of identifying parameters of a system, e.g., an aquifer in general context of modelling, is known as inverse problem. The inverse problem in context to the stream-aquifer interaction with fully penetrating stream has been solved by many investigators (Ferris,1952; Rowe,1960; Pinder et al.,1969; Singh and Sagar, 1977; Reynolds, 1987; Mishra and Jain, 1992; etc.). Most of these solutions have been obtained for sinusoidal and linear stream-stage variations. Stream stages that occur in nature may not always be approximated by mathematical relationship, such as linear, exponential or sinusoidal. Singh and Sagar (1977), Reynolds (1987); and Mishra and Jain, 1992 have taken into account any variation in the stream-stage. The above inverse problem has not been solved for a semi-pervious or partially penetrating stream so far.

In this report, the direct problem (direct problem is the prediction of head in the aquifer when parameters are known) has been solved using the concept of 'retardation coefficient' or 'substitute length' as proposed by Hantush(1965) and Halek and Svec(1979) respectively for modelling a semi-pervious stream. The convolution technique has been employed to determine the aquifer response to the time-varying stream-stage. The variable stream-stage perturbation has been approximated as a train of pulses and accordingly, discrete kernel for aquifer-heads have been obtained.

For the inverse problem, a non linear optimization technique proposed by Marquardt (1963) has been used and a model has been developed for the estimation of parameters in a semi-pervious stream aquifer system. The results of model application on the published data of glacial-outwash aquifer near Cortland, NewYork (Reynolds, 1987) have been discussed.

2.0 REVIEW

This chapter deals with the review of work done by various investigators in the past. In section 2.1, a general review has been presented in chronological order while section 2.2 deals with the critical summary of the past studies.

2.1 GENERAL:

A method for determining aquifer constants from river level fluctuations has been presented by Ferris (1952). He has related the aquifer constants to the groundwater variations corresponding to a sinusoidal variation in the surface water bodies.

Rowe (1960) has derived an equation from which aquifer diffusivity can be estimated. In the derivation, he has assumed linear change of water level with time in the water body. He has compared the aquifer-diffusivity obtained from experimental data for steady state condition with that obtained for unsteady state condition.

Pinder et al. (1969) have developed a numerical model to determine the diffusivity of a homogeneous isotropic aquifer from its response to fluctuations in river stage approximated by a series of incremental steps. They have considered the river to penetrate the full depth of the aquifer and have used the method of super-position.

Singh and Sagar (1977) have evolved an analytical method to determine aquifer diffusivity explicitly from the measurements at the boundary of intersection between a fully penetrating stream and an aquifer. An extra boundary condition, i.e., hydraulic gradient at the interface is specified, which has been used to determine the flow property explicitly. The Boussinesq equation has been chosen as a mathematical model to describe the influence of a rising stream on the water level in the aquifer. The rising water level has been chosen because Dupuit's assumptions are more likely to hold good during this stage. Four types of stream stage hydrographs, i.e., linearly rising water level, exponentially rising water level, water level represented by a sinusoid, and hydrograph approximated by cubic splines, have been used to determine aquifer diffusivity. The results of the model application on the data from the sand tank model, have been presented.

Using the linearized Boussinesq equation, Sagar and Singh (1979) have described the flow in a semi-infinite aquifer with a fully penetrating stream at its boundary. The recorded stream hydrograph and the computed hydraulic gradient at the boundary have been assumed to contain Gaussian noise. The aquifer diffusivity has then been found to be given by the ratio of squares of two normal random variables. The mean and the standard deviation of the diffusivity have been computed for various levels of errors in the boundary data. It has been observed that the mean value of the diffusivity is a function of the errors in data and increases with increasing errors. The increase in the mean diffusivity, however, is not of as much significance as the increase in the standard deviation, which increases at a much faster rate with increasing errors. Any significant increase in the standard deviation would reduce the confidence in the computed diffusivity value. Based on the acceptable value of standard deviation in the diffusivity, one can, therefore, stipulate bounds on permissible errors in data.

Reynolds(1987) has used the flood Wave response model developed by Pinder and others (1969) to calculate the diffusivity of a glacial out-wash aquifer from flood wave response data collected at three sites across a river. The model generates the hydrograph using Duhamel's principle:

$$h_p = \sum_{m=1}^{p} \sum_{n=1}^{\infty} (-1)^{n-1} \Delta H_m \left[\text{erfc}\{0.5U \frac{(2n-2)/(x'/L)+1}{\sqrt{(p-m)}}\} - \text{erfc}\{0.5U \frac{(2n-2)/(x'/L)+1}{\sqrt{(p-m)}}\} \right] \quad ...(2.1)$$

where, $U = x'/\sqrt{(T/\phi)\Delta t}$ and;

L = distance from the river to impermeable boundary, (L);

x' = distance from the point where aquifer response is observed to impermeable boundary, (L);

h_p = head at a distance (L-x) from the river at time $p\Delta t$, (L);

p = total number of time intervals for the simulation, (dimensionless);

ΔH_m = instantaneous rise in stage at time $t=m\Delta t$, where, m= an integer representing a number of time intervals, (L);

T/ϕ = diffusivity of the aquifer (transmissivity divided by storage coefficient), (L^2T^{-1});

$\text{erfc}(\nu) = 1 - 2/\sqrt{\pi} \int_0^{\nu} \exp(-u^2)du$, i.e., complementary error function, (dimensionless) and;

Δt = unit time step duration, in fraction of a day, (T).

Equation (2.1) gives the response of an isotopic homogeneous aquifer of finite width due to stream stage fluctuations in a fully penetrating stream. The following assumptions were made in the derivation of the equation:

i) T and ϕ are constant and uniform,

ii) the component of flow in the aquifer produced by flood wave is essentially one dimensional.

The observed response of the aquifer has been found to resemble with the theoretical response obtained from the model. The matching is poor during the peak period of flood (difference between observed and computed value of head at peak is 19%). He has considered the head variation in an observation well situated very close to the stream boundary as the stream-stage variation rather than using the actual stream-stage fluctuation.

Mishra and Jain(1992) have developed a method for determining the transmissivity and storage coefficient of an aquifer separately using the river-level fluctuations and corresponding head variation in the aquifer at three locations in the vicinity of the stream. They have stated that at one site out of the three sites, observation well should have large diameter. They have proposed a Laplace transform technique for parameter estimation and have compared the results with that obtained using Marquardt algorithm for linear optimization. The applicability of the methods have been tested on the data published by Reynolds(1987).

2.2 CONCLUDING REMARK:

The direct problem of semi-pervious/partially penetrating stream and aquifer interaction has been attempted by Hantush(1965), Hall and Moench(1972), Halek and Svec (1979), and Singh (1994) employing the concept of 'stream resistance' and by Morel-Seytoux and Daly(1975), and Mishra(1987), using the 'reach transmissivity' approach. Morel-Seytoux and Daly(1975), Mishra(1987), and Singh(1994) have used the method of convolution employing the discrete kernels. Singh(1994) have solved for the complete flow field, i.e., head, velocity, discharge, and volume. A detailed review of the subject may be found in Singh(1994).

The inverse problem with fully penetrating stream has been solved by many researchers. Ferris(1952) has considered a sinusoidal variation in the stream-stage; Rowe(1960) has taken the stream-stage to vary linearly with time; Singh and Sagar(1977) have given the explicit analytical expression for determining aquifer diffusivity for linear, exponential, and sinusoidal variations in the stream-stage. They have also considered the case when the stream-stage is approximated by piece-wise cubic splines. Pinder et al.(1969) and Reynolds(1987) have dealt with a numerical model approximating the stream-stage by incremental steps. Mishra and Jain(1992) have used a discrete kernel approach and Laplace transform technique.

The review of the literature shows that the inverse problem has not been solved for a semi-pervious or partially penetrating stream so far.

3.0 PROBLEM DEFINITION

The purpose of the present study is to evolve a method that will enable the determination of the aquifer diffusivity of aquifer in flood plain and the 'stream resistance' (substitute length or retardation coefficient) analyzing the stream-stage variation and corresponding variation of piezometric head in the aquifer in a semi-pervious stream and aquifer system. 'stream resistance' expresses the effect of resistance induced by the curvature of streamlines as well as the effect of other inlet resistances.

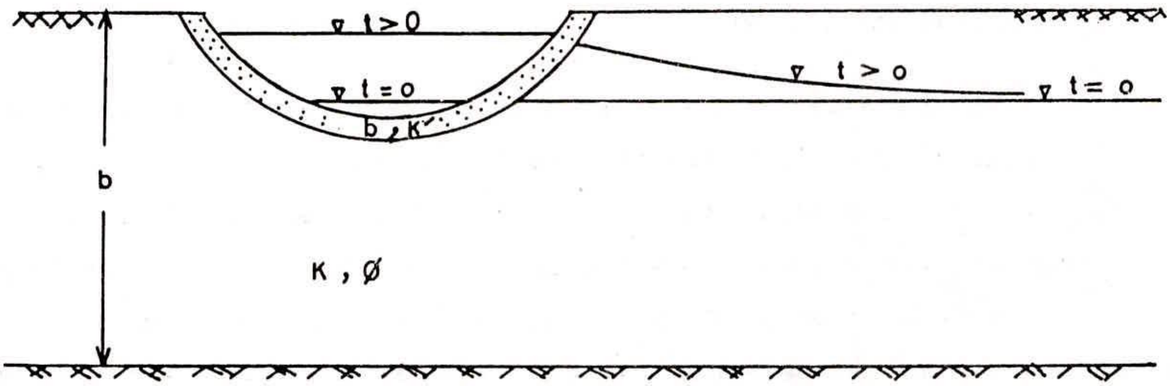
The definition sketch of the problem is shown in figure 3.1(a). A stream with semi-pervious bottom partially penetrates an infinite aquifer. The aquifer is homogeneous and isotropic and has an impervious boundary at the bottom.

Initially, the phreatic surface is horizontal and coincides with the stream-stage. A flood wave passes in the stream. The change in the phreatic level is small as compared to the saturated thickness of the aquifer (so that the theory of flow in confined aquifer is valid for this case).

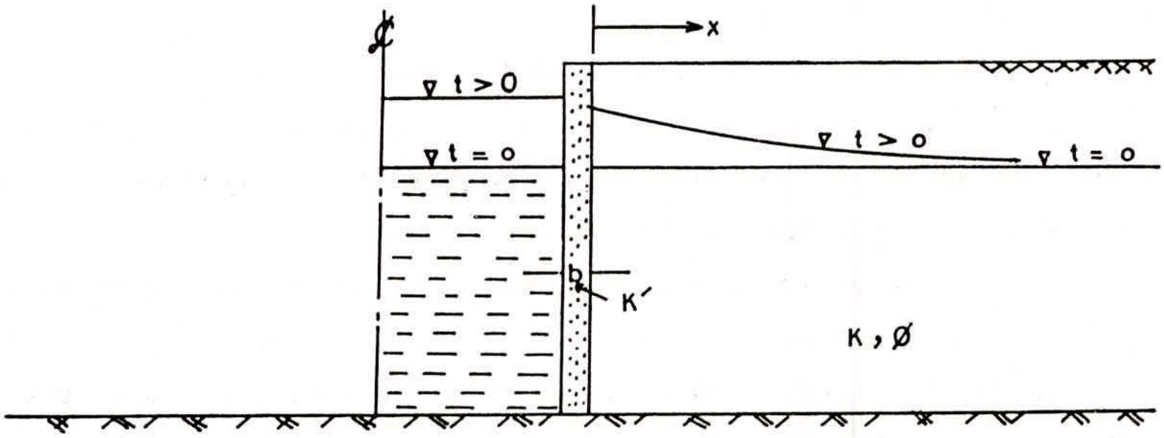
It is required to find the parameters of the stream aquifer system using Marquardt algorithm for the following cases.

1. Using the synthetic data corresponding to Cooper and Rorabaugh's(1963) assymmetric flood.

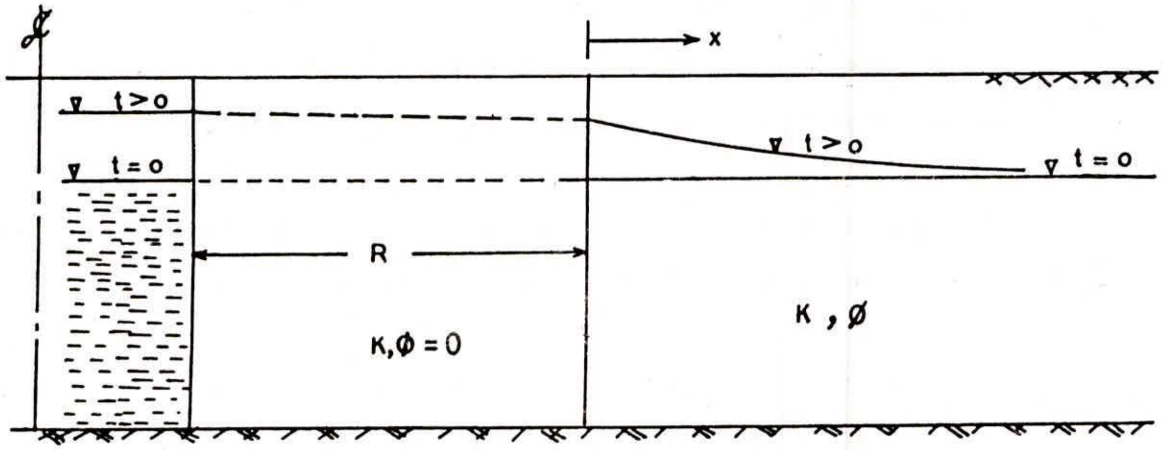
2. Using the published data of glacial-outwash aquifer near Cortland, NewYork (Reynolds, 1987).



(a) Cross-section of a Semi-Pervious Stream



(b) Modelled Semi-Pervious Stream



(c) Equivalent Fully Penetrating Stream with 'Substitute Length'

Fig.3. Definition Sketch of the Problem.

4.0 METHODOLOGY

In this section, first the methodology for direct problem has been discussed and then the methodology for inverse problem has been presented.

4.1 DIRECT PROBLEM:

A partially penetrating river forms a boundary condition of third type (Fourier's condition). Figure 3.1(a) shows a partially penetrating stream with semi-permeable bed and an aquifer. The semi-pervious layer is of thickness b' having hydraulic conductivity K'. In order that analytical solution be obtained, the above problem is transformed as shown in fig. 3.1(b&c) in which the semi-pervious stream has been replaced by a fully penetrating stream. The extra resistance to flow due to semipervious bed of the stream has been taken into account by a fictitious aquifer length R which may be expressed as (Hantush, 1965).

$$R = bK/K' \qquad ...(4.2)$$

where, K is the hydraulic conductivity of the aquifer material.

The parameter, R as defined by Halek and Svec(1979) is 'substitute length' which expresses the effect of resistance induced by the curvature of streamlines as well as the effect of other inlet resistances. Within length R $(-R<x<0)$, there is a pressure flow without storage of water in the aquifer, while for $0<x<\infty$, phreatic surface occurs, the origin being at the boundary of the stream. Hantush(1965) has defined R as 'retardation coefficient' which is the effective thickness of aquifer required to cause the same head-loss as that by semi-pervious stream bank. Obviously, for a fully penetrating stream, R is zero.

Initially, the water-table and the stream water level were taken equal. The water-level in the stream is suddenly raised by H and maintained constant thereafter. Then, the solution for the piezometric head in the aquifer is known from heat transfer theory (Carslaw and Jaeger, 1959) and is given by:

$$\bar{h}(x,t)=H[\mathrm{erfc}(x/\sqrt{4\beta t})-\mathrm{erfc}(x/\sqrt{4\beta t}+\sqrt{\beta t}/R)\exp(x/R+\beta t/R^2)] \quad \ldots(4.3)$$

Where, β is the hydraulic diffusivity of the aquifer and is equal to T/ϕ, (where, T is the transmissivity and ϕ is the storage coefficient of the aquifer. From equation 4.3, the unit step response function for piezometric head, $U(x,t)$ is given by

$$U(x,t)=[\mathrm{erfc}(x/\sqrt{4\beta t})-\mathrm{erfc}(x/\sqrt{4\beta t}+\sqrt{\beta t}/R)\exp(x/R+\beta t/R^2)] \quad \ldots(4.4)$$

Carslaw and Jaeger, (1959) gave the following expression for the instantaneous unit impulse response for the piezometric head, i.e., $u(x,t)$.

$$u(x,t)=\sqrt{\beta/(\pi t)}\exp(-x^2/4\beta t)/R$$
$$-(\beta/R^2)\exp(x/R+\beta t/R^2)\mathrm{erfc}(x/\sqrt{4\beta t}+\sqrt{\beta t}/R) \quad \ldots(4.5)$$

Let the stream water level varies with time, i.e., H=H(t) and the time-span be discretized into a number of uniform time-steps, Δt. The stream-stage perturbation has been approximated as a train of pulses, the stream water level during a time step has been assumed constant equal to the mean value for the time step.

Treating the variable input as a train of pulses, head in the aquifer at the end of n^{th} time-step can be expressed as:

$$h(x,n\Delta t) = \int_{0}^{n\Delta t} H(\tau)u(x,n\Delta t-\tau)d\tau =$$

$$= \sum_{\gamma=1}^{n} 0.5(H_{\gamma-1}+H_{\gamma}) \int_{(\gamma-1)\Delta t}^{\gamma\Delta t} u(x,n\Delta t-\tau)d\tau$$

$$= \sum_{\gamma=1}^{n} 0.5(H_{\gamma-1}+H_{\gamma})\delta_{1}(x,\Delta t,n-\gamma+1) \qquad \ldots(4.6)$$

Where, $\delta_1(x,\Delta t,m)$ is defined as the discrete pulse kernel for piezometric head in the semi-pervious stream and the aquifer system, and is given by,

$$\delta_1(x,\Delta t,m)= \int_{0}^{\Delta t} u(x,m\Delta t-\tau)d\tau =-\Delta F_1(m)-\exp(x/R)\Delta F_3(m) \qquad \ldots(4.7)$$

where,

$F_1(m) =\mathrm{erf}(x/\sqrt{4\beta m\Delta t})$, $F_1(0)=1$, $\Delta F_1(m)=F_1(m)-F_1(m-1)$;

$F_3(m) =\exp(\beta m\Delta t/R^2)\mathrm{erfc}(x/\sqrt{4\beta m\Delta t}+\sqrt{\beta m\Delta t}/R)$; $F_3(0)=0$; and

$\Delta F_3(m)=F_3(m)-F_3(m-1)$.

For fully penetrating stream (R=0), discrete pulse kernel for piezomeric head at x is given by $-\Delta F_1$.

4.2 PARAMETER DETERMINATION USING MARQUARDT ALGORITHM:

If in a semi-perviuos stream and aquifer system, the stream-stage variation and corresponding variation in the piezometric head at a distance from the stream boundary but within the influence zone of the stream, is known , the parameters R and β can be determined using a suitable optimization technique. Since, the equations representing the aquifer head, i.e, (4.6) and (4.7) are nonlinear functions of the parameters, only nonlinear optimization technique can be used to determine the parameters. For the present analysis, Marquardt(1963) algorithm, that minimizes the sum of the squared deviation between observed and calculated response, has been used. The derivatives of the objective function with respect to the parameters have been obtained numerically.

5.0 RESULTS AND DISCUSSION

In this chapter, the parameters of semi-pervious stream and aquifer system have been determined and discussed. Section 5.1 deals with the determination of parameters with synthetic data and in section 5.2, the parameters have been determined using published data of glacial-outwash aquifer near Cartland, NewYork (Reynolds, 1987).

5.1 PARAMETER DETERMINATION WITH SYNTHETIC DATA:

Synthetic data for the piezometric head in a homogeneous and isotropic infinite aquifer at a distance of 100 m (x=100m) from the stream bank were obtained using equations (4.6) & (4.7) for an asymmetric flood wave in the stream with known parameters (R=600m and β=10,000 m^2/hr). The asymmetric flood wave has been assumed to follow the following equation proposed by Cooper and Rorabaugh (1963).

$$H(t)=NH_{max}(1-\cos\omega t)e^{-\delta t} \quad \text{when } t \le t_d n\ t \le t_d$$
$$=0 \quad \text{when } t > t_d \qquad \ldots(5.1)$$

where,

H_{max} =maximum stream-stage at $t=t_c$, (L);

H(t)=stream-stage measured from the initial water level in the stream which is assumed equal to the head in the aquifer at t=0, (L);

t_d =duration of the flood wave in the stream, (T);

ω $=2\pi/t_d$=angular frequency of the flood wave, (T^{-1})

t_c =time to flood peak, (T);

δ $=\omega \cot(\omega t_c/2)$, ($T^{-1}$), and; N $=\exp(\delta t_c)/(1-\cos\omega t_c)$.

The assumed flood wave is shown in fig. 5.1 with t_c=20hr; t_d=50hr; and H_{max}=3m . The piezometric head variation have been obtained for 100 hr. For obtaining the piezometric head in the aquifer at the end of time t, a time-step size equal to t/50 has been used.

The variation of piezometric-head in the aquifer corresponding to the assumed flood-wave in the stream is calculated considering the stream to be fully penetrating and is shown in fig. 5.2 along with the flood wave. considering this variation of piezometric head in the aquifer and the flood wave as known, the aquifer parameter β were determined using Marquardt algorithm with different initial guesses of β and the results have been presented in table 5.1.

Table 5.1

Result of Optimization of β for Fully Penetrating Stream

Sl. No.	Initial guess of β (sq.m/hr)	Optimized Value of β (sq.m/hr)	Objective Function sq. m	No. of Iteration for Opt.
1.	5	10000.00	0.1051E-19	13
2.	5000	10000.00	0.2389E-19	6
3.	500000	10000.00	0.8783E-14	11
4.	5000000	9999.97	0.1368E-10	14

Table 5.1 shows that the convergence is fast even with poor initial guesses of β. The objective function, i.e., the sum of squared deviation between observed and computed heads in the aquifer, is close to zero even with initial guess of β as 5000000 m^2/hr and it took only 14 iteration to converge to optimal value.

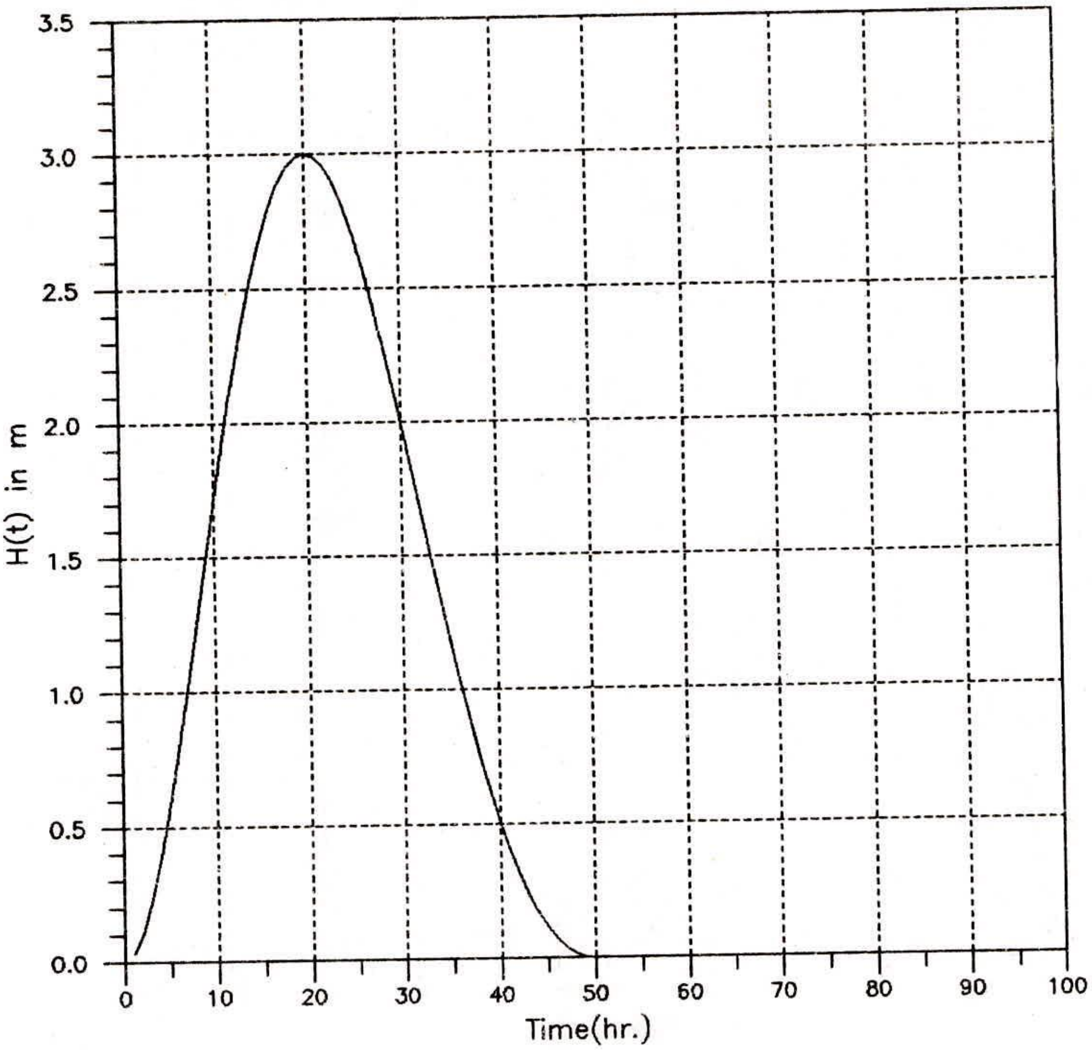


FIG.5.1 A Hypothetical Asymmetric Flood Wave

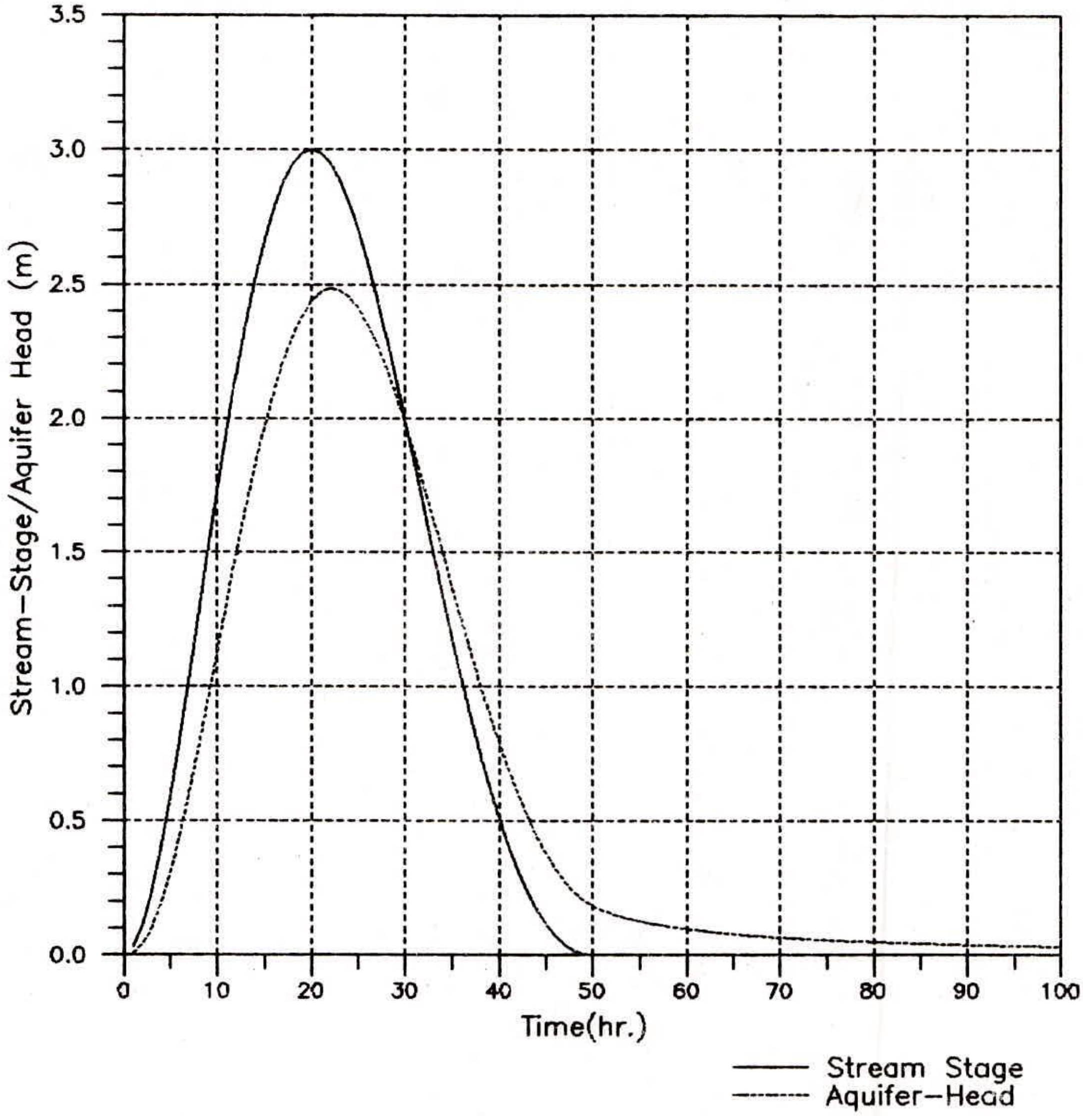


FIG.5.2 Hypothetical Flood Wave and Corresponding Aquifer-Head at x=100m (R=0).

With the above data and considering the stream to be semi-pervious, the result of optimization of the parameters R and β is given in table 5.2.

Table 5.2

Result of Optimization of R & β for semi-pervious Stream

Sl. No.	Initial guesses R (m)	β (sq.m/hr)	Optimized values R (m)	β (sq.m/hr)	Objective Function	No. of Itera-tions
1.	0.1	10	600.0	10000.0	0.4104E-19	18
2.	1.0	1000	600.0	10000.0	0.8772E-21	8
3.	100.0	1000	600.0	10000.0	0.1648E-20	8
4.	6000.0	10000	600.0	10000.0	0.1150E-13	20

The flood wave and the corresponding piezometric head in the aquifer is shown in fig. 5.3. It is observed from Table 5.2 that Marquardt algorithm is efficient in optimizing parameters R and β of the present problem. The synthetic data presented in this section are free from any error. The next section deals with the optimization of parameters with actual field data that may contain observational errors.

5.1 PARAMETER DETERMINATION WITH REYNOLDS'(1987) DATA:

The parameters R and β were optimized using the data reported by Reynolds(1987). He obtained β at three sites in a glacial outwash valley aquifer near Cartland, NewYork. Only at one site, (site 1; Elm Street) stream-stage was recorded. At all the three sites, piezometric-head records at two observation wells are

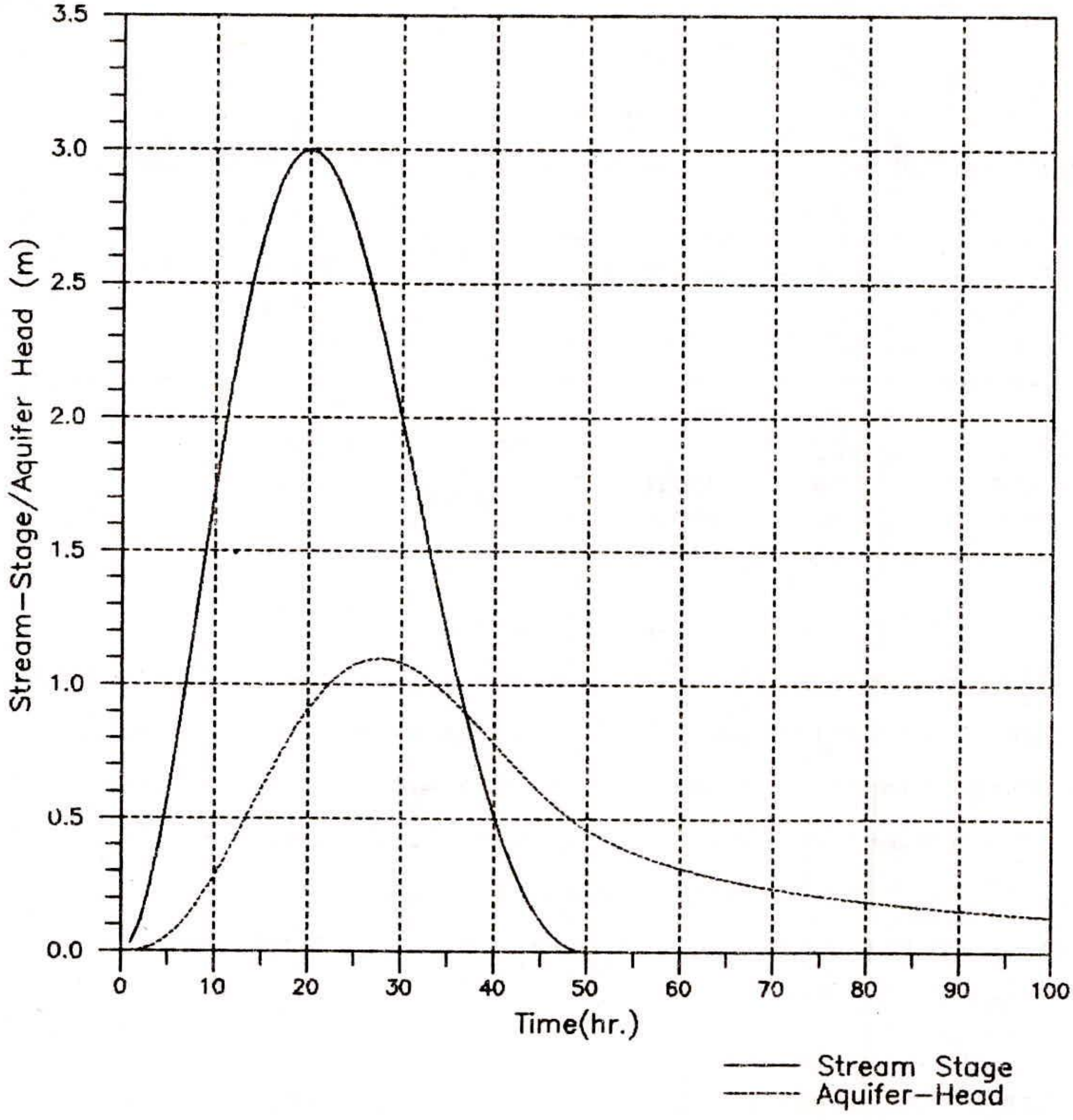


FIG.5.3 Hypothetical Flood Wave and Corresponding Aquifer-Head at x=100m (R=600m).

available. Out of the two observation well at each site, one is very close to the stream-bank (less than 30 ft from the stream bank) and other is 200 ft to 500 ft away from the stream bank.

Reynolds has not used the stream-stage record in his analysis. He calculated β at all the three sites assuming water level in near well (well that is close to the stream bank) as stream-stage. For this he gave the following reasons.

1. The substitution of ground-water levels for stream-stage is required so that the relationship between near well and the distant well (well that is 200 ft to 500 ft away from the stream bank) more closely approximates the assumed theoretical condition for fully penetrating stream;

2. Substitution of stream-bank well for the stream also eliminates the effects of the possibly lower hydraulic conductivity of the stream bed.

The first condition stated above is a conclusion based on the matching of the data with no physical basis. In true sense, the theory of ground-water flow does not conform to such conclusion. The fact contained in second condition highlights the importance of partial penetration and semi-perviousness of stream-bed but the way in which this has been accounted for needs improvement with proper conceptual and physical basis. Therefore, it is not justified to substitute the water level in the stream-bank well for the stream stage. Instead, the stream should be considered as semi-pervious. The present methodology is applicable for semi-pervious stream. Since, the stream-stage is available only at site 1, R and β can be determined only at this site.

The optimized value of parameters were obtained using stream-stage records at site 1 and corresponding ground-water table variations at near well and distant well separately. Distance between the two wells is 152.4m. Exact distance of near well from the stream bank is not reported but it is given that this distance is less than 30ft. Hence, for the present study it has been assumed as 9.0m. Fig. 5.4 shows the digitized stream-stage hydrograph and hydrograph of water table at the two well as reported by Reynolds(1987).

Using the data at distant well the values of the parameters were obtained as R=544.1m and β=13625.0 m^2/hr; whereas with the water table variation at near well, these values were found to be R=105.9m and β=11290.0 m^2/hr. The value of β at near well is less than that at distant well because the thickness of glacial outwash aquifer is less at near well and more at distant well while the present method assumes constant transmissivity. The less value of R at near well as compared to that at distant well shows that the flow lines may be curved at near well and there exists a part of the resistance to flow due to curvature of flow lines and due to semi-pervious bed of the stream between near well and the distant well. Therefore, the values of parameters obtained at distant well may be considered representative for site 1. Fig. 5.5 and fig. 5.6 show the observed and calculated (with optimized parameters) hydrograph of water table at near well and at the distant well respectively.

Reynolds(1987) has obtained value of β=2194.0m^2/hr at distant well for site 1 . He also obtained the specific yield (=0.034) taking maximum saturated aquifer-thickness as 17.68m (58ft) and hydraulic conductivity as 4.19 m/hr (330ft/d). The value of specific yield using present value of β comes out to be

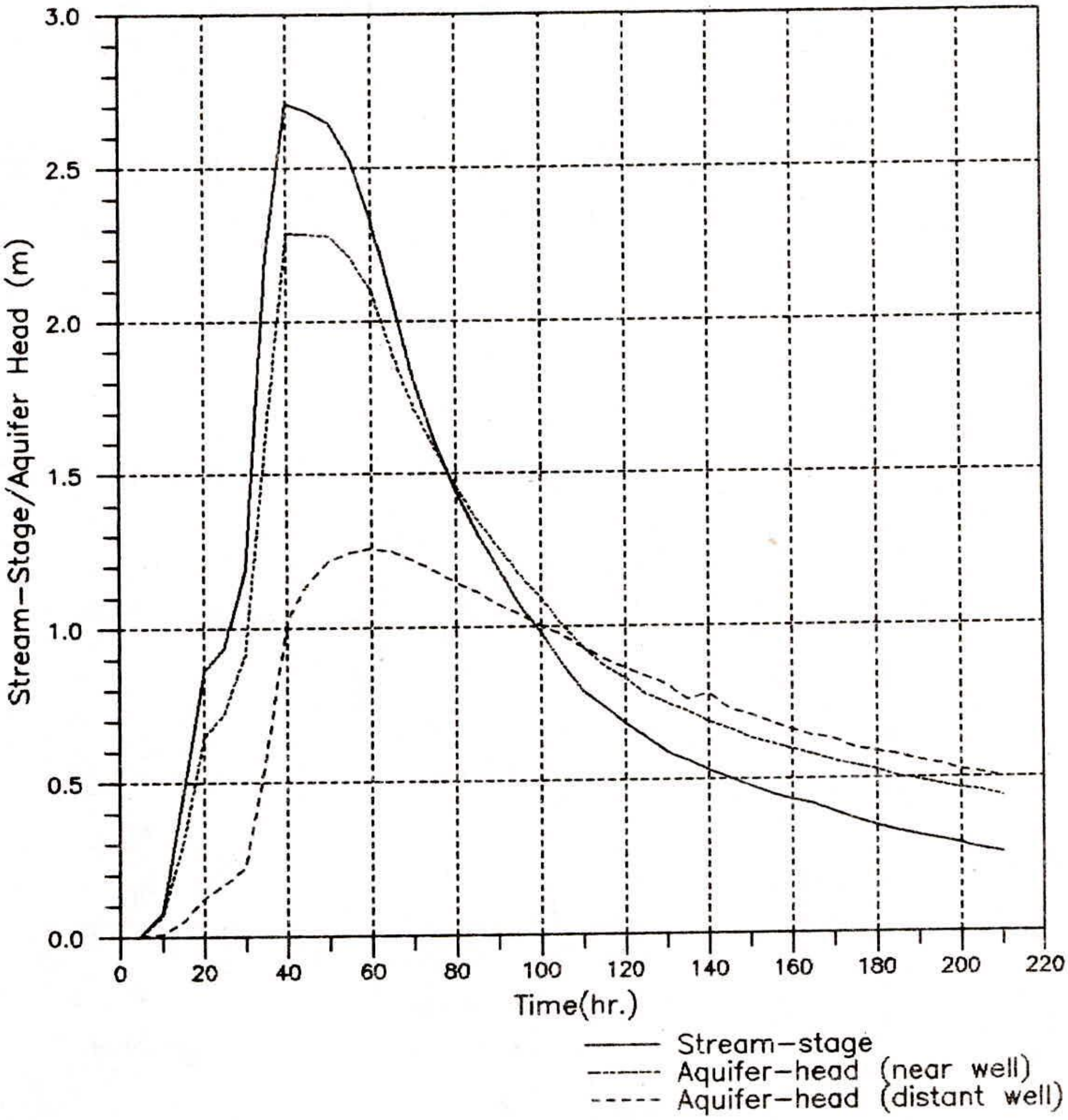


FIG.5.4 Hrdrographs of Stream-stage and Water-tables (Reynolds, 1987)

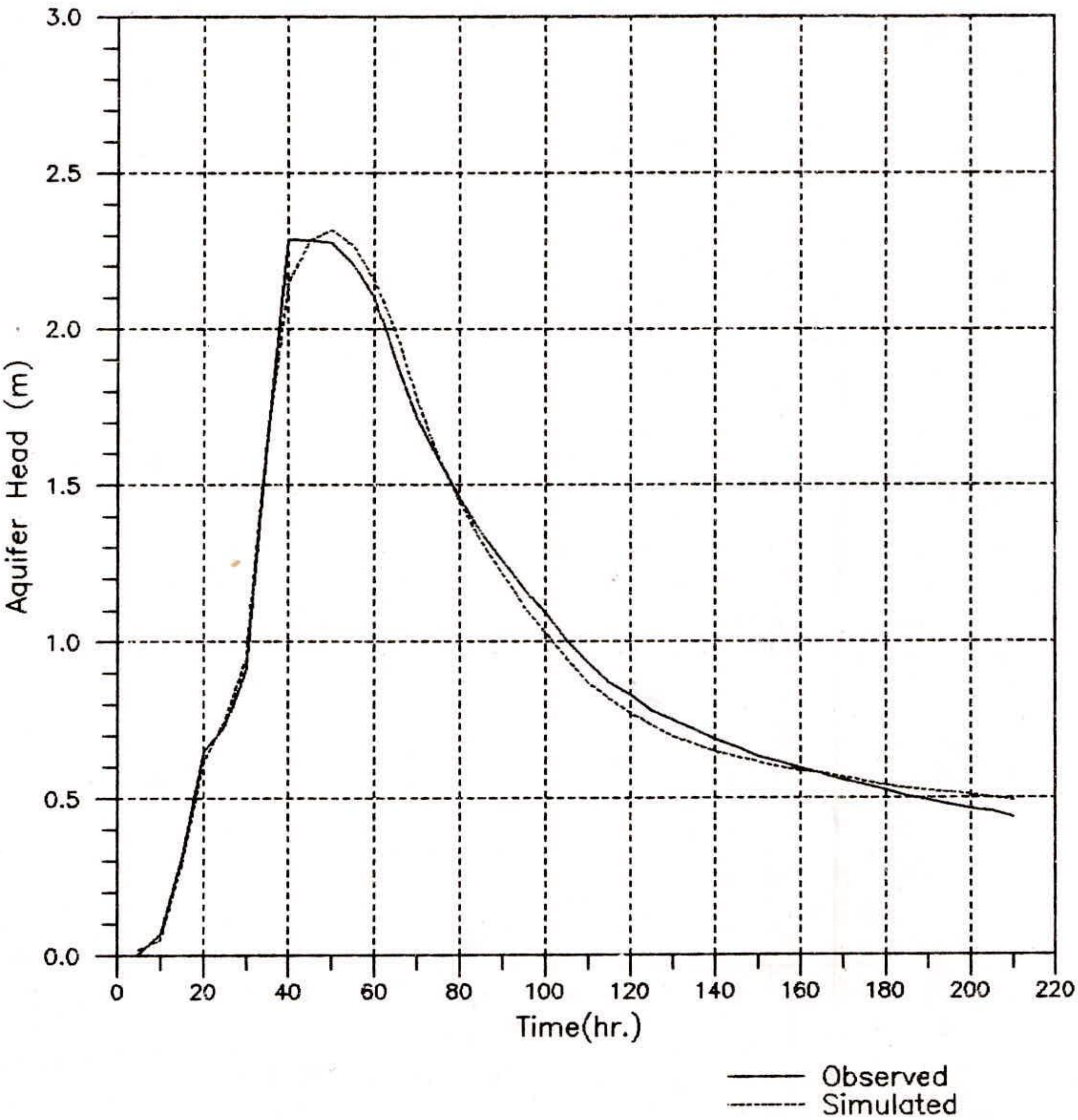


FIG.5.5 Observed and Simulated Hydrographs of Water-table at Near-Well.

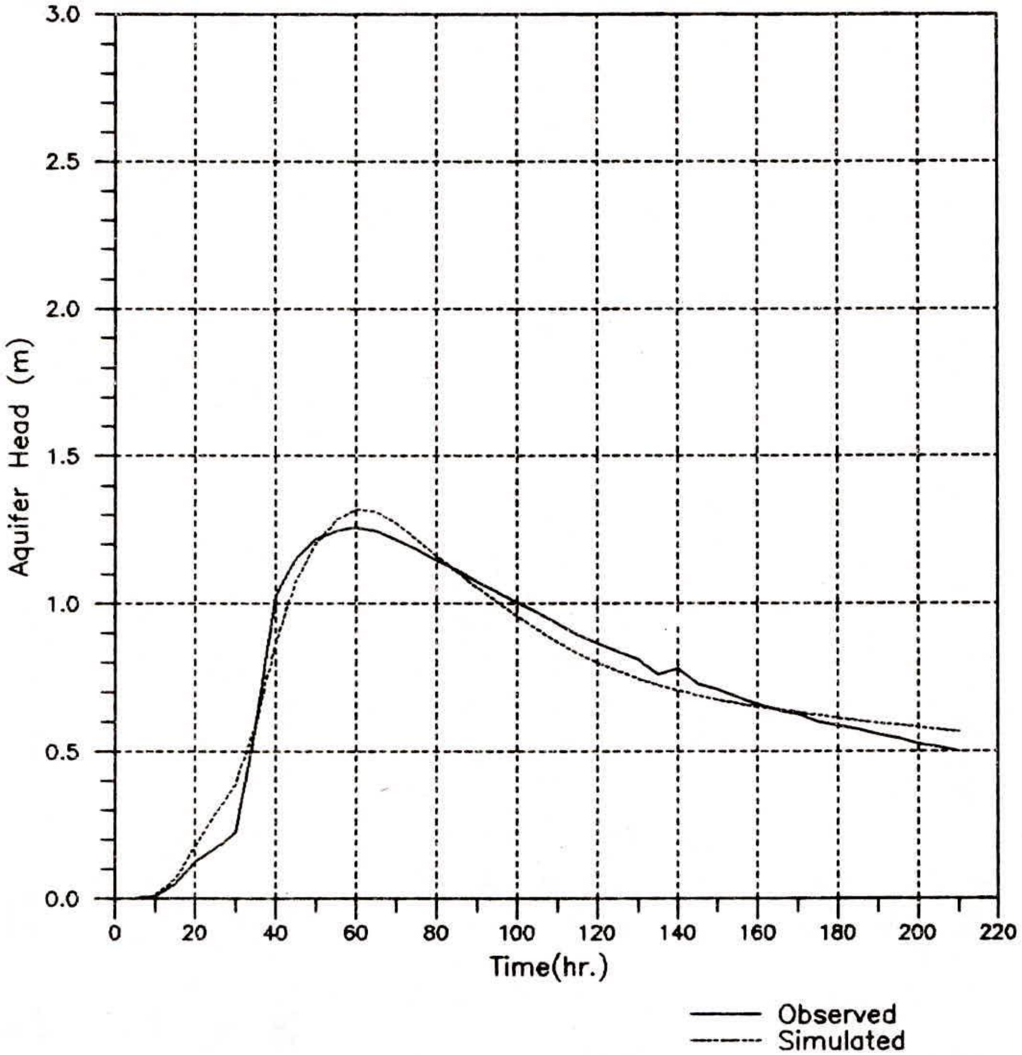


FIG.5.6 Observed and Simulated Hydrographs of Water–table at Distant–well.

0.0054. The simulated water tables at distant well by present method and that obtained by Reynolds(1987) is shown in fig. 5.7.

It is observed from fig. 5.7 that the simulated hydrograph is more close to the observed one as compared to that obtained in Reynolds' simulation.

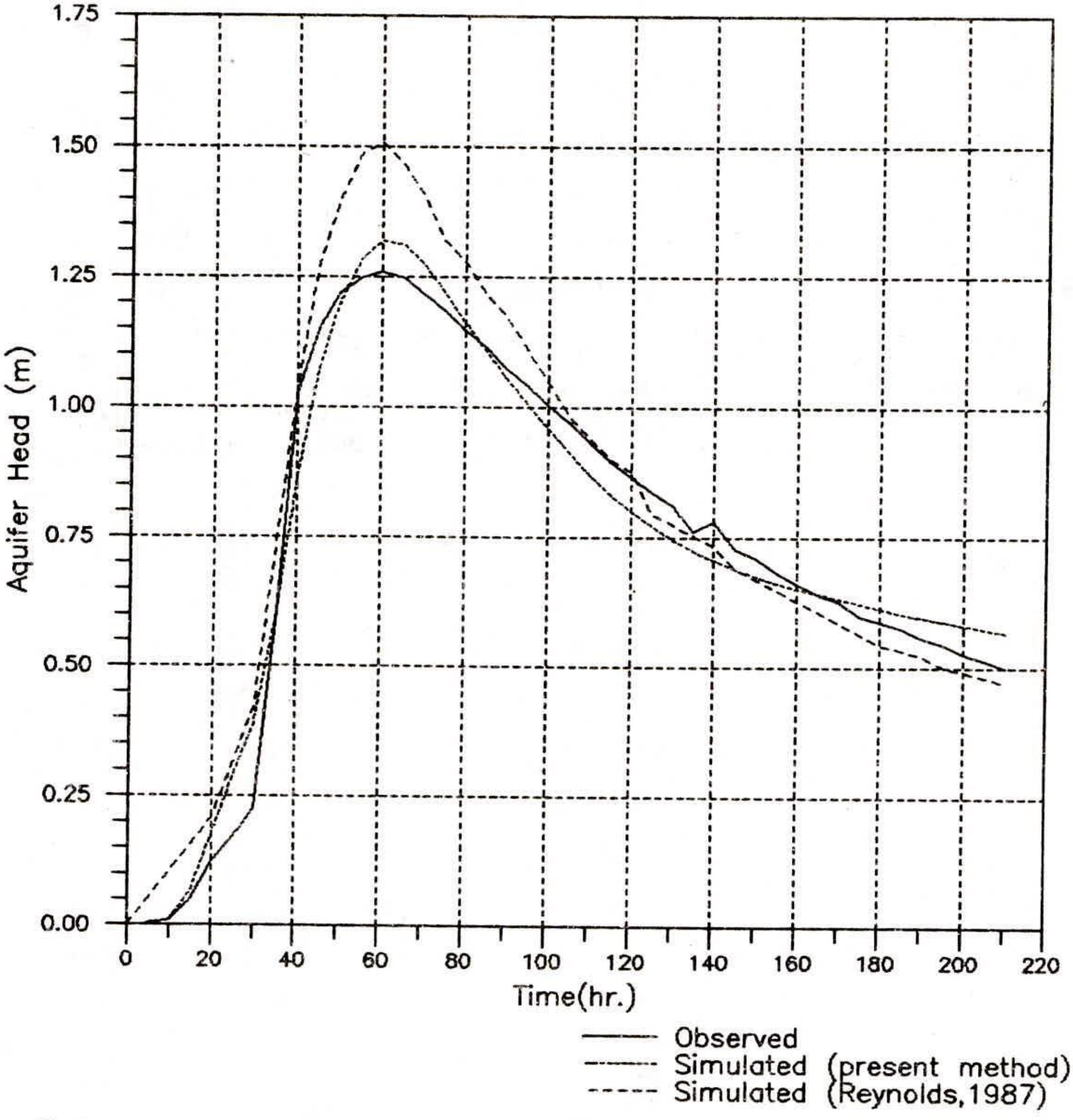


FIG.5.7 Comparison of Simulated Hydrograph at Distant-well with Reynolds' Simulation.

6.0 CONCLUSION

Analytical expression for the induced piezometric head variations in an infinite aquifer on account of stage variations in a semi-pervious stream has been derived. Resistance to flow offered by the semi-pervious bed of the stream has been taken into account. Using a nonlinear optimization technique(Marquardt, 1987) and the derived solution of direct problem, a model has been developed for determination of the parameters (aquifer diffusivity and stream-resistance) for known hydrographs of stream-stage and induced piezometric head in the aquifer. The model can account for any variation in stream-stage.

Applicability of the model has been verified with published data (Reynolds, 1987) and it has been observed that the present method predicts aquifer response more correctly as compared to that as obtained by Reynolds. This shows that the present model is applicable to similar field problems where the stream is partially penetrating and has semi-perviousness at its bed and bank.

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