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Application of Three Aquifer Test Methods for Estimating Hydraulic Properties Within the 100-N Area
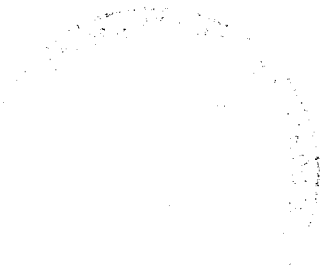
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Executive Summary

This study was conducted for the U.S. Department of Energy by Pacific Northwest Laboratory at the request of the Westinghouse Hanford Company, Facilities Operations Division. The purpose of the study was to better define the range of saturated horizontal hydraulic conductivities in the 100-N Area of the Hanford Site in southeastern Washington for use in a numerical groundwater model.

Three methods were used for determining aquifer properties and are discussed within this report 1) reanalysis of past pumping test data using a pressure derivative method to identify the data in the radial flow regime for analysis by traditional graphical techniques, 2) sinusoidal analysis techniques described in Ferris (1952, 1963) that utilize water-table responses to river-level variations, and 3) the basic flow equation for groundwater.

Based on the results obtained from these three methods, the 100-N Area was divided into two areas with higher hydraulic conductivities in the northeast and lower hydraulic conductivities in the southwest.

A best estimate range of hydraulic conductivity values may be obtained by using the Ferris (composite well/time-lag analysis) and aquifer test reanalysis methods to form the upper bound of the range, and the flow equation and Ferris (composite well/apparent tidal efficiency analysis) method for the lower bound of the range. The hydraulic conductivities in the southwest region of 100-N Area range from 36 to 73 ft/d and in the northeast region 280 to 325 ft/d. The range of hydraulic conductivity presented in this report has narrowed the previous range (104 ft/d to 8,400 ft/d) for the 100-N Area that was based on previous pumping test results.

Acknowledgments

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1.0 Introduction

To predict the flux of strontium-90 entering the Columbia River, the Westinghouse Hanford Company developed a numerical simulation of ^{90}Sr transport for the 100-N Area on the Hanford Site in southeastern Washington using the PORFLO-3 code (Connelly et al. 1991). One of the input parameters for the numerical model, saturated horizontal conductivity, was based on aquifer pumping test data from wells N-27, N-32, and N-34 (Connelly et al. 1991). The aquifer test results yielded a wide range of hydraulic conductivities from 129 to 577 ft/d (39.3 to 176 m/d). Following calibration of the model, a value of 220 ft/d (67 m/d) was used.

In order to increase confidence in the numerical model results, additional hydraulic characterization of the system in the 100-N Area was needed. Three methods to determine aquifer parameters were used to determine which might provide additional information for narrowing the current range of values and providing verification of the input parameters for the numerical model.

The reliability of a groundwater flow model's results rests largely on the knowledge of the hydraulic characteristics of the aquifer. A widely used method for determining aquifer parameters is the constant-rate pumping test. A pumping test consists of pumping groundwater from a well at a constant rate and monitoring the effects on groundwater levels. These effects are analyzed by comparison to theoretical flow equations, often using manual graphical analysis to determine aquifer parameters, primarily transmissivity (T) and storage coefficient (S). Hydraulic conductivity (K) can then be determined from T by dividing by the aquifer thickness.

In addition to the standard pumping tests, there are several analytical methods for estimating aquifer diffusivity (the ratio of transmissivity to the storage coefficient) from the aquifer water-level responses to fluctuation in nearby surface-water bodies (i.e., rivers, oceans, etc.) Published methods of determining aquifer characteristics from groundwater-level response to river or tide fluctuations rely on analytical or semi-analytical solutions to simplified hydraulic equations. All of the methods assume that the aquifer is semi-infinite (bounded on one side by the fluctuating water body) and has no mean gradient. The Ferris (1952, 1963) methods assume purely sinusoidal fluctuations, as do Gregg (1966), Carr and Van der Kamp (1969), and Erskine (1992). Cooper and Rorabaugh (1963) derived a solution for a damped sinusoid to better represent river-level changes during a flood event, and Pinder et al. (1969) provide a semi-analytical method for summing the instantaneous response to small changes in river level, which allows accurate representation of arbitrary river-level changes. Other methods for evaluating aquifer properties from well response to flood-wave events have been presented by Rowe (1960) as modified by Hantush (1961) and Reynolds (1987).

The study described in this report used three methods to determine aquifer properties: 1) analysis of pumping-test data using pressure-derivative diagnostics, 2) evaluation of river-induced groundwater changes with the Ferris (1963) method, and 3) evaluation of river-induced changes using a flow equation method that numerically solves the one-dimensional continuity equation. Data for both water- and river-level were collected from a network of groundwater monitoring wells at the 100-N Area (Figure 1.1) and are reported in Gilmore et al. (1989, 1990).

The first method used was the reanalysis of data from past aquifer pumping tests to determine transmissivity values. There have been at least 13 aquifer pumping tests run in the 100-N Area. The results of these pumping tests were reviewed. Tests that were determined to be most representative were reanalyzed using pressure derivative diagnostic plots that can identify which data are in the radial flow regime and can be analyzed by additional graphical techniques to determine transmissivity values. The results of these analysis are presented in Section 2.0. Two analytical methods using the groundwater response to river-level changes were applied to the 100-N data for both a comparison with the pumping test data and to provide for a spatially weighted hydraulic characterization. The first method uses the analytical method described in Ferris (1952, 1963) (Section 3.0) and the second method uses the continuity equation for flow (Bear 1979, p. 113) (Section 4.0). A discussion (Section 5.0) follows the descriptions of each of the three analysis techniques.
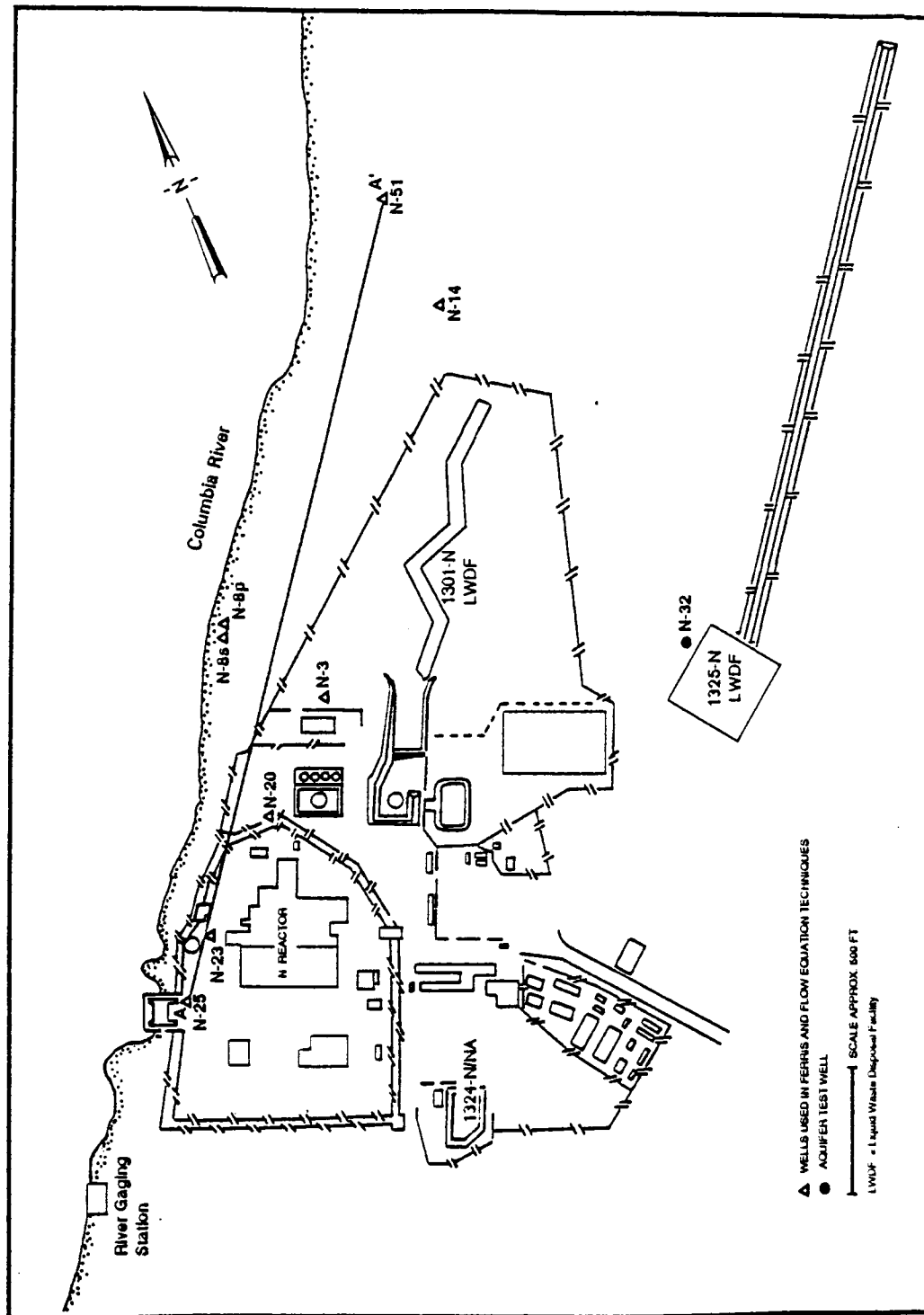
Figure 1.1. Location Map of Monitoring Wells

Columbia River

River Gaging Station

N REACTOR

N-86 N-8p

N-85

N-3

N-20

N-23

N-25

N-51 A'

N-14

1301-N LWDF

1325-N LWDF

N-32

1324-N/NA

△ WELLS USED IN FERHS AND FLOW EQUATION TECHNIQUES

● AQUIFER TEST WELL

SCALE APPROX 500 FT

LWDF = Liquid Waste Disposal Facility

1.3

2.0 Aquifer Test Analysis

2.1 Background

Estimates of hydraulic properties of the unconfined aquifer in the 100-N Area are reported in Kipp and Mudd (1973) and PNL (1988). Estimates of transmissivity ranged between 5200 ft²/d to 420,000 ft²/d and estimates of hydraulic conductivity ranged between 92 ft/d and 5300 ft/d, varying by a factor of approximately 50. These estimates were calculated using data from tests conducted in wells 199-N-14, 199-N-15, 199-N-27, 199-N-32, 199-N-34, 199-N-39, 199-N-64, 199-N-67, 199-N-69, and 199-N-70. The data from wells 199-N-69 and 199-N-70 were taken from an unreleased 1989 report by PNL researchers T. J Gilmore, S. M. Goodwin, and D. R. Newcomer.

These estimates were derived by applying log-log type curve matching methods (e.g., Theis 1935) or the Jacob (1946) straight-line solution to pumping-test data. These tests ranged in duration from a few hours to 24 hours of pumping and the discharge rates ranged from a few gallons per minute to as much as 153 gpm. These variations of discharge rates and test duration are due in part to the variability of hydraulic properties within the aquifer, but are also reflective of lack of consistency in pumping test design.

2.2 Methods

A cursory review of previous hydraulic test analysis of well tests conducted in the area indicates that the majority of the reported property estimates (i.e., T and K) are not strictly valid. The lack of validity in the previous analyses is attributed to a variety of adverse test conditions that were evident in the reviewed test data, as well as reliance on invalid analytical assumptions that were applied in the data analysis. These adverse test conditions and assumptions include the:

- nearby presence of the Columbia River that acts as a recharge boundary during hydraulic testing

- effects of waste-water disposal at nearby facilities (e.g., 1325-N and 1301-N cribs) and the surface disposal of groundwater pumped during constant-rate pumping tests

- assumption that Theisian test conditions held during testing (e.g., fully penetrating well, homogeneous confined aquifer of infinite extent, etc.)

- short-duration of many of the tests that did not extend into late-time radial flow conditions

- the lack of test data corrections for unconfined aquifer dewatering and partial penetration effects, and

- misapplication of log-log and semi-log, straight-line analysis solutions for test data not displaying radial flow conditions.

The following briefly highlights several sources of error associated with previous hydraulic test analysis within 100-N Area wells.

Hydraulic analysis of constant-rate pumping tests are usually performed using log-log type-curve matching and/or semi-log, straight-line methods. During the reanalysis of selected pumping test results, a number of analysis errors were identified. For example, hydraulic conductivity estimates provided in Kipp and Mudd (1973) from single-well tests conducted in wells 199-N-14 and 199-N-15 are based on log-log, type-curve analysis of pumping test drawdown and recovery data. The log-log, type-curve matching method, however, is only strictly correct for analysis of observation well test response (i.e., multiple-well interference tests). Quantitative log-log analysis of the test response at the pumping well is not completely valid, since a number of factors contribute to drawdown and recovery that are not normally accounted for in the analytical procedure. These nonformational response phenomena that affect water-level response at the pumping well include wellbore storage, formation damage or stimulation due to drilling/well construction practices, pumping variations, partial aquifer penetration, etc. (e.g., Weeks 1979; Earlougher 1977).

Another identified cause for the wide range in cited hydraulic conductivity values is the inappropriate application of straight-line analysis. The use of the Jacob (1946) straight-line analysis for pumping test data is strictly valid only for test data exhibiting radial flow conditions and for test times that satisfy the prescribed time, t, criteria specified in the following relationship:

$$t \geq (r^2 S)/(4 T \mu) \tag{1}$$

where r = effective well radius (for pumping well analysis)
S = storativity
T = transmissivity
μ = dimensionless parameter equal to 0.01.

In reviewing the original data sets and analysis by various workers in the 100-N Area, a number of misapplications of the straight-line analysis method were identified for test data not exhibiting radial flow conditions or satisfying the μ criteria relationship. Recent developments in log-log diagnostic analysis have greatly facilitated the identification of the test data that can properly be analyzed using straight-line analysis methods. These recently developed diagnostic methods rely on converting pumping drawdown and recovery pressure data to derivative form and analyzing the shape of the pressure derivative versus time on a log-log

2.2

diagnostic plot. As noted by others (e.g., Bourdet and Alagoa 1984, Clark and Golg-Racht 1985), pressure derivative diagnostic plots can be used to precisely identify regions within the pumping test data set that are controlled by nonformational (e.g., wellbore storage, skin effects, etc.) and formational factors (homogeneous vs. heterogeneous formation response, radial flow conditions, etc.).

Pumping test data from well 199-N-32 were the only data selected to be reanalyzed for hydraulic property determination. Hydraulic properties could not be estimated for other tests in the 100-N Area because of the reasons previously cited.

A 24-hr constant-rate pumping test was conducted at well 199-N-32 in March 1984, shortly after well completion. Discharge rates varied during the course of the test, and averaged approximately 100 gpm. Data records indicate (i.e., as suggested by recoverying water levels) that the well was still developing during the later stages of the pumping test. Because of the discharge rate variability and presence of well development during the test, drawdown test data were not reanalyzed. Recovery test data for well 199-N-32 following termination of pumping, however, were reanalyzed utilizing the methods described in Spane (1992). Briefly described, the analysis procedure included the following steps:

- Diagnostic analysis of the recovery water-level response using a log-log pressure change and pressure change derivative plot.

- Quantitative semi-log analysis of the radial flow portion of the recovery water-level data record, as indicated from the log-log pressure derivative plot.

Recovery data were corrected for unconfined aquifer de-watering using the procedure presented in Jacob (1963). Partial penetration effects were evaluated with applicable software presented in Walton (1987). The analysis indicated that partial-penetration effects were relatively constant during the radial flow period of the test data and, therefore, of no significant consequence to the straight-line analysis of test data exhibiting radial flow conditions. Results of the straight-line analysis indicate a transmissivity of 5,640 ft. Details of the test and reanalysis, including test plots, are presented in Appendix A.

3.0 Ferris Method for Aquifer Parameter Estimation

This section presents the estimates of hydraulic diffusivity for the unconfined aquifer using the analytical method described in Ferris (1952, 1963). The use of sinusoidal analysis techniques is detailed for analysis of river-induced well hydrograph responses for selected unconfined aquifer wells in the 100-N Area. Wells exhibiting discernable river-induced sinusoidal responses are shown on the map in Figure 1.1. In all, eight wells were identified, which are located at distances between 80 ft and 440 ft from the river. All but two of the wells (N-3 and N-8p) are completed with well screen sections at the top of the unconfined aquifer. Figure 3.1 shows a schematic of the well screen sections within the unconfined aquifer for wells analyzed in this investigation. The unconfined aquifer ranges from approximately 50 to 65 ft thick in the region where the wells are located.

The theoretical basis of the analytical solution of Ferris (1952, 1963) requires that the river-stage response can be successfully approximated by a sinusoidal wave, which dissipates with distance from the rivershore boundary. Ferris (1952, 1963) presents two analytical methods for calculating aquifer diffusivity (the ratio of transmissivity to the storage coefficient) based on correlating observation well response to river-stage variation. The two techniques include the time-lag and stage-ratio (i.e., apparent tidal efficiency) methods. The two equations, which are based on analogous heat flow equations (Carslaw and Jaeger 1959), permit calculation of the change in groundwater level for a given distance from the river and for a given aquifer diffusivity.

The time-lag and stage-ratio data can be analyzed collectively. In this situation, the ratio of the well stage to river stage (stage ratio method) or time-lag (time-lag method) calculated for each observation well site is plotted versus distance to the river. A best-fit straight line is then constructed through the data and the appropriate match-point information used in the analysis equations provided in Ferris (1952, 1963) to calculate the average areal hydraulic diffusivity for the aquifer in this region. If the aquifer storativity (S) is independently known (or can be reliably estimated) then estimates for transmissivity (T) can also be determined by dividing the diffusivity by S.

Implicit assumptions in the Ferris sinusoidal analysis method are that 1) the river fluctuation can be closely approximated by a sinusoidal pattern, 2) the aquifer is homogeneous, 3) the river fully penetrates the aquifer and 4) no significant vertical ground-water flow components are present. To ensure that the first assumption was correct, river-stage data between November 1989 and September 1991 were reviewed to identify a period of cyclic sinusoidal fluctuations. Because
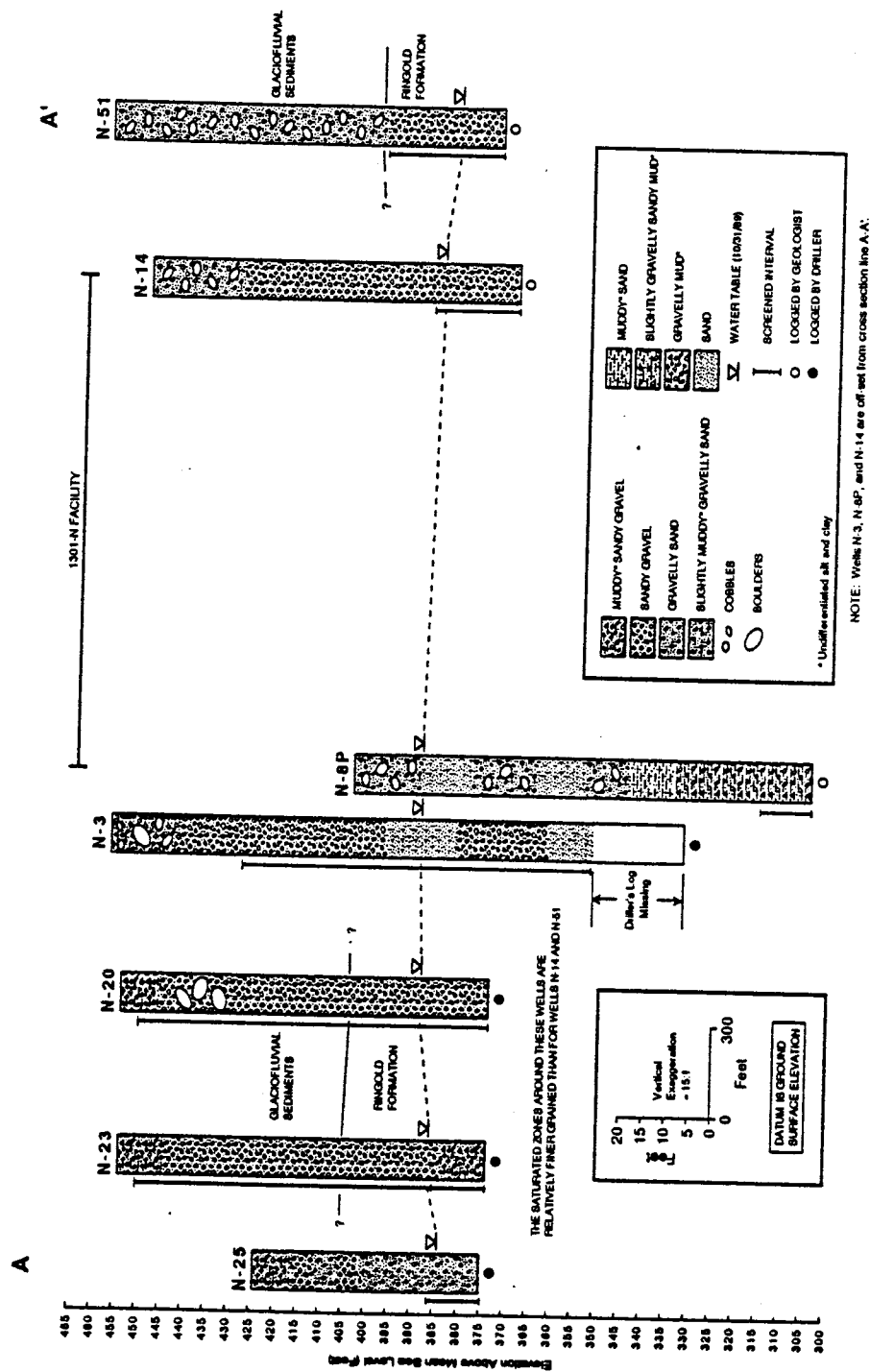
Figure 3.1. Cross Section of Geology Parallel to the Columbia River

of the highly regulated nature of the Columbia River, extended periods of cyclical river-stage fluctuation patterns are not common. Daily cyclical patterns, however, appear to be more frequent between October and December.

Assumption 2 is a common condition for applying most standard hydrologic analytical methods. Because of the nature of the alluvial deposits comprising the unconfined aquifer on the Hanford Site, heterogeneities are expected to occur with increasing size of the area investigated. In an effort to minimize the effects of heterogeneity, wells examined in this study were all located in proximity to the river (i.e., within 450 ft) and, in most cases, completed in the upper-section of the aquifer (see Figure 3.1).

Assumption 3 implies that river and groundwater exchange occurs immediately, with no delay or attenuation of the induced response. In most cases this probably does not occur, and is evident on composite time-lag or observed tidal efficiency vs. distance plots for the wells analyzed. In these situations, Ferris (1963) states that a reference well adjacent to the river may be used for calculating the actual river-stage response transmitted to the aquifer, thereby improving the estimation capabilities. For this investigation, river-stage and reference-well responses were used in the analysis of monitoring well hydrograph patterns.

Assumption 4 is probably valid for situations in which monitor wells that are used in the hydraulic property assessment are not located immediately adjacent to the river and/or completed in a thick aquifer. For this investigation, wells were located between 80 ft and 440 ft from the river and had shallow completions. These conditions would minimize the presence of significant vertical groundwater flow components.

Hourly water-level measurements for the respective wells and the river are generally available for the period December 1989 to February 1991. As indicated, most wells have only limited periods of uninterrupted continuous record. For the analysis, an effort was made to find time periods that were 1) continuous periods without interruption, 2) common with other wells, and 3) reflective of Columbia River stage-elevation data that could be approximated by daily cyclical behavior. Based on these data screening criteria, two time periods (December 30-31, 1989, and October 18-21, 1990) were selected for analysis of well hydrograph response. Figure 3.2 shows an example of the Columbia River fluctuation pattern and associated well hydrograph response for selected wells during these two time periods. As indicated, the well hydrograph responses exhibit a slightly delayed (lagged) and attenuated signal in comparison to the river-stage behavior.

The general analysis procedure consisted of using computer software to shift the river-stage record by specified increments of time (i.e., time-lag) and then statistically analyzing the correlation of the river-stage and well hydrograph records. The correlation method followed the procedure described in Clark
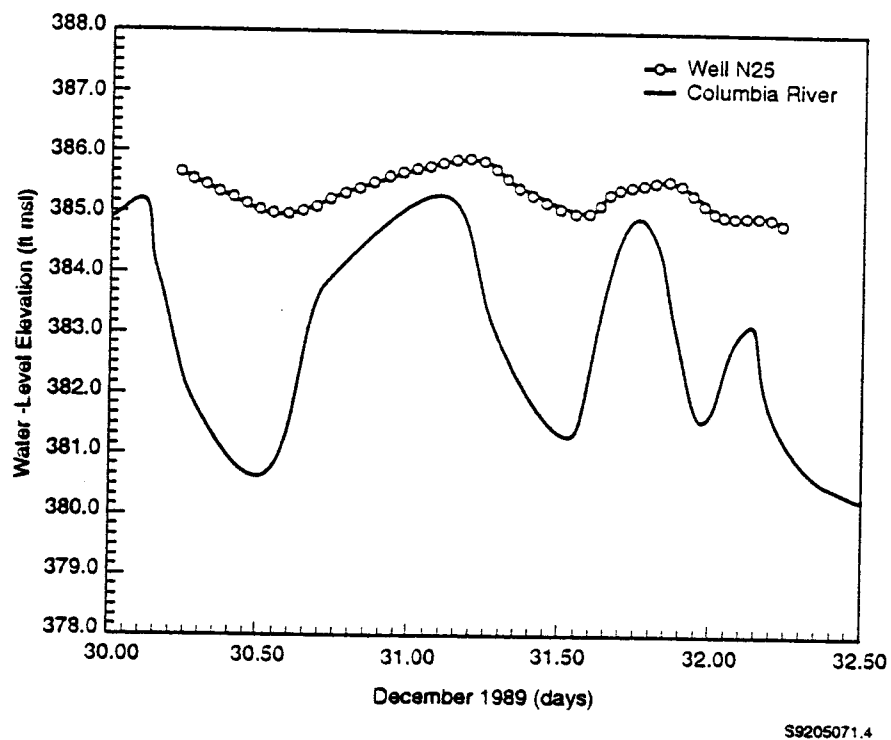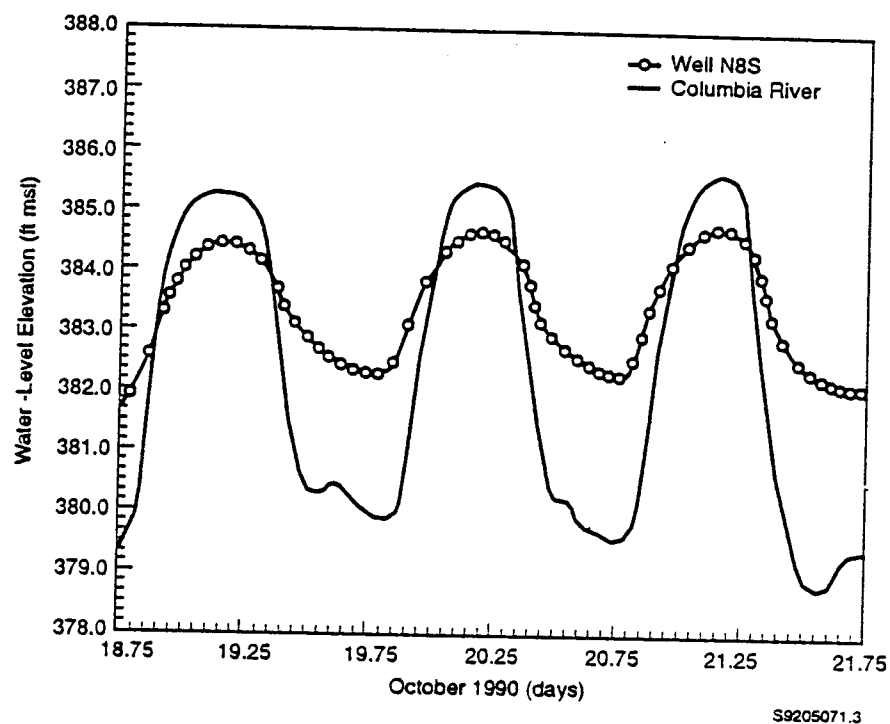
Figure 3.2. Columbia River Fluctuation (a) with Well N-8s Hydrograph and (b) with Well N-25 Hydrograph

(1967), which was originally developed for calculation of the analogous well barometric efficiency factor based on correlated atmospheric pressure and well water-level changes. For this investigation, the Clark (1967) method was used for determining the apparent tidal efficiency factor (TE_a), which was based on relating changes of hourly river-stage elevation with hourly well water-level measurements. The method outlined in Clark (1967) is superior to the peak-ratio method described in Ferris (1963), because it can account for the effects of extraneous trends (such as aquifer recharge) and reduce the effect of individual data errors.

The calculated time-lag for each well analyzed was designated as the value that provided the highest linear correlation coefficient and lowest standard error of estimate for the river-stage and well water-level change correlation. Figure 3.3 shows a final analysis figure for one of the analyzed wells. Included in the figure are results of the statistical analysis for the specified time-lag. As indicated, the apparent tidal efficiency, TE_a, is equal to the slope of the linear regression fit to well water-level changes versus the change in river-stage.
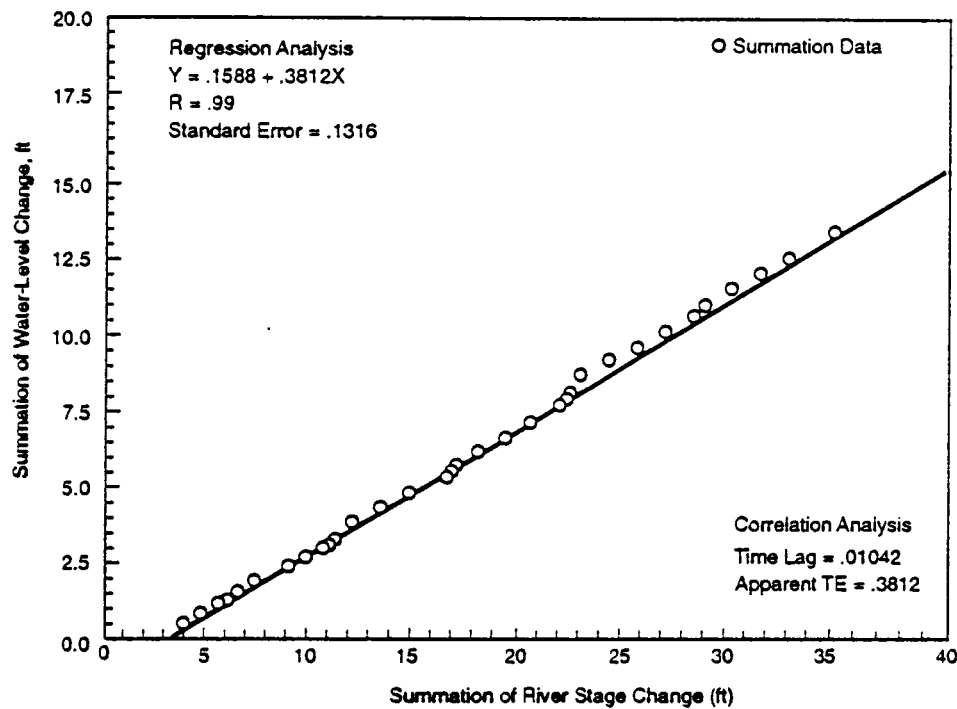
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Figure 3.3. Well Water-Level Changes Versus Change in River-Stage. Slope of Regression Line Equal to Apparent Tidal Efficiency (TE_a).

3.5

After the T_L and TE_a were calculated for each well, the hydraulic diffusivity, T/S, for the intervening aquifer materials between the observation well and the river were determined using the following relationships presented in Erskine (1991):

Time Lag

$$T_L = x\,[(t_o\,S)/4\pi\,T)]^{1/2} \qquad (2)$$

or re-arranging

$$T/S = (x^2\,t_o)/(T_L^2\,4\pi)$$

Apparent Tidal Efficiency

$$TE_a = \exp\,(-x[(\pi\,S)/(t_o\,T)]^{1/2}) \qquad (3)$$

or re-arranging

$$T/S = (x^2\,\pi)/t_o(\ln TE_a)^2$$

where x = distance from river to well location
       t_o = period of river-stage sinusoidal fluctuation.

Time-lag and tidal efficiency calculation results for each well analysis are presented in Table 3.1. Included in the table are estimates for hydraulic diffusivity that are based on both the time-lag method (Equation 2) and apparent tidal efficiency method (Equation 3). Considerable variability exists for the estimates of hydraulic diffusivity, which range over 3 orders of magnitude based on time-lag calculations, but only over 1 order of magnitude from apparent tidal efficiency determinations. Additionally, the wells have been grouped into their spatial location as shown in Figure 3.1. Wells N-3, N-8p, N-8s, N-20, N-23, and N-25 comprise a group of wells located along the southwest section of the A - A' cross-sectional line (see Figure 3.2), while wells N-14 and N-51 are approximately 1700 and 2200 feet, respectively, to the northeast. The northeast pair of wells have generally higher diffusivity values than wells in the southwest group.

Time-lag and tidal efficiency calculation results of Table 3.1 were also analyzed collectively by plotting the calculated time-lag or apparent tidal efficiency for each observation well versus distance to the river and constructing a best-fit straight line through the data. With the appropriate match-point information, Equations 2 and 3 were used to calculate the average areal aquifer diffusivity.

3.6

Table 3.1. Summary of Well Hydrograph/River-Stage Analysis

Well	Distance From River (ft)	Time Lag (day)	Apparent TE	Diffusivity Estimate T/S (ft²/d) Time Lag	Apparent TE
Southwest Group Wells					
N-3	300	.3313	.00705	65,270	11,520
N-8p	80	.006944	.1798	10,560,000	6,830
N-8s	80	.01042	.3812	4,691,000	21,614
N-20	190	.1250	.04543	183,900	11,870
N-23	330	.2854	.02714	106,400	26,290
N-25	130	.06042	.1793	368,400	17,980
Composite Analysis[a]				49,890	20,620
Northeast Group Wells					
N-14	440	.2479	.01943	250,700	39,160
N-51	320	.01736	.1264	27,040,000	75,220

(a) Figure 3.4-3.5.

Figures 3.4 and 3.5 show the composite analysis results for time-lag and apparent tidal efficiency, respectively. The least-squares regression line shown in each figure is representative only of wells comprising the southwest group and, therefore, are not reflective of results obtained for wells N-14 and N-51. Composite analysis was not attempted for the two northeast wells, because of the presence of only two wells for this region. Their results, however, are shown for comparison purposes with wells analyzed for the southwest cross-sectional area. As indicated in Figures 3.4 and 3.5, hydraulic diffusivity estimates for the unconfined aquifer in this region based on composite time-lag and apparent tidal efficiency analysis are 49,890 ft²/d and 20,620 ft²/d, respectively.

Ideally, the least-squares regression line would be expected to pass through the origin. The fact that the regression lines shown in Figures 3.4 and 3.5 have negative y-intercept values suggests that the river "acts" as if it is closer to the wells analyzed. Site inspection of the shoreline in the 100-N Area reveals the possibility that because of manmade modifications (rip-rap emplacement, etc.) the shoreline may be actually closer than surface-distance measurements would indicate (i.e., in some locations 20 to 30 ft closer).

To examine the effects of decreasing the river distance, the well hydrograph data were reanalyzed using well N-8s to simulate the river response. This caused a reduction of 80 ft in the calculated distance for monitoring wells to the "river."
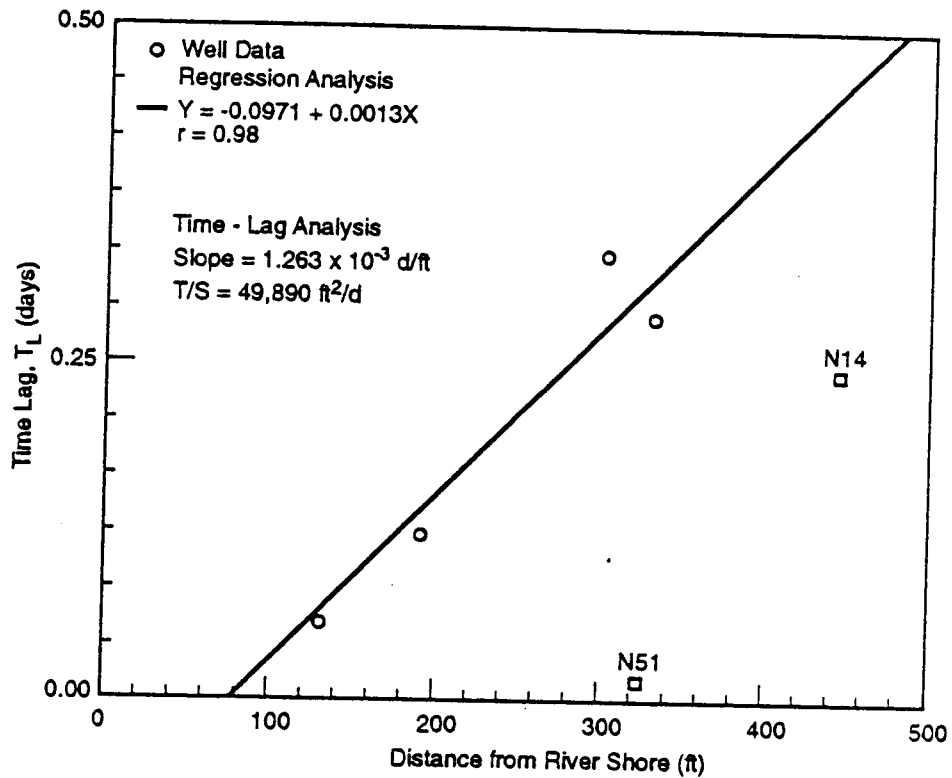
Figure 3.4. Composite Analysis Results for Time Lag Method
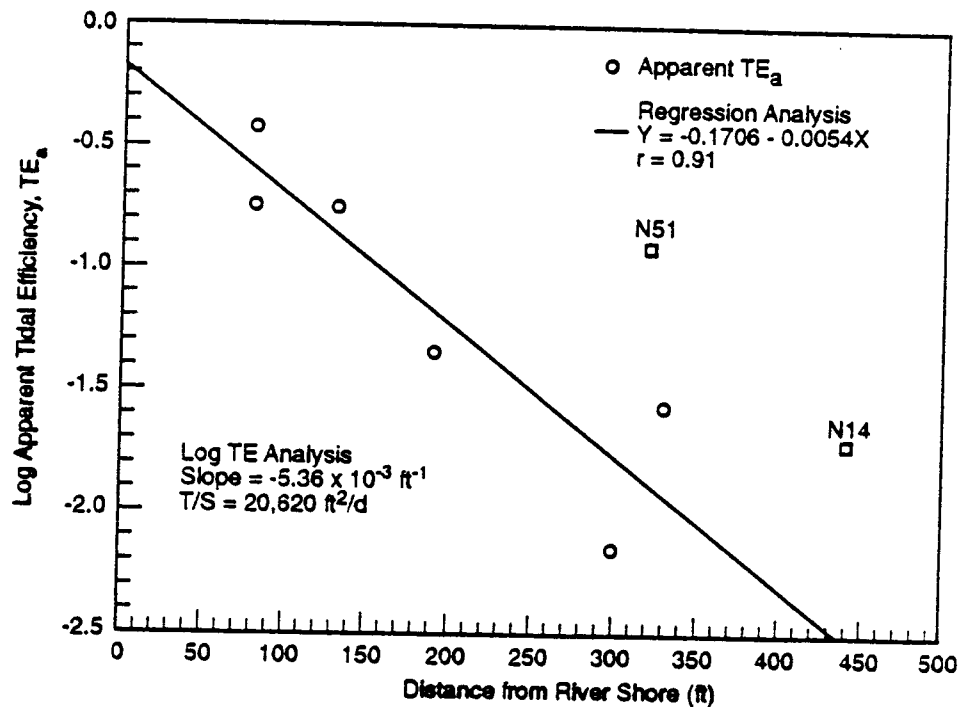
Figure 3.5. Composite Analysis Results for Apparent Tidal Efficiency (TE_a)

3.8

The use of a reference well to simulate river response is also the preferred procedure to use for those situations where the river does not completely penetrate the aquifer (e.g., Reynolds 1987). Results of the reanalysis are presented in Table 3.2. (Note: Wells N-8s and N-8p cannot be analyzed because in this procedure they represent the shoreline). As indicated in Table 3.2, the variability in hydraulic diffusivity estimates was reduced significantly, especially for estimates calculated using the time-lag method. Of particular importance is the more close corroboration for hydraulic diffusivity values obtained with the two methods (i.e., except for well N-51). In most cases, hydraulic diffusivity estimates were within a factor of about four for each well.

Figures 3.6 and 3.7 show the composite well analysis results for the time-lag and apparent tidal efficiency methods, using well N-8s as the point of applied river stress. As previously stated, wells N-14 and N-51 were not used in the composite analysis and are shown only for comparison purposes. Examination of Figures 3.5 and 3.7 indicates that similar average hydraulic diffusivity values were obtained from the composite analysis of apparent tidal efficiency vs. distance (i.e., comparing the river derived values vs. that obtained using well N-8s as the river stress). A slightly lower average value for the southwest region, however, was obtained based on the composite time-lag analysis (i.e., 36,480 ft²/d vs. 49,890 ft²/d), as shown in Figures 3.4 and 3.6.

Table 3.2. Summary of Well Hydrograph/River-Stage (Well N-8s) Analysis

Well	Distance From River (ft[a])	Time Lag (day)	Apparent TE	Diffusivity Estimate T/S (ft²/d)	
				Time Lag	Apparent TE
Southwest Group Wells					
N-3	220	.3604	.01598	29,650	8,890
N-20	110	.1063	.1368	85,290	9,600
N-23	250	.3000	.06649	55,260	26,730
N-25	50	.05958	.4512	56,040	12,400
Composite Analysis[b]				36,480	20,010
Northeast Group Wells					
N-14	360	.2196	.05530	213,900	48,590
N-51	240	.00625	.3310	117,300,000	147,980

(a) Calculated distance to river reduced by 80 ft distance to well N-8s.

(b) Figure 3.6-3.7
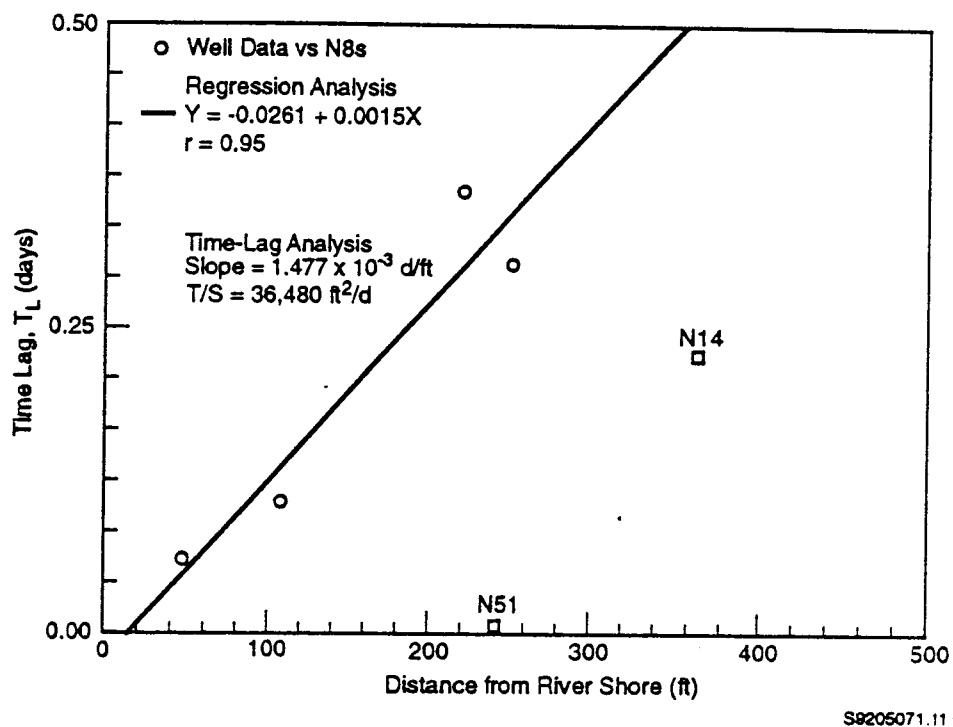
Figure 3.6. Composite Well Analysis Results for Time Lag Method Using Well N-8s as Point of Applied River Stress
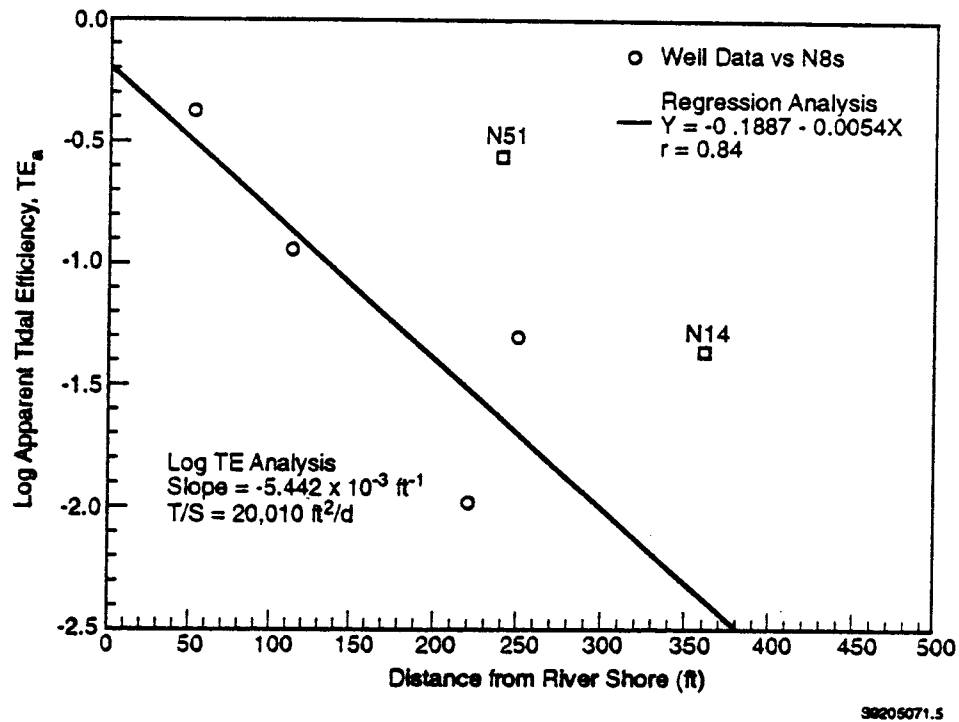
Figure 3.7. Composite Well Analysis Results for Apparent Tidal Efficiency (TE$_a$) Using Well N-8s as Point of Applied River Stress

3.10

In summary based on composite well analysis, a best estimate range for the average hydraulic diffusivity for the unconfined aquifer in the southwest cross-sectional area is 20,010 ft²/d to 49,890 ft²/d. The range in average hydraulic diffusivity varies by a factor of two (i.e., for both composite analyses). The diffusivity for the unconfined aquifer in the vicinity of the northeast group of wells appears to be more than twice that calculated for the region encompassing the southwest area wells.

The Ferris sinusoidal method does not permit calculation of the individual components of hydraulic diffusivity (i.e., transmissivity and storativity). However, if the value for transmissivity of approximately 5,640 ft²/d reported from the reanalysis of well N-32 (see Section 2.0) is representative of the unconfined aquifer in this area, then a storativity range of between 0.11 and 0.28 is indicated. This storativity range estimate is within the range for specific yield commonly reported for the unconfined aquifer on the Hanford Site. [Note: storativity or specific yield for an unconfined aquifer is the yield of the aquifer per unit area and unit drop of the water table (Bear 1979)].

4.0 Aquifer Parameter Estimation Using the Groundwater Flow Equation

This section describes results from the analysis to determine aquifer diffusivity using the basic flow equation for groundwater. As with the Ferris Method (Section 3.0) the purpose of this analysis was to obtain average areal estimates of hydraulic diffusivity for the unconfined aquifer.

The Boussinesq equation is the basic continuity equation for groundwater flow in a phreatic aquifer with a horizontal impervious base. The equation is derived on p. 113 of Bear (1979).

$$\frac{\partial}{\partial x}\left(h\frac{\partial h}{\partial x}\right) + \frac{N}{K} = \frac{S}{K}\frac{\partial h}{\partial t} \tag{4}$$

where h = h(x,t) is the height of the water table
        N = N(x,t) is an external source/sink term
        S = Specific yield of the aquifer
        K = hydraulic conductivity of aquifer.

Equation 4 is nonlinear because of the product $\partial h/\partial x$. Two methods of linearization to facilitate a solution are described in Bear (1979, p. 115). The linearized version of Equation 4 is shown in Equation 5.

$$\frac{\partial^2 h}{\partial x^2} = \frac{S}{T}\frac{\partial h}{\partial t} \tag{5}$$

Note: Diffusivity = kh/S = T/S.

Equation 5 is only valid when changes $\partial h/\partial x$ and $\partial h/\partial t$ are small relative to the total height h. Equation 5 can then be solved using a one-dimensional, fully implicit finite-difference technique (Fletcher 1988). In finite-difference form Equation 5 becomes:

$$-\frac{K\bar{h}}{S}\frac{\Delta t}{(\Delta x)^2}h_{i-1}^{j+1} + \left(1 + 2\frac{K\bar{h}}{S}\frac{\Delta t}{(\Delta x)^2}\right)h_i^{j+1} - \frac{K\bar{h}}{S}\frac{\Delta t}{(\Delta x)^2}h_{i+1}^{j+1} = h_i^{j} \tag{6}$$

where h_i^j = water-table height at the i-th spatial node and the j-th time step
Δx = constant grid spacing
Δt = constant time step.

Equation 6 is a system of linear equations that can be solved for h_i at new time $j+1$, given initial condition h_i at time j and boundary conditions h_1 and h_{imax} at time $j+1$. Because the system of equations forms a tridiagonal matrix, it can be efficiently solved using Thomas' algorithm (e.g., Anderson et al. 1984).

Measured river-level data was used with Equation 6 in an iterative technique to determine aquifer diffusivity by matching predicted changes in groundwater levels with observed changes. A computer program (Appendix B) was used to repeatedly solve Equation 6 and update estimates of diffusivity until a good fit between data and predictions was obtained. The fit was judged by comparing the average absolute change in measured and predicted groundwater elevations for the last week of the simulation. Diffusivity was adjusted until these values agreed to within the specified accuracy (0.1%). The program converged rapidly (usually less than 10 iterations). The numerical solutions were calculated with grid spacing $\Delta x = 50$ ft, and time steps $\Delta t = 800$ or 1600 s using as input linear interpolations of river-level data recorded at 3600-s or 7200-s intervals (1 or 2 hr). The program was validated by comparison with the analytical solution of Pinder et al. (1969).

A sample result of the program is presented in Figure 4.1, which shows the predicted and measured groundwater levels for Well 199-N8s in October 1990. After an initial spin-up period, the predicted changes in groundwater elevation closely resemble the measured changes, both in amplitude and in timing. The constant offset of approximately 1 ft is attributed to errors in the distant-boundary condition and, because the program examines changes in water levels rather than absolute magnitude, the offset has little effect on the diffusivity estimate.

Assumptions of the flow equation method are that 1) the aquifer is phreatic with an impervious base, 2) the river fully penetrates the aquifer and 3) no significant vertical groundwater flow components are present (Dupuit assumption). The first two assumptions are satisfied at the 100-N Area as discussed in Section 3.0. Assumption 3 is probably valid for situations where monitor wells are not located immediately adjacent to the river and/or completed in a thick aquifer. For this investigation, wells were located between 80 ft to 440 ft from the river and have shallow completion intervals suggesting that vertical groundwater flow components were not significant.

4.2
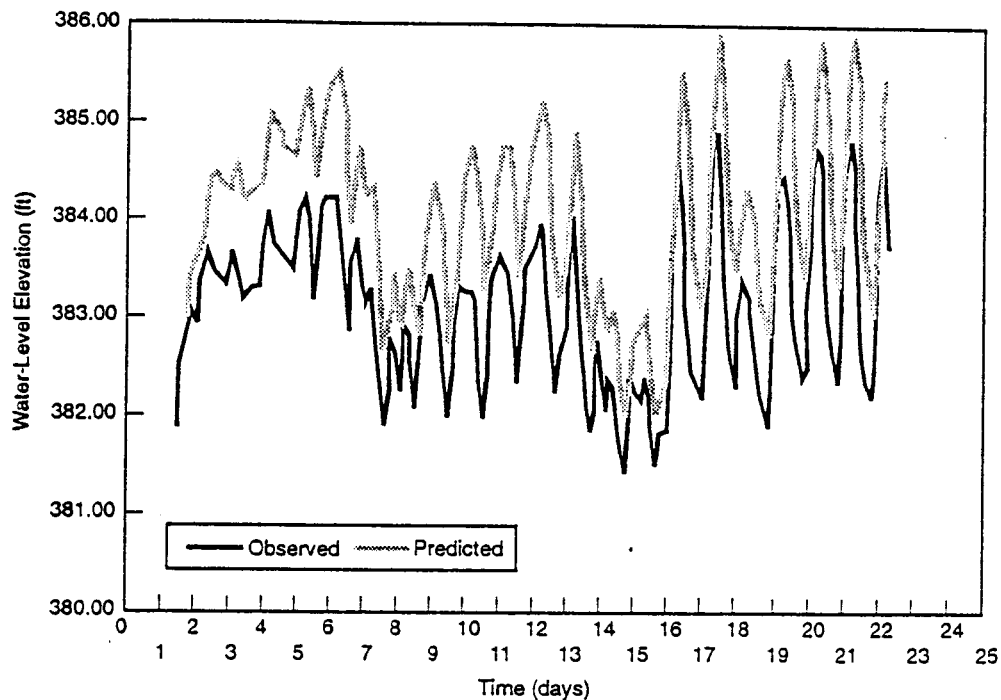
S9205071.6

Figure 4.1. Comparison of Observed and the Predicted Water Levels for
Well N-8s (October 1990)

The original data sets used in this analysis are from Gilmore et al. (1990)
and the length of each set used as input to the computer program was for one
month. In order to compare the diffusivity values of this method with the results
from the Ferris method in Section 3.0, the same wells and periods of analysis
(December 1989 and January 1990) were used. In addition, a period during the
high river stage in June 1990 was used in the flow equation analysis. The June
1990 data could not be used in the Ferris method because the water-level fluctua-
tions during this period were not cyclic and did not approximate a sinusoidal wave.
The calculated diffusivity values, however, from both the high and low river stage
showed no appreciable difference when the flow equation method was used (see
Table 4.1).

The flow equation method provides the aquifer parameter, diffusivity; how-
ever, the numerical model requires the aquifer parameter, horizontal hydraulic con-
ductivity. The hydraulic conductivity can be calculated from diffusivity by knowing
the two variables: specific storage and aquifer thickness. These variables can be
specified in the flow equation computer program input file for the calculation of
hydraulic conductivity.

The calculation of hydraulic conductivity (k) from a diffusivity value is very
sensitive to the value assumed for the specific storage of the aquifer. For example,

4.3

Table 4.1. N Area Flow Equation Method Diffusivity

Well	Period	Flow Equation Method Diffusivity (ft^2/d)
N-8s	Jan-90	32,000 (64)[a]
	Oct-90	10,000 (20)
N-8p	Jan-90	13,500 (27)
	Jun-90	26,500 (53)
	Oct-90	22,000 (44)
N-20	Jan-90	10,000 (20)
	Jun-90	11,000 (22)
	Oct-90	11,000 (22)
N-23	Dec-89	21,500 (43)
	Jan-90	20,500 (41)
N-25	Dec-89	21,500 (43)
	Jan-90	12,500 (25)
N-51	Dec-89	100,500 (201)
	Jan-90	179,500 (359)
	Jun-90	207,500 (415)

(a)   Hydraulic conductivity values (ft^2/d) are in parenthesis and are based on s = 0.1 and aquifer thickness at 50 feet.

a change in the specific storage from 0.1 to 0.2 changes the resultant T and K values by a factor of 2. Specific storage is approximately equal to effective porosity and is defined as yield of the aquifer per unit height for a unit drop in water-table height (Bear 1979, p. 88). To directly measure specific storage, a sediment core sample can be tested for bulk density. However, the specific storage of the aquifer in the 100-N Area is likely to vary both laterally and vertically due to the heterogenous nature of the sediments. Many samples would be needed to adequately define the average specific storage over the study area. In a compilation of hydrologic parameters of the Columbia Basin, Gephart et al. (1979) reported values of storage coefficients for the unconfined aquifer ranging from 0.01 to 0.1. For this report, a conservative specific storage value of 0.1 was assumed. This is a conservative value because it results in a higher groundwater flux.

The aquifer in the 100-N Area is approximately 35 to 60 ft thick. The bottom of the aquifer is formed by a Mud unit within the Ringold Formation composed of 75% mud (undifferentiated silt and clay) at approximately 100 ft below the land surface (350 ft elevation) based on geologic logs of the now abandoned 699-86-60 well under the 1312 LERF facility and well N-5.

Another variable that is required for the flow equation method is distance from the river to the well. Distances were calculated from an aerial photograph of the 100-N Area. The length of the 1325 Crib was used for scale. The digitized locations of the wells and the nearest point to the river were measured and then rounded to the nearest 10 ft. The river shore varies daily and seasonally so the distance from the well to the river is an approximation.

The results of applying this analytical technique to the 100-N data are shown in Table 4.1. The resultant T and K values for each of the wells are very consistent with this method. Although the flow equation method is simplification of flow in the aquifer, the results provide a good approximation of the aquifer properties and compare favorably with the aquifer test reanalysis results (Section 2.0) and Ferris method (Section 3.0).

5.0 Discussion

The values for diffusivity and hydraulic conductivity derived from each of the three methods presented in Sections 2.0 through 4.0 are summarized in Table 5.1. The actual values derived from both the Ferris method (Section 3.0) and the flow equation (Section 4.0) are for diffusivity, but have been converted to hydraulic conductivity for comparison purposes using the relationship for hydraulic diffusivity (Kb/S=hydraulic diffusivity) and assuming a specific storage of 0.1 and an aquifer thickness of 50 ft. Hydraulic conductivity values are presented in parenthesis.

Examination of the diffusivity values of the wells in Table 5.1 indicates that the wells can be divided into two groups based on location, with wells N-14 and N-51 located in the northeast forming one group and all other wells (N-8s, N-8p, N-20, N-23, N-23, N-25, N-32) located in the southwest comprising another distinct group. For both the Ferris method and the flow equation method, wells located in the northeast of the 100-N Area (N-14, N-51), display diffusivity values that are significantly higher than wells comprising the group located in the southwest of the 100-N Area. For the flow equation method, the diffusivity results are an order-of-magnitude different between the two groups. This variation based on location appears to have a hydrologic basis and may be caused by lithologic changes in the aquifer's sediments to the northeast of the 1301-N LWDF (Gilmore et al. 1990).

In comparing results obtained for the three methods, it is evident that the Ferris method (based on time-lag analysis) and aquifer test reanalysis yield the highest values and the flow equation and Ferris method (based on apparent tidal efficiency analysis) the lowest values. These methods were used to estimate the upper and lower limits for a range of diffusivity values (hydraulic conductivity values in parenthesis). Because the southwest and northeast groupings of wells discussed above appear to comprise distinct groups, the two groups will be treated separately.

For the southwest grouping of wells the reanalysis of the aquifer pumping test forms the upper limit of this range at 112,800 ft^2/d (295 ft/d). The lower bounds of this range, 18,000 ft^2/d (36 ft/d), is the average of the wells analyzed using the flow equation method. The composite analysis results of all the wells using the Ferris technique falls in the middle, ranging between 20,010 ft^2/d (40 ft/d) based on the composite/well apparent tidal efficiency analysis, and 36,480 ft^2/d (73 ft/d) - based on the composite/well time-lag analysis (Table 3.2). Although data from the two wells that form the northeast groupings of wells are limited, a range of diffusivity values can be estimated using the results of the two techniques. The average for the two wells using the Ferris technique forms the lower limit of 140,000 ft^2/d (280 ft/d) and the upper limit of 162,500 ft^2/d (325 ft/d) is the average for well N-51 using the flow equation method. (Note: The unrealistic value obtained for well N-51 using the Ferris time-lag method was not included in the lower limit calculation). When these values are converted to

hydraulic conductivity (with s = 0.1 and aquifer thickness = 50 ft), these results compare favorably with the reported ranges of hydraulic conductivities within the Columbia Plateau for the Ringold Formation of 10^1 to 10^2 ft/d (Gephart et al. 1979).

Table 5.1. N Area Diffusivity, Flow Equation and Ferris Method

Well	Period	Flow Equation Method Diffusivity (ft^2/d)	Ferris Method Diffusivity (ft^2/d)	Aquifer Reanalysis Diffusivity (ft^2/d) at S= 0.15-0.05
N-8s	Jan-90	32,000 (64)		
	Oct-90	10,000 (20)	11,500 (23)[a]	
N-8p	Jan-90	13,500 (27)		
	Jun-90	26,500 (53)		
	Oct-90	22,000 (44)	7,000 (14)[a]	
N-3	Oct-90		8,890[b]	
N-14	Dec-89		131,000 (262)[c]	
N-20	Jan-90	10,000 (20)		
	Jun-90	11,000 (22)		
	Oct-90	11,000 (22)	10,000 (20)[b]	
N-23	Dec-89	21,500 (43)	27,000 (54)[b]	
	Jan-90	20,500 (41)		
N-25	Dec-89	21,500 (43)	12,500 (25)[b]	
	Jan-90	12,500 (25)		
N-32				37,600 - 112,800 (295)
N-51	Dec-89	100,500 (201)		
	Jan-90	179,500 (359)		
	Jun-90	207,500 (415)		
	Oct-90		148,000 (296)[b]	

(a) Based only on apparent tidal efficiency using distance to river (Table 3.1).
(b) Based on only apparent tidal efficiency analysis value (Table 3.2).
(c) Average of both apparent tidal efficiency analysis and time-lag analyses (Table 3.2).

Note: Hydraulic conductivity values in parenthesis (ft^2/d).

Results of this study indicate that despite the variability exhibited for the three methods, the range for hydraulic conductivity has been narrowed considerably in comparison to the range initially used for this region based on pumping test results (104 ft/d to 8,400 ft/d). (Note: the range of values used in Connelly et al. (1991) representing aquifer test values from wells N-27, N-32, N-34 was from 129 to 577 ft/d).

This study narrowed the range of aquifer parameter values for the 100-N Area and has identified two distinct areas of different diffusivity values. It was also determined that analytical techniques for determining aquifer properties based on a comparison of well and river stage water-level fluctuations compare favorably with more traditional aquifer pumping tests and provide valid results in the 100-N Area of the Hanford Site.

6.0 References

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Appendix A

Aquifer Test of Well 199-N-32

Appendix A

Well 199-N-32

A 24-hr constant-rate pumping test at well 199-N-32 was initiated on March 26, 1984. Discharge rates varied during the course of the test, and averaged approximately 100 gpm. Observation well 199-N-33, located 515 ft from well 199-N-32, was monitored during the test, but no drawdown was observed.

Available test data records indicate (i.e., as suggested by recovering water levels) that the well 199-N-32, was still developing during the later stages of the pumping test. Because of the discharge rate variability and presence of well development during the discharge test, drawdown test data were not reanalyzed. Recovery test data for well 199-N-32 following termination of pumping, however, were evaluated diagnostically using a log-log pressure change and pressure change derivative plot to identify characteristic flow regimes and conditions exhibited within the recovery data. A quantitative semi-log analysis of the radial flow portion of the recovery water-level data record, as indicated from the log-log pressure derivative plot, was then applied. A detailed description of the use of log-log diagnostic and semi-log straight-line analysis methods is presented in Spane (1992).

Figure A.1 shows that log-log recovery data and recovery data derivative plot following the constant-rate pumping test. The recovery derivative was calculated using the Agarwal time equivalent function following the procedure outlined in Bourdet and Alagoa (1984). Using the Agarwal time equivalent function accounts for the length of the drawdown time period, and allows recovery plots to be analyzed with drawdown type curves. The time equivalent function, Δt_e, is defined by Agarwal (1980) as:

$$\Delta t_e = (t \times t')/(t + t') \qquad (A.1)$$

where, t = duration of the pumping test; (T)

  t' = time since pumping terminated; (T)

As shown in the Figure A.1, diagnostic analysis of the log-log recovery data and data derivative plot indicates the following effects and conditions during the recovery period:

- wellbore storage effects dominant up to a Δt_e ≈ 0.15 min

- delayed-yield conditions significant during the time period $\Delta t_e \approx 0.15$ min to 6 min

- radial flow conditions established between the time period $\Delta t_e \approx 6$ min to 200 min, and

- presence of a recharge boundary after a $\Delta t_e \approx 200$ min.

The presence of a recharge boundary in the later stages of recovery are attributed either to wastewater discharges occurring at the 1301-N crib facility, surface discharge of groundwater withdrawn during the constant-rate pumping test, or the recharge boundary condition imposed by the Columbia River.

Figure A.2 shows the semi-log, straight-line analysis of recovery data utilizing the Theis (1935) recovery method. Only recovery data that exhibit radial flow conditions (i.e., $\Delta t_e = 6$ to 200 min; {t+t'}/t') were analyzed. As indicated in the figure, an estimate of 5,640 ft²/d was obtained form the straight-line analysis. Recovery data shown in Figures A.1 and A.2 were corrected for unconfined aquifer de-watering using the procedure presented in Jacob (1963). Partial penetration effects were evaluated with applicable software presented in Walton (1987). The analysis indicated that partial-penetration effects were relatively constant during the radial flow period of the test data and, therefore, of no significant consequence to the straight-line analysis of test data exhibiting radial flow conditions.
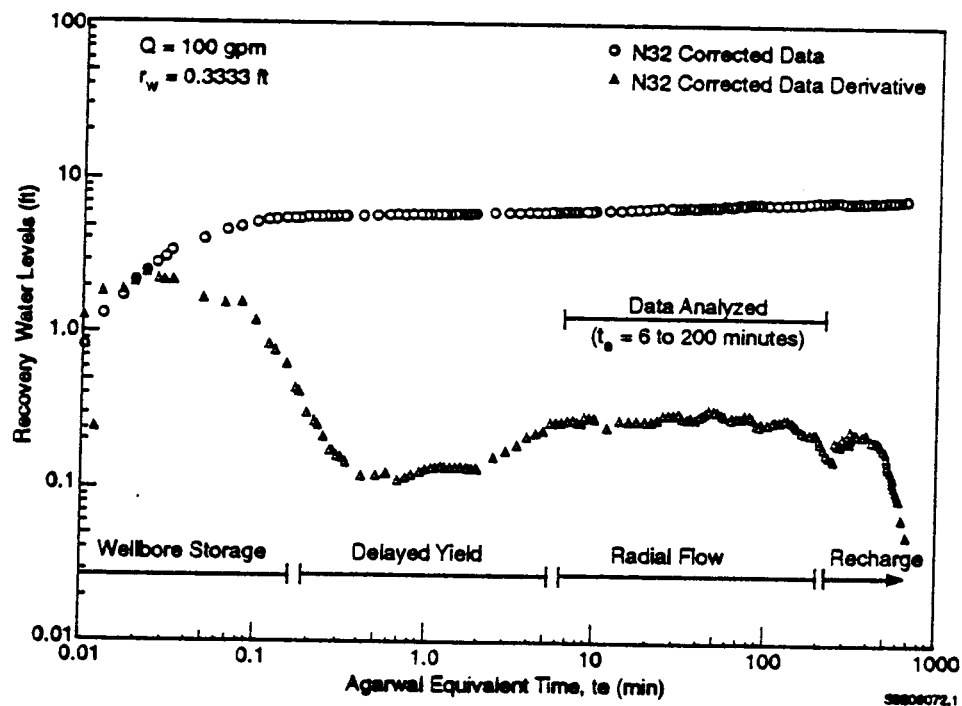
Figure A.1. Log-Log Plot of Water-Level Recovery Data and Pressure Derivative of Recovery Data
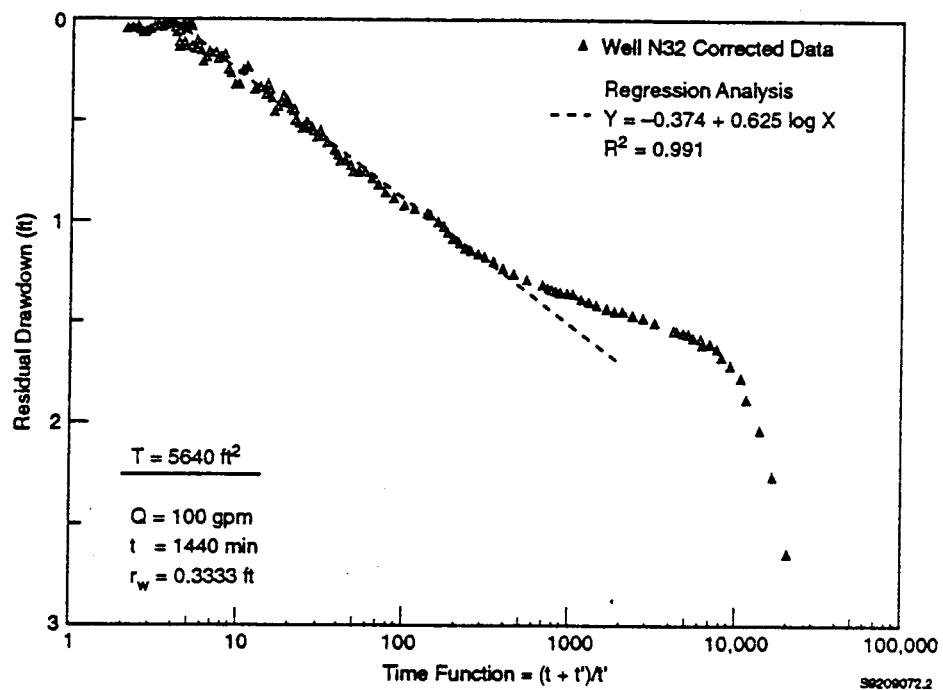
Figure A.2. Semi-Log Plot of the Water-Level Recovery Data

The chart contains the following labels:

- ▲ Well N32 Corrected Data
- Regression Analysis
- Y = −0.374 + 0.625 log X
- R^2 = 0.991
- Residual Drawdown (ft)
- Time Function = (t + t')/t'
- T = 5640 ft^2
- Q = 100 gpm
- t = 1440 min
- r_w = 0.3333 ft

38209072.2

Appendix B

Computer Code for Aquifer Parameter Estimation

```
c****************************************************************
c  tr.for
c
c  Program to estimate aquifer transmissivity
c
c  Uses finite difference solution to river-well problem
c  and iterates to best fit of predicted results with recorded
c  well data.  Best fit is judged by comparing average magnitude of changes
c  in well height and predicted well height during last week of simulation.
c
c  Written by: Chris Sherwood, Battelle PNL
c  Last revised: 3/23/92 on Gateway 486
c
c  This code will run either either fully implicit or Crank-Nicholson
c  formulation for the linear heat equation.
c  Results are written for one x position.
c
c  Present configuration allows for IMAX=501 nodes in horizontal and
c  MMAX=4000 input records.
c
c  Special caveats:
c      Input times are ignored; the program assumes time 1 = 0 and
c      subsequent times are t+dt, output times are same as input times
c      for river data.
c
c  Input file WR.IN:
c      First record contains up to 80 characters of comments
c      Second record contains file name of river input file
c      Third record contains file name of river data
c      Fourth line contains parameters on one line,
c      separated by blanks or commas.  Parameters are are floating-point
c      numbers read in general format, except for first
c      one, which is an integer:
c
c      numerical_scheme - switch: 0=implicit, 1=Crank-Nicholson
c      s - storativity
c      dx - delta x
c      xb - distance to boundary where h is held constant
c      xw - distance to well where h(t) output is recorded (actually, closest
c          computational node is used).
c      hrinit - initial river elevation
c      hbmean - mean boundary elevation (together, hrinit and hbmean establish
c          initial slope; if hbmean < hwmean from data, hwmean is used)
c      aqthick - thickness of aquifer
c      dt - time step of input time series
c      tmult - factor to shorten time-step in model: actual delta t used in
c          model is dt/tmult.  Output h at xw is only written at dt intervals.
c
c  Input file called rivdatfile
c      Format 1:
c      ttt.ttt rrr.rrr
c      ASCII file containing up to MMAX times and river stages
c      at constant time interval dt.
c
c  Input file called welldatfile
c      Format 1:
c      ttt.ttt www.www
c      ASCII file containing up to MMAX times and well elevations
c      at constant time interval dt.
c
c  Output file TROUT.DAT:
c      ASCII file with up to MMAX records, each containing:
c      t - time from river height input file
```

```
c       riv - river height from input file
c       well - well-water elevation from input file
c       wellp - model-predicted height at xw
c
c***************************************************************
        program tr
        parameter(IMAX = 501, MMAX = 4000, ACC = 0.001, MAXIT = 50 )
        integer m, mcount, wcount, errno, iwell
        character*80 comment
        character*30 infile /'tr.in'/
        character*30 outfile /'trout.dat'/
        character*30 rivdatfile /'rvrmmmyy.dat'/
        character*30 welldatfile /'nnnmmmyy.dat'/
        character*30 err_mess(1) /'Fatal error.'/
        character*9 cdate
        character*8 ctime
        character*30 scheme_name(2) /'Implicit Scheme',
     *                               'Crank-Nicholson Scheme'/
        character*30 version /'tr.for, version of 3/16/92'/

c       ...set dimensions: IMAX is x dimension, MMAX is time
        real rtime(MMAX), riv(MMAX), well(MMAX), wellp(MMAX)
        real x(IMAX), hx(IMAX)
        real k /1.0/
        real s /1.0/

c       ...misc intermediate values
        integer n, nstart, npts
        real meanr, meanw, meanwp
        real trans, kw, kwp

c       ...Model domain - space
        real dx /50./
        real xb /5000.0/
        real xw /1500./

c       ...Model domain - time
        real dt /60.0/
        real tmult /4.0/

c       ...Initial conditions
        real hrinit  /350./
        real hbmean  /350./
        real aqthick /50.0/

c       ...Run information
        call date( cdate )
        call time( ctime )
        write( *,'(1x,a)') version
        write( *,1000) cdate, ctime
1000    format( 1x,'Run on ',a,' at ',a,'.')

c       ...Read model parameters
        open( unit=5, file=infile, status='old')
        read(5,'(a80)') comment
        read(5,'(a30)') rivdatfile
        read(5,'(a30)') welldatfile
        read(5,*) numerical_scheme, s, dx, xb, xw, hrinit,
     &            hbmean, aqthick, dt, tmult
        close(5)

c       ...List model parameters
        write(*,'(1x,a30)') comment
```

B.2

```
      write(*,1005) rivdatfile, welldatfile
1005 format(1x,'River data file: ',a30,/,
    &        1x,'Well data file : ',a30)
      write(*,'(1x,a)') scheme_name(numerical_scheme+1)
      write(*,1010) s,dx,dt
1010 format(1x,'S = ',F9.4,/,
    &        1x,'dx= ',F9.4,'      dt= ',F9.4)
      write(*,1020) hrinit, hbmean, aqthick
1020 format(1x,'hrinit = ',F7.2,' hbmean = ',F7.2,/,
    &        1x,'aqthick = ',F7.2 )

c     ...Read in time-history of river data
      open(5,file=rivdatfile,status='old')
      write(*,2000) rivdatfile
2000 format(1x,'Reading river input file: ',a)
      mcount = 0
      do 200 m=1,MMAX
         read(5, *, end = 209) rtime(m), riv(m)
         mcount = mcount+1
200  continue
209  close(5)
      if( mcount .eq. 0 ) then
         write(*,*) 'Error reading input file.'
         stop
      else if( mcount .eq. MMAX ) then
         write(*,*) '...done.  Read ',mcount,' points.'
         write(*,*) '***WARNING***'
         write(*,*) 'Array full; may not have read all points.'
      else
         write(*,*) '...done.  Read ',mcount,' points.'
      endif

c     ...Read in time-history of river data
      open(5,file=welldatfile,status='old')
      write(*,2001) welldatfile
2001 format(1x,'Reading well input file: ',a)
      wcount = 0
      do 201 m=1,mcount
         read(5, *, end = 219) wtime, well(m)
         wcount = wcount+1
201  continue
219  close(5)
      if( wcount .eq. 0 ) then
         write(*,*) 'Error reading input file.'
         stop
c
c   bug fix -- changed .le. to .lt. -- SK Wurstner  3/23/92
      else if( wcount .lt. mcount ) then
         write(*,*) '...done.  Read ',wcount,' points.'
         write(*,*) '***WARNING***'
         write(*,*) 'Well data shorter than river data.'
c        ...process fewer points
         mcount = wcount
      else
         write(*,*) '...done.  Read ',m,' points.'
      endif

c     ...calculate means of river and well data
      call mv( riv, mcount, hrmean, rsdev, 0 )
      call mv( well, mcount, hwmean, wsdev, 0 )
      if(hbmean .lt. hwmean) then
         hbmean = hwmean
         write(*,*) '*** Specified hbmean < hwmean ***'
```

B.3

```fortran
               write(*,*) 'Calculated value of hwmean substituted'
            endif
            write(*,2100) hrmean, rsdev, hwmean, wsdev, hbmean
 2100 format( 1x,'hrmean: ',f8.3,' std. dev: ',f9.5,/,
     &1x,'hwmean: ',f8.3,' std. dev: ',f9.5,/,
     &1x,'hbmean: ',f8.3 )

c     ...calculate average change for last
c         week of data
      npts = 60.*60.*24.*7./dt
      nstart = max( 2, mcount-npts )
      call absmean( well, mcount, npts, meanw )
      call absmean( riv, mcount, npts, meanr )
      write(*,2200) npts, meanr, meanw
 2200 format(1x,'npts: ',i4,/,
     &1x,'Mean delta r: ',F8.4,/,
     &1x,'Mean delta w: ',F8.4)

c     ...estimate K from ratio of changes
      call xpinder( meanw, meanr, xw, trans, s, dt )
      kw = trans/aqthick
      write(*,2300) kw
 2300 format(1x,'Kw from hw/hr: ',f8.4 )

      it = 1
      k = kw
      kwp = 0.
c     do 10 while (it .lt. MAXIT .and. abs( kw-kwp ).gt.ACC )
c     Microsoft compiler (older version) will not accept do while loop
c     Modified to do loop and if block --- SK Wurstner 3/19/92
      do 10 ii=1,MAXIT
         if(it .lt. MAXIT .and. abs( kw-kwp ).gt.ACC ) then
         alpha = (k/s) * (dt/tmult)/(dx*dx)
         call wrsub( numerical_scheme, alpha,
     &              dx, xb, xw, hrinit,
     &              hbmean, aqthick, dt, tmult,
     &              mcount, errno, iwell,
     &              riv, wellp, x, hx )
         if( errno .gt. 0 )then
            write(*,3000) err_mess
 3000       format(1x,'***',a,'***')
            stop
         endif
c     ...calculate average delta well and wellp for last
c         week of data
         npts = 60.*60.*24.*7./dt
         n = npts
         call absmean( well, mcount, npts, meanw )
         call absmean( wellp, mcount, npts, meanwp )
         if( n .ne. npts) write(*,*) 'Dropped values in averaging.'
         write(*,4000) npts, meanw, meanwp
 4000    format(1x,'npts: ',i5,'   Mean delta w: ',F8.4,
     &   '   Mean delta w predicted: ',F8.4)
c     ...estimate kwp from ratio of changes
         call xpinder( meanwp, meanr, xw, trans, s, dt )
         kwp = trans/aqthick
         write(*,4300) kwp
 4300    format(1x,'Kwp from hwp/hr: ',f9.5)
         k = k*kw/kwp
         it = it+1
         write(*,4400) it, k
 4400    format(1x,'It: ',i3,' K: ',f9.5)
         endif
```

B.4

```
      10 continue

c        ...Write results
         write(*,5000) k, k*aqthick
   5000 format(1x,'Final estimates:',/,
      &          1x,'K= ',f9.5,/,
      &          1x,'T= ',f9.5)

         write(*,*) 'Writing results to trout.dat...'
         open(unit=6, file=outfile, status='unknown')
         t=0.
         do 500 m=1,mcount
            write(6,*) rtime(m), riv(m), well(m), wellp(m)
            t = t+dt
  500    continue
         close(6)
         write(*,*) 'done.'
         end

c*********************************************************************
c
         subroutine wrsub( numerical_scheme, alpha,
      *              dx, xb, xw, hrinit,
      *              hbmean, aqthick, dt, tmult,
      *              mcount, errno, iwell,
      *              riv, wellp, x, hx )
c
c*********************************************************************
         parameter(IMAX = 501, MMAX = 4000)
         integer numerical_scheme, errno, mcount, iwell
         real dx, xb, xw, hrinit, hbmean, aqthick, dt, tmult
         real riv(*), wellp(*), x(*), hx(*)

         real bb(IMAX), dd(IMAX), aa(IMAX), cc(IMAX)

         integer i, kk, n, m
c        ...Boundary condition at river.
c           (Calculated from input river heights.)
         real dh

c        Set initial conditions, which assume constant slopes
         n = xb/dx + 1
         if( n .gt. IMAX )then
            write(*,*) '***ERROR***'
            write(*,*) 'Arrays not large enough for specified xb and dx.'
            write(*,*) 'xb = ',xb,' and dx = ',dx,' requires IMAX = ',IMAX
            errno = 1
            return
         endif
         do 100 i=1,n
            x(i) =   (i-1)*dx
            hx(i) = hrinit + ((hbmean-hrinit)/xb) * x(i)
  100    continue

c        Find x node near well at xw
         iwell = (xw/dx) + 1
         write(*,1500) iwell, x(iwell)
c 1500 format(1x,'Well data from x(',i4,') = ',F7.3)
c        Changed format to accomodate larger numbers
   1500 format(1x,'Well data from x(',i4,') = ',F9.3)
         wellp(1) = hx(iwell)

c        --- Model starts here ---
```

```
            write(*,3000)
 3000 format(1x,'Calculating'\)
c     For each time step....
      t = 0.
c     write(*,*) t, riv(1), wellp(1)
      do 400 m=2,mcount
          if( mod(m,100).eq.0 ) write(*,3500)
 3500   format('.'\)
c         ...Solve model at shorter time step dt/tmult
          dh = (riv(m)-riv(m-1))/tmult
          do 300 kk=1,tmult
              t = t+dt/tmult
              if( numerical_scheme .eq. 0 ) then
c                 ...implicit formulation
                  do 210 i=2,n-1
                      bb(i) = -alpha * aqthick
                      dd(i) = 1+2*alpha* aqthick
                      aa(i) = -alpha * aqthick
                      cc(i) = hx(i)
  210             continue
              else
c                 ...Crank-Nicholson
                  do 220 i=2,n-1
                      bb(i) = -0.5 * alpha * aqthick
                      dd(i) = 1+alpha * aqthick
                      aa(i) = -0.5 * alpha * aqthick
  220             continue
                  do 230 i=2,n-1
                      cc(i) = 0.5 * alpha * aqthick * hx(i-1) +
     *                        (1-alpha * aqthick) * hx(i) +
     *                        0.5 * alpha * aqthick * hx(i+1)
  230             continue
              endif
c             ...Boundary conditions (force hbmean and hx(t))
              cc(1) = hx(1) + dh
              cc(n) = hbmean
              bb(1) = 0.
              dd(1) = 1.
              aa(1) = 0.
              bb(n) = 0.
              dd(n) = 1.
              aa(n) = 0.
c             ...Solve tridiagonal matrix
              call sy(1, n, bb, dd, aa, cc)
c             ...Copy answer to hx
              do 250 i=1,n
                  hx(i) = cc(i)
  250         continue
  300     continue
c         ...Save results for only one x location
          wellp(m) = hx(iwell)
c         write(*,*) t, riv(m), wellp(m)
  400 continue
      write(*,*) 'done.'
c     --- Model ends here ---
      return
      end
c****************************************************************
c
      subroutine sy(il, iu, bb, dd, aa, cc)
      integer il, iu
      dimension bb(*), dd(*), aa(*), cc(*)
c
```

B.6

```fortran
c     Solves tridiagonal system of equations by elimination
c
c     From: Anderson, Tannehill, and Pletcher, 1984:
c     "Computational Fluid Mechanics and Heat Transfer"
c     Appendix A.
c
c     il = subscript of first equation
c     iu = subscript of last equation
c     bb = coefficient behind diagonal
c     dd = coefficient on diagonal
c     aa = coefficient after diagonal
c     cc = constant vector on input; solution on return
c
c************************************************************
      integer i, j
      real r

c        ...establish upper triangular matrix
      lp = il+1
      do 10 i = lp,iu
         r = bb(i)/dd(i-1)
         dd(i) = dd(i)-r*aa(i-1)
         cc(i) = cc(i)-r*cc(i-1)
 10   continue
c
c        ...back substitution
      cc(iu) = cc(iu)/dd(iu)
      do 20 i = lp,iu
         j = iu-i+il
         cc(j) = (cc(j)-aa(j)*cc(j+1))/dd(j)
 20   continue
      return
      end
c************************************************************
c
      subroutine absmean( data, length, npts, rmean )
c
c     calculates mean of absolute value of change in last npts of data
c     actual n found returned in npts
c
c************************************************************
      real data(*)
      integer length, npts
      integer i,nstart,n
      real rmean
      nstart = max( 2, length-npts+1 )
      n = 0
      rmean = 0.
      do 100 i=length,nstart,-1
         n=n+1
         rmean = rmean + abs( data(i)-data(i-1) )
 100  continue
      if( n .ne. npts ) then
         write(*,*) 'Warning: n= ',n,' and npts= ',npts
      endif
      rmean = rmean/n
      npts = n
      return
      end
c************************************************************
c
      SUBROUTINE DATE(CDATE)
c
```

```
C       VAX-LOOK ALIKE SHELL FOR MICROSOFT RUN-TIME DATE CALLS
C
C*******************************************************************
        CHARACTER*9 CDATE
        CHARACTER*3 CMO(12) /'Jan','Feb','Mar','Apr','May','Jun','Jul','Au
       +g','Sep','Oct','Nov','Dec'/
        INTEGER*2 IYR,IMO,IDA
            CALL GETDAT(IYR,IMO,IDA)
        WRITE(CDATE,1000) IDA,CMO(IMO),IYR-1900
  1000 FORMAT(I2,'-',A,'-',I2)
        RETURN
        END
C*******************************************************************
C
        SUBROUTINE TIME(CTIME)
C
C       VAX-LOOK ALIKE SHELL FOR MICROSOFT RUN-TIME TIME CALLS
C
C*******************************************************************
        CHARACTER*8 CTIME
        INTEGER*2 IHR,IMN,ISC,ISHUN
            CALL GETTIM(IHR,IMN,ISC,ISHUN)
        WRITE(CTIME,1000)IHR,IMN,ISC
  1000 FORMAT(I2,':',I2,':',I2)
        RETURN
        END
C*******************************************************************
c
        integer function iexp2(n)
c
c...returns exponent of 2 greater such that ipow2 is less than n
c
C*******************************************************************
        integer n
        j = 1
        do 100 i=1,100
            iexp2 = j
            ipow2 = 2**i
            if (ipow2 .gt. n) goto 110
            j = j + 1
  100   continue
  110   iexp2 = j-1
        return
        end
C*******************************************************************
c
        subroutine mv(data,n,rmean,sdev,iflag)
c
c   calculates mean and standard deviation of data
c   when iflag = 0, data is returned unchanged
c        iflag = 1, mean is subtracted from the data
c        iflag = 2, mean is subtracted from the data, then
c                   data is divided by standard deviation (normalized)
c        iflag = 3, data is divided by standard deviation
c
c   last revised: 2/19/88
c
C*******************************************************************
        real data(*), rn, rmean, sdev
        double precision sum, sumsq
        real rsdev, rrmean
        integer n
        sum = 0.
```

B.8

```fortran
         sumsq = 0.
         do 100 i=1,n
            sum = sum + data(i)
            sumsq = sumsq + data(i)*data(i)'
  100    continue
         rn = float(n)
         rmean = sum/rn
         var = (rn*sumsq - sum*sum)/(rn*(rn-1.))
         if(var.gt.0.) then
            sdev = sqrt(var)
         else
            sdev = 1.
            write(*,*) 'Warning: variance = ',var
         endif
         if(iflag.eq.0) then
            return
         else if (iflag.eq.1) then
            rrmean = rmean
            rsdev = 1.
         else if (iflag.eq.2) then
            rsdev = sdev
            if(sdev.le.0.0) then
               rsdev = 1.
               write(*,*) 'Std. dev. <= zero: normalization not performed.'
            endif
            rrmean = rmean
         else if (iflag.eq.3) then
            rsdev = sdev
            if(sdev.le.0.0) then
               rsdev = 1.
               write(*,*) 'Std. dev. <= zero: normalization not performed.'
            endif
            rrmean = 0.0
         else
            pause 'Illegal value of IFLAG in subroutine MV'
            return
         endif
         do 200 j=1,n
            data(j) = (data(j)-rrmean) / rsdev
  200    continue
         return
         end
c******************************************************************************
c
      subroutine lfit(data,n,a,b)
c
c   returns linear fit parameters a anb b, where
c
c         estimate of data(x) = a + bx
c
c         where x = n*deltatx
c
c   Equations 11.16a anb 11.16b in Bendat and Piersol, 1986, p. 364.
c
c   Written by Chris Sherwood, Battelle, PNL
c
c   Last revised: 2/22/88
c
c******************************************************************************
      parameter (DELTAX = 1.)
      integer n
      real data(n)
      real a,b
```

```fortran
      integer i
      double precision sumx, sumxn, da, db,dn
      sumx = 0.0
      sumxn = 0.0
      dn = dble(n)
      do 100 i=1,n
        sumx = sumx + dble(data(i))
        sumxn = sumxn + dble(i)*dble(data(i))
100   continue
      da = ( 2.*(2.*dn+1)*sumx - 6.*sumxn ) / (dn*(dn-1.))
      db = ( 12.*sumxn - 6.*(dn+1.)*sumx ) / (DELTAX*dn*(dn-1.)*(dn+1.))
      a = real(da)
      b = real(db)
      return
      end
c****************************************************************
c       .
      subroutine detr(data,n,a,b)
c
c   Applies linear fit parameters a anb b to data, where
c
c           estimate of data(n) = a + bn           .
c
c           residual of data(n) = data(n) - estimate(n)
c
c   Returns residual in original data array.
c   Equations 11.16a anb 11.16b in Bendat and Piersol, 1986, p. 364.
c
c   Written by Chris Sherwood, Battelle, PNL
c
c   Last revised: 2/22/88
c
c****************************************************************
      integer n
      real data(*),a,b
      integer i
      do 100 i = 1,n
        data(i) = data(i) - (a + b * float(i))
100   continue
      return
      end
c****************************************************************
c
      real function erfcc( x )
c
c     Complementary error function
c     from Press et al. 1986 "Numerical Recipes", p. 164
c
c****************************************************************
      real x
      real t, z

      z=abs(x)
      t=1.0/(1.0+0.5*z)
      erfcc = t*exp(-z*z-1.26551223+
     &    t*(1.00002368+t*(0.37409196+t*(0.09678418+
     &    t*(-0.18628806+t*(0.27886807+t*(-1.13520398+t*(1.48851587+
     &    t*(-0.82215223+t*0.17087277)))))))))
      if(x .lt. 0.) erfcc = 2.0-erfcc
      return
      end
c****************************************************************
c
```

B.10

```fortran
      real function xerfcc( erfcx )
c
c     Calculates x given erfc(x) by Newton's method
c     Returns when accuracy XACC is reached or maximum
c     number of iterations MAXIT is reached.
c
c     Based on RTNEWT in Press et. al. 1986, p. 254
c     Written by:  Chris Sherwood, Battelle MSL
c
c******************************************************************
      parameter( XACC=0.0001, MAXIT = 100 )
      real erfcx
      real dx, f, df
      real erfcc

      dx = 1
      xerfcc = .5
      do 10 i=1,MAXIT
         f = erfcc( xerfcc )-erfcx
         df = ( erfcc( xerfcc+5*XACC ) - erfcc(xerfcc-5*XACC) )
     &        /(10*XACC)
         dx = f/df
         xerfcc = xerfcc-dx
         if( abs(dx) .lt. XACC ) return
 10   continue
      write(*,*) 'MAXIT reached in xerfcc.'
      return
      end
c******************************************************************
c
      subroutine xpinder( hw, hr, x, trans, stor, time )
c
c     Calculates transmissivity trans from amplitudes of
c     river variations hr, well variations hw, storativity stor,
c     and time.  Based on approach of Pinder (1969).
c     (Equation 5.6 of Sherwood and Newcomer, 1992)
c
c     Written by: Chris Sherwood, Battelle MSL
c
c******************************************************************
      real hw, hr, x, trans, stor, time
      real arg, xerfcc
      arg = xerfcc( hw/hr )
      trans = (stor/time)*(0.5*x/arg)**2
c     write(*,*) 'hw/hr=',hw/hr,' arg = ',arg,' trans = ',trans
      return
      end
```

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