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Estimating Aquifer Hydraulic Properties Using the Ferris Method, Hanford Site, Washington

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1.0 INTRODUCTION

Aquifer hydraulic properties play an important role in the transport of contamination via groundwater. Both the amount of hazardous constituents that can be held in an aquifer and the rate of movement are strongly dependent on properties such as the capacity to store liquids and the ease with which liquids may flow through the aquifer. Computer models that predict the flow of groundwater, along with any contamination it may contain, rely on aquifer hydraulic properties as variables.

Aquifer hydraulic properties may be measured or estimated by several methods. Pump tests in monitoring wells are commonly run for this purpose. Estimates also may be derived from a knowledge of the geologic materials that provide a framework for the aquifer. Some researchers have investigated the relationship between fluctuating water levels in streams and corresponding fluctuations in nearby groundwater wells. Inferences regarding aquifer properties are then made by analyzing the changes in characteristics of these fluctuations with increasing distance from the stream.

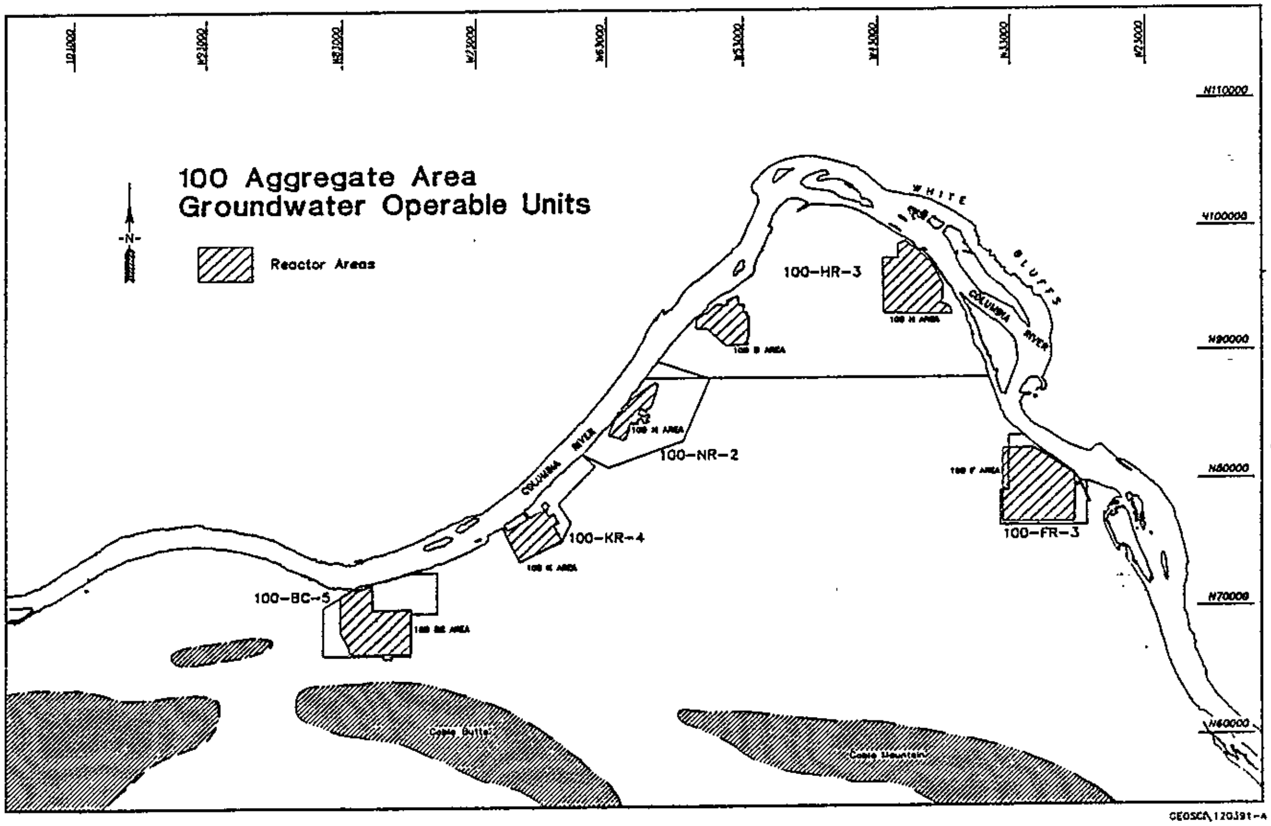
This report focuses on the latter method. It contains a review of previous work of a similar nature on the Hanford Site, as well as an application of the method to recently collected water level data for the Site.

1.1 PURPOSE AND SCOPE

This investigation contributes to several tasks described in work plans (e.g., DOE-RL 1992) associated with groundwater operable units for the 100 Areas (Figure 1-1). Work Plan Task 6, "Groundwater Investigation," of each work plan provides for estimating aquifer properties and characterizing river/groundwater interaction. Appendix D-1, "Surface Water/Sediment Investigation for the 100 Areas," which is a part of each groundwater work plan, describes the installation of river stage recorders and data loggers in shoreline monitoring wells.

These tasks are oriented towards providing a better understanding of the flow of contaminated groundwater from the Hanford Site into the Columbia River. Because a part of this understanding will come from modeling groundwater flow, a knowledge of aquifer hydraulic properties is necessary. This investigation has explored one method of estimating several aquifer hydraulic properties.

Figure 1-1. Location Map for 100 Aggregate Area Groundwater Operable Units.



The investigation described in this report was completed under a requirement contained in 100 Aggregate Area Milestone M-30-04 of the *Hanford Federal Facility Agreement and Consent Order* (Tri-Party Agreement) (Ecology et al. 1990), which states:

> *"Submit a report (secondary document) to EPA and Ecology evaluating the interaction of the Columbia River and the unconfined aquifer for aquifer hydraulic properties."*

Discussions among Tri-Party Agreement participants have resulted in a well-defined scope for this investigation. The scope includes (1) evaluating published methods for inferring aquifer properties from stream/groundwater interaction; (2) determining the suitability of various methods for the Hanford Site; (3) collecting data from the 100 Areas; (4) applying the preferred method to Site-specific data; and (5) comparing the results to estimates derived by other means.

1.2 RIVER/GROUNDWATER INTERACTION: A SYNOPSIS

As it passes through the Hanford Site, the Columbia River can be described generally as a "gaining" stream, since groundwater flowing under the Site ultimately discharges into the river. This has been the case both prior to and during Hanford Site operations. Most of this discharge takes place out of sight, through the submerged part of the river channel. A minor portion of discharge occurs along the riverbank, where groundwater seepage can be observed during periods of low river levels. During periods of high river levels, river water flows into the riverbank, where it either mixes with groundwater or overlies the groundwater. When the river level falls again, both the river water and groundwater stored in the riverbank flow back into the river. This phenomenon is referred to in technical literature as "bank storage." Newcomb and Brown (1961) describe bank storage along the Hanford Reach of the Columbia River.

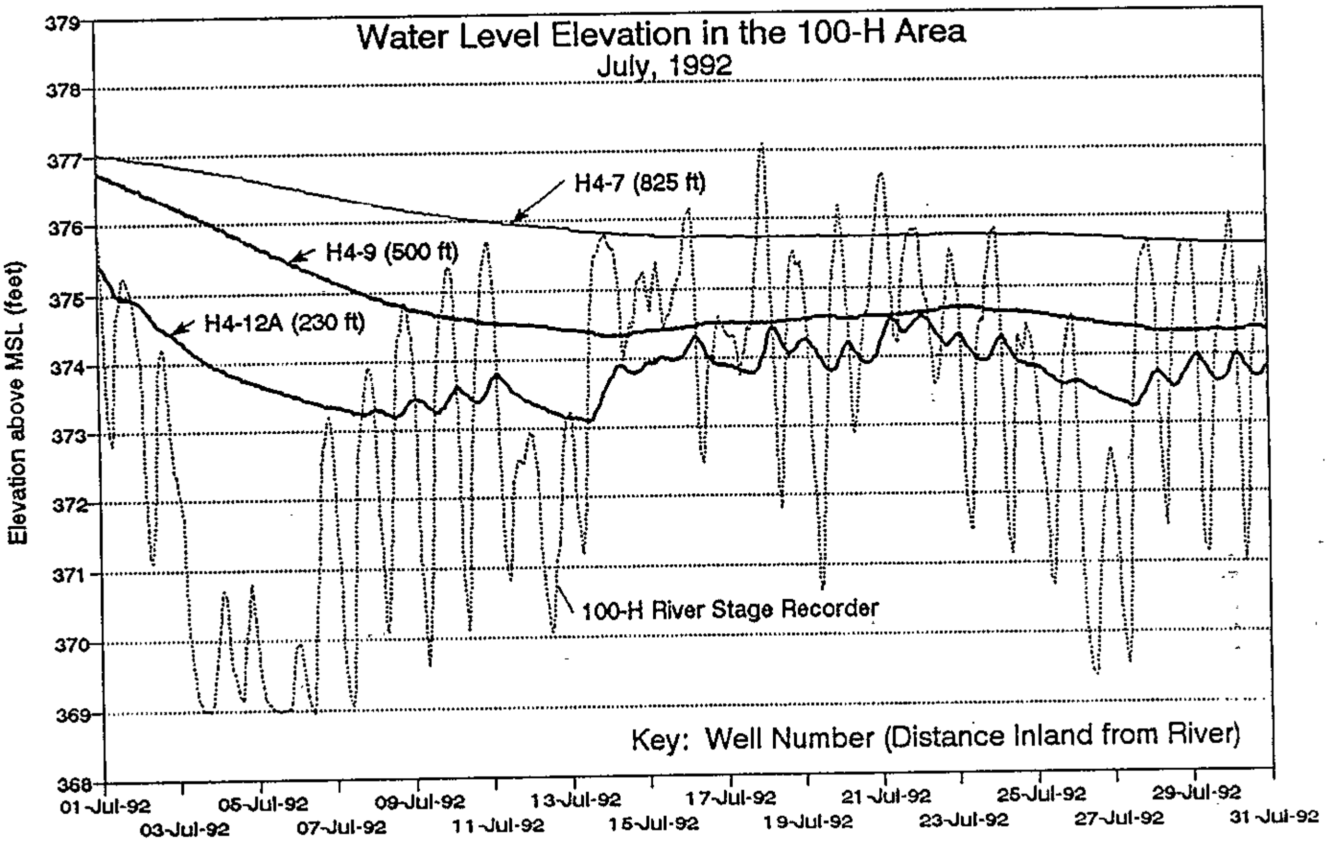
The Columbia River level rises and falls on a regular basis. River levels are related primarily to power-generating operations by upstream dams and cause daily fluctuations. Levels are related secondarily to the seasonal variability in natural runoff of precipitation. Along the 100 Areas shoreline, daily fluctuations may result in an elevation change of 6 to 8 ft, while seasonal fluctuations may cover an 8- to 10-ft range. The amount of river water that flows into the riverbank is directly related to the height and duration of high water levels in the river.

As the river fluctuates up and down, a pressure wave is transmitted inland through the groundwater. Daily changes in river levels are observed easily in wells at distances of several hundred feet or less inland from the river (Figure 1-2). Weekly, monthly, and seasonal changes are observable at correspondingly greater distances inland, which range to thousands of feet. The inland distances to which pressure waves from river fluctuations propagate vary with the magnitude and duration of the fluctuations, as well as the geologic characteristics of the aquifer.

The pressure wave from river fluctuations can be observed much farther inland than the extent to which river water invades the riverbank during high river levels. For river water to actually flow inland, the river level must be higher than the nearby groundwater surface and must remain high for a sufficiently long period for water to flow through the sediments. Typically, this inland flow of river water is restricted to within several hundred feet or less of the shoreline.

In addition to water movement in and out of the riverbank (i.e., one-dimensional flow perpendicular to the shoreline), there is a component of flow towards the downstream direction (Newcomb and Brown 1961). Because the river flows downstream in response to an elevation gradient, groundwater and bank storage also tend to travel downstream, although at a considerably slower rate than the river flow. Think of bank storage water as zig-zagging downstream along the shoreline of a fluctuating river. Finally, to complete the three-dimensional flow picture, vertical components of flow are induced by a stream that does not fully penetrate the aquifer, and by water table gradients associated with the bank storage phenomenon.

Figure 1-2. Typical Hydrographs for the Columbia River and Monitoring Wells Along the 100 Areas Shoreline.



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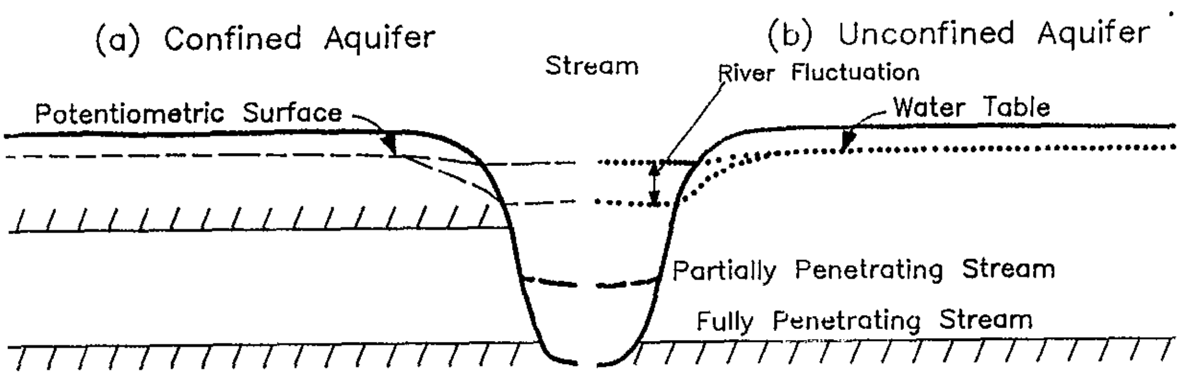
1.3 RELATIONSHIP BETWEEN AQUIFER PROPERTIES AND WATER LEVELS

Hydrologists have long been intrigued by the response of groundwater, as observed in wells, to a stress applied to the aquifer at an aquifer boundary. An example of stress is hydrostatic pressure induced by a fluctuating stream, flood event, or ocean tide. Numerous attempts have been made to describe mathematically the response of the aquifer to an induced stress, and to use the relationship to infer hydraulic properties (Appendix A summarizes the literature on this topic). Figure 1-3 illustrates the various aquifer/stream configurations that must be considered for this research.

Some success has been achieved for confined aquifers that are fully penetrated by a stream (Figure 1-3a), where there is free hydraulic interchange between the stream and the aquifer and flow is predominantly one-dimensional. Efforts have been less successful for unconfined aquifers that are partially penetrated by a stream (Figure 1-3b), since three-dimensional flow has a stronger influence on the interchange. (This disadvantage can be minimized by choosing observation wells at the greatest distance possible where the pressure wave passage can be observed.) A significant difficulty yet to be overcome involves describing mathematically the free surface of the water table for an unconfined aquifer as it responds to a nearby fluctuating stream that partially penetrates the aquifer.

Figure 1-3. Aquifer/Stream Configurations for (a) Confined Aquifer and (b) Unconfined Aquifer.



The aquifer/stream configuration shown in Figure 1-3b best describes conditions along the Hanford Site shoreline of the Columbia River. That is, the Columbia River partially penetrates the unconfined aquifer. The areas of interest with regard to contaminant transport are located relatively close to the shoreline, such that choosing distant observation wells becomes irrelevant. The heterogeneous geologic characteristics of the unconfined aquifer, which include wide variations in sediment size, sorting, and consolidation, as well as frequent facies changes caused by a fluvial depositional environment, all combine to place formidable challenges to creating a tractable mathematical relationship between water levels and hydraulic properties.

The aquifer hydraulic properties of most interest in groundwater contamination investigations relate to the capability of the aquifer to (1) store water (e.g., storativity, storage coefficient) and (2) transmit water (e.g., transmissivity, hydraulic conductivity). Hydraulic diffusivity, which reflects the combined effects of storage and transmitting properties, is an aquifer characteristic that may be estimated from analysis of water level fluctuations. All of these properties are controlled or defined by the type of aquifer, which may be unconfined, semiconfined, or confined.

The several methods previously investigated for inferring hydraulic properties from water level data are discussed in more detail in Appendix A and Chapter 2.0. The method chosen for evaluation during this project involves measuring several characteristics of sinusoidal fluctuations in river levels and corresponding fluctuations in observation wells located some distance inland from the river. It is referred to as the Ferris method (Ferris 1952, 1963). The pulse produced by a rise and fall in river level is observed in the well at a reduced amplitude and at a later time (see Figure 1-2). The ratio between river pulse height and well pulse height, and the lag time between the river and the well, are each used to infer hydraulic diffusivity, which is the ratio of transmissivity to storativity.

2.0 PREVIOUS INVESTIGATIONS ON THE HANFORD SITE

Several investigations to infer aquifer properties by analyzing water level data have been undertaken on the Hanford Site. None has as yet been shown to be useful for detailed mapping of spatial variations in aquifer properties. This is perhaps due to the limited geographic coverage of the data sets that were available. It also may be due to inherent difficulties caused by heterogeneities in Hanford Site aquifers (e.g., Poeter and Gaylord 1990). However, these previous analyses and the current analysis serve to test the possibility that spatial variations could be delineated using water level data, given a sufficiently comprehensive data set.

2.1 BIERSCHENK (1959)

This landmark report on aquifer characteristics and groundwater movement for the Hanford Site contains the results of several methods for estimating aquifer hydraulic properties, including the analysis of cyclic fluctuations in wells. Bierschenk used the Ferris method to infer "transmissibility" (transmissivity) and "field permeability" (hydraulic conductivity) for several wells located between the Gable Butte/Gable Mountain trend and the Columbia River. He used average values over periods of 3 to 12 years for water level ranges in the river and individual wells. These averages were then used as stage ratios in the Ferris equation, which relates stage ratio to coefficients for transmissibility and storage. Inferences regarding transmissibility and field permeability were thus based on cyclic fluctuations due to seasonal changes in water levels caused by annual flood crests in the Columbia River.

His results are summarized in Table 2-1. The appendix from his report, which describes the analysis, is reproduced in Appendix D, along with a location map for the wells he used. Bierschenk does not discuss the assumptions and limitations of the Ferris method. He describes his results as "tentative estimates . . . that serve merely to demonstrate the applicability, usefulness, and limitation of the method . . ." The values obtained do compare within an order of magnitude to estimates derived by other means.

2.2 100-N AREA STUDIES

In 1960, an analysis of aquifer hydraulic properties was performed in the 100-N Area to help evaluate the performance of a proposed waste water disposal facility (Brown and Rowe 1960). The analysis method (Rowe 1960) was inspired by the earlier work of Ferris (1952). Estimates for transmissibility and storage coefficient were derived from a linear change in river stage with time, in contrast to the sinusoidal fluctuations that were analyzed by the Ferris method. Unfortunately, an error was present in the initial method used by Rowe (1960), although it was subsequently discovered and corrected (Hantush 1961).

Table 2-1. Summary of Results from Cyclic Fluctuation Data for the Northern Hanford Site (modified from Bierschenk 1959).

Hydrologic unit	Well number	Transmissibility (gal/day/ft)	Transmissivity (ft^2/day)
Glaciofluviatile (Hanford gravels and overbank deposits)	699-60-60		
	-61-66		
	-65-72	2,300,000	307,487
	-63-90		
	-66-103		
	-57-29	610,000	81,551
	-62-32	790,000	105,615
Glaciofluviatile and Ringold Formation	699-63-25	130,000	17,380
	-67-77	190,000	25,401
	-70-68	240,000	32,086
	-HAN-23	260,000	34,759
Ringold Formation	699-71-84	15,000	2,005
	-72-88	51,000	6,818
	-92-38	32,000	4,278

NOTE: Transmissibility in (gal/day/ft) is converted to transmissivity in (ft^2/day) by dividing by 7.48. A storage coefficient of 0.06 was assumed for glaciofluviatile units and 0.1 for the mixed unit and Ringold Formation.

In spite of the error, estimates for transmissibility along the 100-N Area shoreline were consistent with earlier estimates made by Bierschenk (1959) for the northern part of the Hanford Site. Brown and Rowe (1960) estimated a transmissibility range of 30,000 to 60,000 gal/day/ft (4,011 to 8,021 ft^2/day), assuming a storage coefficient of 0.1. This range compares favorably with the range for the Ringold Formation shown in Table 2-1.

Newcomer (1988) conducted research on the interaction between the Columbia River and Hanford Site groundwater using water level data from 100-N Area during late 1987 and early 1988. He analyzed river and well water level data collected at 15-minute intervals. Using statistical methods to compare river data and well data, he described the correlation, time lag, and attenuation of river stage changes as they propagate landward. Although he did not attempt to infer aquifer hydraulic properties from the data, he noted the potential usefulness of his analysis in calibrating flow models for the near-river subsurface flow system.

Current work in progress to better define the groundwater flow regime and transport of strontium-90 in the 100-N Area has used the Ferris method to infer aquifer properties from cyclic water level fluctuations (Gilmore et al. 1992). The work compares the estimates for hydraulic conductivity that are derived from three independent methods: reinterpretation of pump test data, analysis of cyclic fluctuations in water levels, and application of the basic flow equation for groundwater. The goal of this work is to provide better hydraulic conductivity values for use in a numerical groundwater flow model.

Initial results suggest that the range of hydraulic conductivity values that is considered in the numerical model for the area can be reduced by applying these three methods. The results indicate that the 100-N Area can be divided into two regions, with hydraulic conductivities in the ranges of approximately 36 to 215 ft/day and 325 to 606 ft/day (Gilmore et al. 1992). These ranges are compared to the previously used range estimate of 104 to 8,400 ft/day. However, these results are tentative and the analysis is still in progress.

2.3 300 AREA STUDIES

Investigation of the interaction between Columbia River fluctuations and the water table underlying the 300 Area is being conducted as part of the compliance groundwater monitoring program for the 300 Area Process Trenches (C. R. Sherwood and D. R. Newcomer, Pacific Northwest Laboratory, personal communication). The study focuses on using statistical methods to describe the relationship between river stage and nearby groundwater levels, for the purpose of better understanding contaminant transport behavior. A variety of statistical analyses were performed on data collected at 15-minute intervals during 1987 and 1988. Auto-spectral and cross-spectral analysis methods are used to predict water levels in wells from river stage data. Continuing research suggests the possibility of using these methods to infer aquifer properties as well.

An extensive water level data collection program is currently in progress in the 300 Area as part of a CERCLA remedial investigation (DOE-RL 1990). Task 4c, "Hydraulic Properties," involves determining aquifer hydraulic properties to help understand the geohydrologic system, as well as the rate and direction of contaminant migration. Several methods are proposed for determining aquifer properties, including (1) single-well pumping and slug tests, (2) multiple well pumping tests and tracer tests, and (3) analysis of cyclic fluctuations in water levels in response to river stage changes. The analysis presented in Chapter 3.0 of this report uses data from the 300 Area data logger network.

2.4 OTHER RELATED INVESTIGATIONS

Aimo (1987) investigated the effect of a fluctuating stream on water quality in aquifers near the stream bank. He describes a flushing zone in which groundwater is diluted by the inflow of stream water. The extent of this flushing zone is controlled by (1) the volume of water involved in bank storage, (2) the relationship between aquifer diffusivity and the rate of rise in stream level, and (3) the magnitude and duration of stream level fluctuations. While his analysis of the problem does not result in estimates for aquifer hydraulic properties, it provides considerable insight into the problem of modeling the interaction between contaminated groundwater and an adjacent gaining stream.

3.0 APPLICATION OF FERRIS METHOD

Analysis of cyclic water level fluctuations in the Columbia River and adjacent groundwater monitoring wells was selected as the preferred method to test the feasibility of inferring aquifer hydraulic properties from water level data. This selection was prompted by the literature review of various methods (Appendix A); discussions with U.S. Geological Survey staff who have extensive experience with the subject (E. P. Weeks, U.S. Geological Survey, personal communication); and discussions with Hanford Site operable unit managers and their consultants.

Tri-Party Agreement milestone M-30-04 pertains to 100 Aggregate Area investigations. However, no water level data for the river and nearby wells in the 100 Areas had been acquired by the time of this analysis that fit the requirements for analysis by the Ferris method. Consequently, to meet the intent of the milestone by its due date, water level data from the 300 Area were used to investigate the feasibility of applying the Ferris method.

3.1 INTRODUCTION

The diffusivity of the shallow unconfined aquifer adjacent to the Columbia River in the 300 Area was estimated using the Ferris stage ratio and lag time methods (Ferris 1952, 1963). Diffusivity of an aquifer is defined as the aquifer transmissivity divided by its storativity. Pressure transducers installed in several groundwater wells and a river stage recorder in the 300 Area provided simultaneous hourly measurements of water levels. River stage fluctuations were approximated as simple harmonic motion. The attenuation and lag time between the river stage sinusoid and the response in the groundwater wells were measured. Using these data, aquifer diffusivity was estimated. The analysis was restricted to wells within approximately 1,700 ft of the river, since the groundwater response to the daily river stage fluctuations is damped out beyond that distance. The travel time of the wave front downstream along the river was assumed to be negligible compared to the lag time through the aquifer.

3.2 THE FERRIS METHOD

Ferris (1952, 1963) described two methods for estimating aquifer diffusivity from the groundwater response to river stage fluctuations. The equations were adapted from an analogous analysis of heat flow through solids and were derived for one-dimensional flow to a fully penetrating stream that is freely connected to the aquifer. When river stage fluctuations are approximated by simple harmonic motion, he calculated diffusivity from either (1) the attenuation between the river stage amplitude and the amplitude of corresponding water level fluctuations in nearby groundwater wells, or (2) the time lag between the river stage oscillations and the corresponding signal arriving in the groundwater wells. Ferris (1952) measured the attenuation and time lag parameters from the relative maximum and minimum points of the river stage and groundwater well water level hydrographs. One problem with both of these methods is that they rely on individual points (the relative extrema) and thus, if data collection errors or noise influence these points, the final result might be strongly influenced by such deviations.

To alleviate this sensitivity, methods that utilize more of the data set have been developed to calculate the time lag and attenuation. Erskine (1991) applied a least-squares-fitting routine to adjusted piezometer readings to determine the time lag. He then used the tidal efficiency factor, which is defined as the ratio of the standard deviations of the well and river water level readings, to determine the signal attenuation.

Gilmore et al. (1992) applied a correlation procedure for calculating barometric efficiency (Clark 1967) to determine the analogous apparent tidal efficiency. This procedure has two advantages over the Erskine method: (1) water levels do not have to remain symmetrical about their means, and (2) the mean water level does not have to remain constant from period to period. This procedure is advantageous when a limited data set (i.e., few cycles and a limited number of observation wells) is available. It helps prevent outliers from overly influencing the result.

In the analysis that follows, diffusivity is calculated using peak-to-peak measurement data. This was done because several cyclic data sets are available and the undue influence of potential outlier data is thereby avoided. For comparison, however, the correlation procedure used by Gilmore et al. (1992) was also applied to a sample data set.

3.3 WATER LEVEL DATA USED IN THE ANALYSIS

Three lines, each consisting of two wells, are included in the analysis (a location map is shown in Figure 3-1). The first line is formed by wells 399-1-7 and 399-1-2 and is located near the northern edge of the 300 Area. The second line, wells 399-3-9 and 399-3-12, runs between the 316-1 and 316-3 facilities, which is approximately 2,300 ft south of the first line. The third line, wells 399-4-9 and 399-4-1, is located approximately 1,100 ft south of the second line. The SWS-1 river stage recorder is located approximately 200 ft from well 399-4-9 and provides river stage fluctuation data for all three lines of wells.

The water level data analyzed come from two time intervals: May 17-21, 1992 and May 25-30, 1992. It was during these intervals that the data exhibited the cyclic behavior that meet the assumptions inherent in the Ferris methods. Furthermore, the stage ratios, time lags, and apparent tidal efficiencies were measured relative to two references: river stage recorder SWS-1 and monitoring well 399-4-9. All of the hydrographs, along with corresponding stage ratio, time lag, and apparent tidal efficiency measurements, are presented in Appendix C. An example hydrograph for the northernmost line of wells is shown in Figure 3-2. The stage ratios and time lags measured from this hydrograph are tabulated in Table 3-1. Table 3-2 contains the stage ratios, time lags, and apparent tidal efficiencies (determined from plots also found in Appendix C) for all of the data sets. Example plots for the stage ratio (logarithmic) and time lag (linear) measurements, from which input variables for the Ferris equation are obtained, are presented in Section 3.4, "Results" (see Figure 3-3).

Figure 3-1. Location Map for 300 Area Wells and River Stage Recorders.

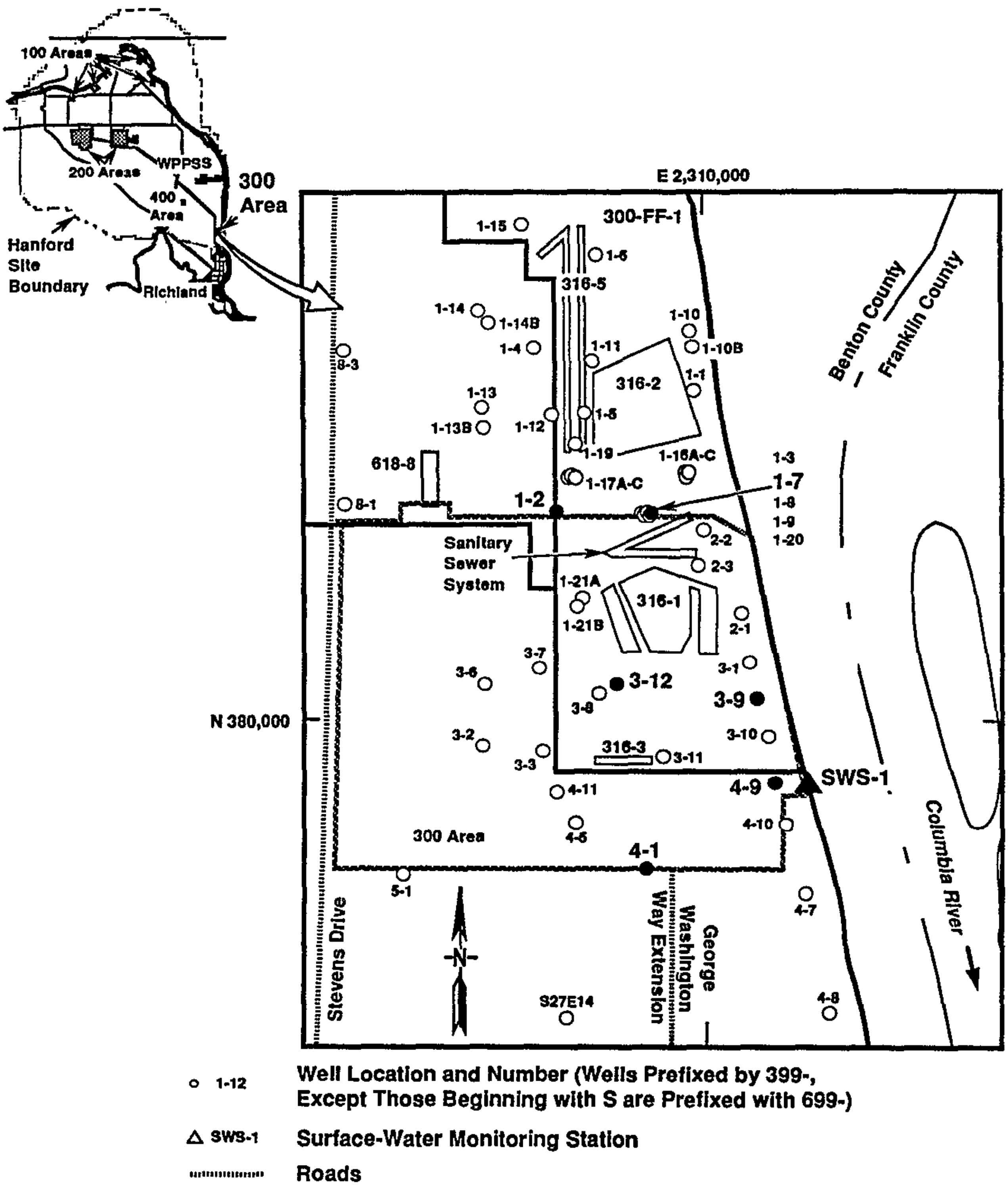
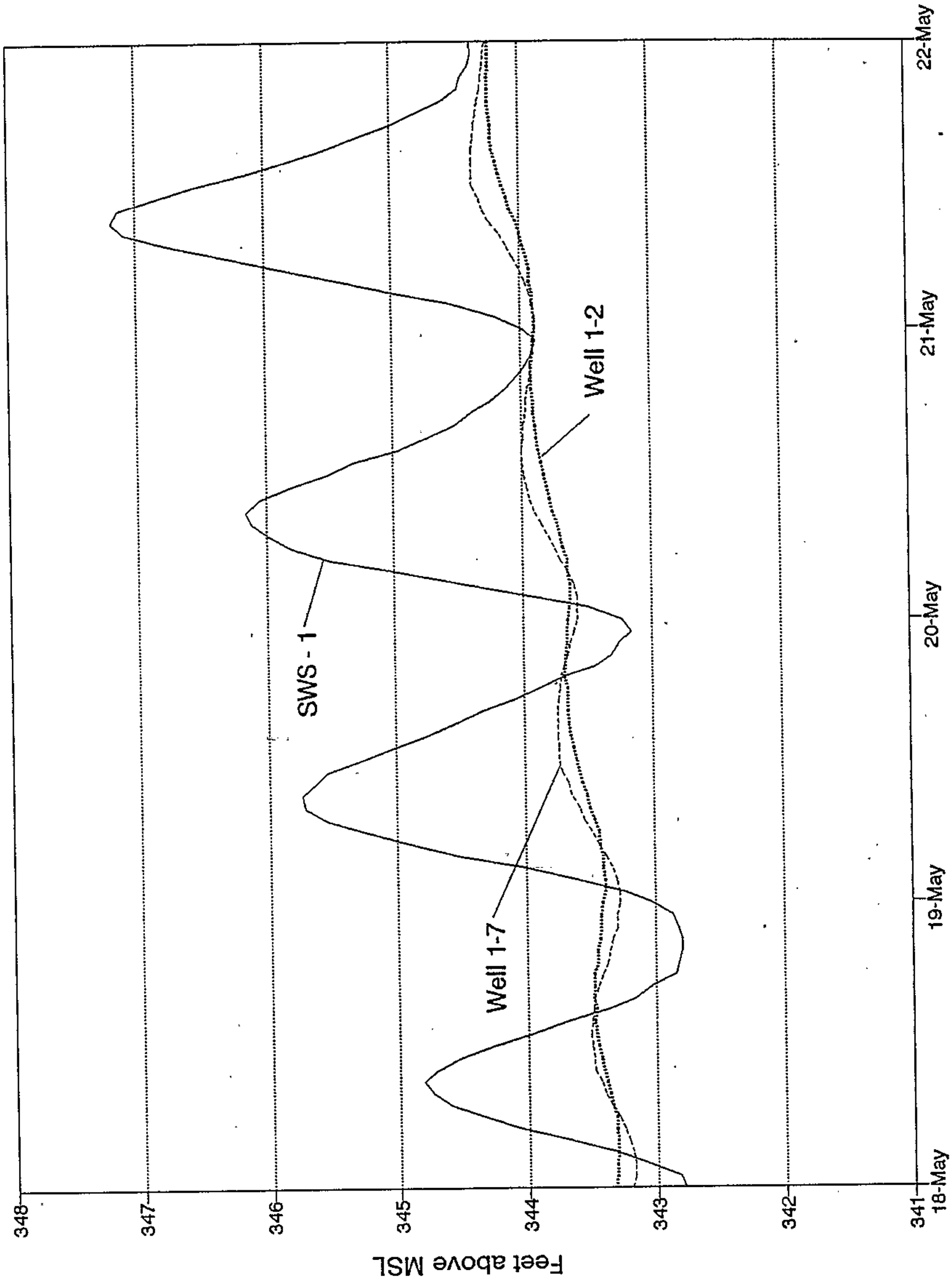


Figure 3-2. Example Hydrographs for Northern Line of Wells.



9 3 1 2 7 5 2 0 9 1 9

Table 3-1. Example Data Table for Water Level, Stage Ratio, and Lag Time Measurements for Northern Line of Wells.

May 17-21

	SWS - 1 Elevation (feet)	Change (feet)	Well 399-1-7 Elevation (feet)	Change (feet)	Change Ratio	Well 399-1-2 Elevation (feet)	Change (feet)	Change Ratio (feet)
Minimum	342.64		343.17			343.3		
Maximum	344.8	2.16	343.51	0.34	0.157407	343.47	0.17	0.078704
Minimum	342.8	-2	343.27	-0.24	0.12	343.39	-0.08	0.04
Maximum	345.74	2.94	343.74	0.47	0.159864	343.68	0.29	0.098639
Minimum	343.16	-2.58	343.57	-0.17	0.065891	343.63	-0.05	0.01938
Maximum	346.15	2.99	343.99	0.42	0.140468	343.92	0.29	0.09699
Minimum	343.88	-2.27	343.88	-0.11	0.048458	343.9	-0.02	0.008811
Maximum	347.18	3.3	344.36	0.48	0.145455	344.24	0.34	0.10303
Avg. Rising Limbs					0.150799			0.094341
Avg. Falling Limbs					0.078117			0.02273
Overall Avg.					0.119649			0.063651

	SWS - 1 Date	Hour	Well 399-1-7 Hour	Lag Time (days)	Well 399-1-2 Hour	Lag Time (days)	Lag Time Between Wells (days)
Minimum	May 17	1800	2600	0.333333	2700	0.375	0.041667
Maximum	May 18	900	1300	0.166667	1600	0.291667	0.125
Minimum	May 18	2000	2400	0.166667	2600	0.25	0.083333
Maximum	May 19	900	1500	0.25	2200	0.541667	0.291667
Minimum	May 19	2300	2500	0.083333	2600	0.125	0.041667
Maximum	May 20	900	1300	0.166667	1900	0.416667	0.25
Minimum	May 20	2300	2400	0.041667	2400	0.041667	0
Maximum	May 21	900	1300	0.166667	2200	0.541667	0.375
Avg. Lag Time (days)				0.171875		0.322917	0.151042

Period = ~ 1 day

Distance (River to 399-1-7) 700 feet
Distance (River to 399-1-2) 1400 feet
Distance (399-1-7 to 399-1-2) 700 feet

Table 3-2. Stage Ratios, Lag Times, and Apparent Tidal Efficiencies, Measured Relative to (a) River Stage Recorder SWS-1 and (b) Well 399-4-9.

Well number	Distance from source (ft)	Stage ratio		Time lag ratio (days)		Efficiency ratio
		May 17-21, 1992	May 25-29, 1992	May 17-21, 1992	May 25-29, 1992	May 17-21, 1992
(a) Source: River Stage Record SWS-1						
399-1-2	1,400	0.064	0.057	0.323	0.271	0.033
399-1-7	700	0.120	0.156	0.172	0.177	0.093
399-3-12	1,200	0.088	0.086	0.292	0.255	0.054
399-3-9	200	0.139	0.141	0.167	0.146	0.112
399-4-1	1,400	0.152	0.101	0.276	0.245	0.057
399-4-9	300	0.102	0.156	0.146	0.125	0.111
(b) Source: Well 399-4-9						
399-1-2	1,100	0.668	0.644	0.177	0.146	0.355
399-1-7	400	0.785	0.997	0.026	0.052	0.756
399-3-12	900	0.574	0.548	0.146	0.130	0.546
399-4-1	1,100	0.417	0.363	0.130	0.120	0.598

3.4 RESULTS

The northernmost well line (399-1-7 and 399-1-2) shows the greatest uniformity of results. Diffusivity values are all within the same order of magnitude--10^6 ft^2/day (Table 3-3). The agreement between the two methods for the entire line is very good and averages 2.2 x 10^6 ft^2/day, with no individual measurement differing from the average by more than a factor of two. The stage ratio method yields a higher diffusivity value for the data collected between May 17-21 than for that collected between May 25-30, but the reverse is true for the lag time method.

Examining the individual segments of the line using the lag time method showed that the region between the river and well 399-1-7 has a lower diffusivity than either the segment between the two wells or the entire line, whereas the region between the wells has the highest diffusivity. This pattern holds true for both periods of time. As stated previously, however, the differences are relatively small and may not be significant with respect to conclusions regarding variations in aquifer properties.

Table 3-3. Results of Stage Ratio and Lag Time Methods for Calculating Aquifer Diffusivity for the 399-1-7 and 399-1-2 Well Line.

Stage ratio method		
Region	Time span	Diffusivity (ft^2/day)
River => 399-1-2	May 17-21	3.84×10^6
River => 399-1-2	May 25-29	1.50×10^6
Lag time method		
River => 399-1-2	May 17-21	1.50×10^6
River => 399-1-7	May 17-21	1.32×10^6
399-1-7 => 399-1-2	May 17-21	1.71×10^6
River => 399-1-2	May 25-29	2.13×10^6
River => 399-1-7	May 25-29	1.24×10^6
399-1-7 => 399-1-2	May 25-29	4.44×10^6

The two southernmost lines exhibit much greater variability in diffusivity values, although both lines follow similar patterns (Tables 3-4 and 3-5). The stage ratio method produces diffusivity values for the entire lines one order of magnitude higher than the lag time method (10^7 versus 10^6 ft^2/day). Like the northernmost line, the stage ratio method yields higher diffusivities using the May 17-21 data than the May 25-29 data, with the reverse being true for the lag time method. Comparing the individual segments along both lines also reveals the same similarities. The segment nearest the river has the lowest diffusivity (in both cases by an order of magnitude) than either the diffusivity along the entire line or the region between the wells. On this basis, it can be surmised that the aquifer diffusivity nearest the river (i.e., within 400 ft) is significantly lower than that farther inland (beyond 1,000 ft).

Treating the six wells as a composite, the two methods produce overall diffusivity values for the area that differ by less than an order of magnitude (Table 3-6). The results also remain fairly constant for the two time periods. The agreement between these methods is in contrast to the results from the two southernmost lines of wells, which showed disparity between the two methods. Based on the agreement shown by the two methods for the composite, the disparity may be due to the small number of wells (i.e., two) contained in the individual lines. However, the values determined for the northern line of wells are in agreement with the composite values.

Table 3-4. Results of Stage Ratio and Lag Time Methods for Calculating Aquifer Diffusivity for the 399-3-9 and 399-3-12 Well Line.

Stage ratio method		
Region	Time span	Diffusivity (ft^2/day)
River => 399-3-12	May 17-21	1.46×10^7
River => 399-3-12	May 25-29	1.26×10^7
Lag time method		
River => 399-3-12	May 17-21	1.35×10^6
River => 399-3-9	May 17-21	1.15×10^5
399-3-9 => 399-3-12	May 17-21	5.09×10^6
River => 399-3-12	May 25-29	1.76×10^6
River => 399-3-9	May 25-29	1.50×10^5
399-3-9 => 399-3-12	May 25-29	6.65×10^6

Table 3-5. Results of Stage Ratio and Lag Time Methods for Calculating Aquifer Diffusivity for the 399-4-9 and 399-4-1 Well Line.

Stage ratio method		
Region	Time span	Diffusivity (ft^2/day)
River => 399-4-1	May 17-21	2.32×10^7
River => 399-4-1	May 25-29	1.95×10^7
Lag time method		
River => 399-4-1	May 17-21	2.05×10^6
River => 399-4-9	May 17-21	3.37×10^5
399-4-9 => 399-4-1	May 17-21	5.68×10^6
River => 399-4-1	May 25-29	2.60×10^6
River => 399-4-9	May 25-29	4.58×10^5
399-4-9 => 399-4-1	May 25-29	6.71×10^6

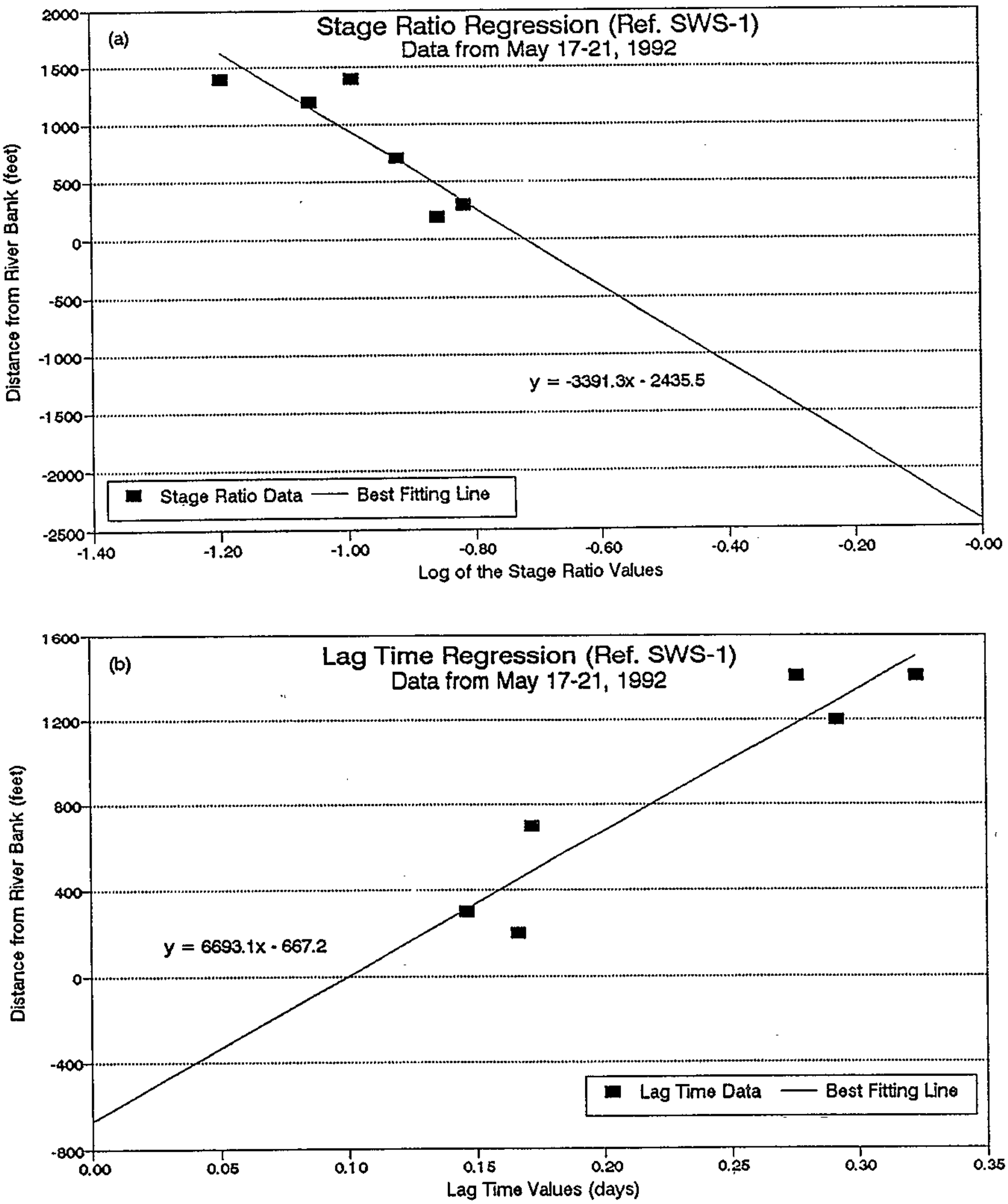
Table 3-6. Results of Stage Ratio and Lag Time Methods for Calculating Aquifer Diffusivity for the Composite Group of Wells.

Composite stage ratio method				
Date	Pulse reference	Regression line slope	Regression line intercept	Diffusivity (x 10^6 ft^2/day)
May 17-21	SWS-1	-3391.4	-2435.5	6.77
May 25-29	SWS-1	-2559.7	-1598.2	3.85
May 17-21	399-4-9	-1974.3	430.3	2.29
May 25-29	399-4-9	-1496.6	540.4	1.32
Composite time lag method				
May 17-21	SWS-1	6693.1	-667.2	3.56
May 25-29	SWS-1	8509.1	-861.7	5.76
May 17-21	399-4-9	4698.7	312.1	1.76
May 25-29	399-4-9	7466.7	38.9	4.44
Efficiency ratio method				
May 17-21	SWS-1	-2372.8	-1872.9	3.34
May 17-21	399-4-9	-1780.8	404.2	1.88

The high magnitude of the y-intercept values associated with the calculated regression lines (Figure 3-3) indicates that the river stage recorder location (SWS-1) may not be truly representative of the origin of the pressure wave within the aquifer. Ideally, the regression lines for both methods should pass through the origin. For the stage ratio method, the origin represents zero signal attenuation at zero distance. For the time lag method, it represents zero time lag at zero distance. By failing to pass through the origin, the calculated lines indicate that because of conditions along the riverbank, the pressure waves through the aquifer do not originate at the river/aquifer interface or the actual hydrologic conditions are not amenable to the model assumptions. That is, actual hydrologic conditions do not meet the assumptions required by the Ferris method.

To test the hypothesis that river/aquifer interface anomalies are influencing the calculations, a well near the river was substituted as the source of the pressure pulse. Well 399-4-9 was selected for this purpose, because its water level response had the shortest average lag time with the river. The time lags and stage ratios were then calculated relative to this well. (Although well 399-3-9 is closer to the riverbank, it had a greater lag time than 399-4-9, indicating a potentially anomalous connection to the river.)

Figure 3-3. Semilogarithmic Plots for (a) Composite Stage Ratio Data and (b) Composite Lag Time Data.



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Using the data from well 399-4-9 as the source or input signal resulted in a substantial decrease in the magnitude of the y-intercept values of the regression lines. Unfortunately, the values still remain too high to be considered acceptable and remain on the order of the distances between the wells. The intercept numbers would have to be much closer to zero before the calculated diffusivity values can be viewed with confidence.

If the magnitude of the regression y-intercept value is indicative of model reliability, then the time lag method proves to be the more reliable method of determining aquifer diffusivity. In all comparable cases, the time lag method produces regression lines with y-intercept values less than the stage ratio method, usually by at least an order of magnitude.

Using the efficiency method described in Clark (1967) does not affect the diffusivity calculations significantly. The efficiency values determined by this technique yield diffusivity values that are essentially indistinguishable from values determined using only the relative extrema data points from the well hydrographs. Furthermore, this method is plagued by extremely high y-intercept values on its regression lines. Because this technique does not appear to offer any advantage to the analysis, it was only applied to the first data set (May 17-21, 1992) for use as a comparison.

3.5 DISCUSSION

Numerous factors may contribute to the variability in the calculated diffusivity values and high y-intercept values of the regression lines. These factors include:

- Heterogeneity in aquifer lithology, geometry, or structure, resulting in the transmissivity and storativity to be highly variable in the space domain of the river and wells.

- Hydraulic gradients, both transitory and permanent, that are neither normal to nor directed toward the river.

- Curvilinear flow lines as a result of the partial penetration of the stream, causing a violation of the one-dimensional flow assumption.

- Aquifer response to each wave may be dependant on the magnitude and duration of previous waves or other preceding trends in groundwater elevation.

- Variable aquifer thickness resulting from the passage of the pressure wave--transmissivity is proportional to aquifer thickness and therefore becomes variable.

- Specific yield and storativity relate to two distinctly different phenomena but are used interchangeably in the governing equation for the method.

Any or all of these factors may be responsible for the variability observed in the diffusivity values. Attempts to filter the influence of these factors from the true aquifer characteristics may not be possible.

The diffusivity values obtained from the two Ferris methods are on the order of 10^6 ft^2/day. Assuming a storativity value of 0.1 results in transmissivity values on the order of 10^5 ft^2/day. This range in transmissivity is consistent with values reported by Spane (1991) and Bierschenk (1959) for pumping tests conducted in the 300 Area. These tests were conducted in several wells that are screened in the shallow unconfined aquifer. A summary of these test results is presented in Table 3-7.

Table 3-7. Summary of Pumping Tests Results.

Well number (300 Area)	Transmissivity (x 10^5 ft^2/day)	
	Spane (1991)	Bierschenk (1959)
399-1-13	1.1	
399-1-18A	10.0	
399-1-14	1.9	
399-1-10	2.0	
399-1-16A	0.1	
399-3-2		4.3
399-3-6		8.5
399-3-7		15.0

Because the Ferris methods indicate transmissivity values within the range of the pumping tests, it is, perhaps, a reasonable method for estimating aquifer properties, given a suitable set of water level observations. Whether the model can be used to delineate aquifer heterogeneities or to provide more accurate estimates of aquifer properties than currently exist is still uncertain. This results from the need to combine data from several observation points, in order to create a composite plot for regression analysis. Spatial precision is thus sacrificed to obtain the improved reliability yielded by the regression analysis.

Transmissivity estimates can be derived from water level data where suitable fluctuation patterns exist. However, experience gained thus far on the Hanford Site suggests that obtaining suitable data sets is not guaranteed for any particular region. For example, the three lines of data loggers at 100-B, 100-H, and 100-F, respectively, did not produce cyclic fluctuation records that can be interpreted with any confidence during approximately the first 6 months of their operation (Appendix B contains example records from those areas).

Where suitable cyclic fluctuation records exist, such as the 300 Area, the analysis methods described by Ferris (1952) do not appear to result in a significant improvement over other methods currently in use to estimate aquifer properties. While some agreement is present among the various methods, there is no improvement in spatial resolution as the result of applying the Ferris method. Using the Ferris method also requires a substantial commitment to gathering field data, data management, and data analysis to produce aquifer property information.

Information to support risk assessments involving contaminated groundwater must consider two geographic scales: (1) a source-specific scale, such as a localized plume at a reactor area, and (2) a regional scale. To predict the movement of contaminants at the source-specific scale, direct observation of plume movement using data from wells and shoreline seepage may prove to be the most effective and cost-efficient approach. Analysis of cyclic fluctuations in water levels does not appear to contribute to this objective, based on the results of this investigation.

Predictions of contaminant transport at a regional scale involve a numerical groundwater flow model. A risk assessment that considers the cumulative impact that contaminated groundwater from the northern Hanford Site has on river water quality is an example of this scale. The flow model uses aquifer hydraulic properties as variables, and the accuracy of the predictions will depend to some degree on the spatial distribution of aquifer properties over the region of interest. Analysis of water level fluctuations may improve information on this spatial variability. However, the improvement may be in the form of relative changes in aquifer properties rather than increased accuracy of actual values.

4.0 REFERENCES

Aimo, N. J., 1987, *Preliminary Numerical Study of the Effects of Fluctuating River Level on Solute Transport Within Riverbank Aquifers*, Master's Thesis, Washington State University, Pullman, Washington.

Bierschenk, W. H., 1959, *Aquifer Characteristics and Ground-Water Movement at Hanford*, HW-60601, June 1959, Hanford Atomic Products Operation, General Electric Company, Richland, Washington, pp. 29-31 and Appendix VI.

Brown, D. J. and P. P. Rowe, 1960, *100-N Aquifer Evaluation*, HW-67326, November 1960, Hanford Atomic Products Operation, General Electric Company, Richland, Washington. [Note: Hantush (1961) contains corrections to equations used in this report].

Clark, W. E., 1967, "Computing the Barometric Efficiency of a Well," *American Society of Civil Engineering, Journal of the Hydraulics Division*, Vol. HY 4, pp. 93-98.

DOE-RL, 1990, *Remedial Investigation/Feasibility Study Work Plan for the 300-FF-5 Operable Unit, Hanford Site, Richland, Washington*, DOE/RL 89-14, June 1990, U.S. Department of Energy, Richland Field Office, Richland, Washington.

DOE-RL, 1992, *Remedial Investigation/Feasibility Study Work Plan for the 100-BC-5 Operable Unit, Hanford Site, Richland, Washington*, DOE/RL 90-08, Rev. 0, July 1992, U.S. Department of Energy, Richland Field Office, Richland, Washington.

Ecology, EPA, and DOE, 1990, *Hanford Federal Facility Agreement and Consent Order*, Vol. 1 and 2, Washington State Department of Ecology, U.S. Environmental Protection Agency, and U.S. Department of Energy, Olympia, Washington.

Erskine, A. D., 1991, "The Effect of Tidal Fluctuation on a Coastal Aquifer in the UK," *Ground Water*, Vol. 29, No. 4, pp. 556-562.

Ferris, J. G., 1952, "Cyclic Fluctuations of Water Level as a Basis for Determining Aquifer Transmissibility," *U.S. Geological Survey, Ground-Water Note*, No. 1, April 1952, 17 pp.

Ferris, J. G., 1963, "Cyclic Water-Level Fluctuations as a Basis for Determining Aquifer Transmissibility," in R. Bentall (compiler), *"Methods of Determining Permeability, Transmissibility, and Drawdown," U.S. Geological Survey Water-Supply Paper 1536-I*, Washington, D.C., pp. 305-318.

Gilmore, T. J., F. A. Spane, Jr., D. R. Newcomer, and C. R. Sherwood, 1992, *Application of Three Aquifer Test Methods for Estimating Hydraulic Properties Within the 100-N Area*, PNL-8335, September 1992, Pacific Northwest Laboratory, Richland, Washington.

Hantush, M. S., 1961, "Discussion of Paper by P.P. Rowe, 'An Equation for Estimating Transmissibility and Coefficient of Storage From River Level Fluctuations,'" *Journal of Geophysical Research*, April 1961, Vol. 66, No. 2, pp. 1310-1311.

Newcomb, R. C. and S. G. Brown, 1961, "Evaluation of Bank Storage Along the Columbia River Between Richland and China Bar, Washington," *Geological Survey Water-Supply Paper 1539-I*, U.S. Geological Survey, Washington, D.C.

Newcomer, D. R., 1988, *Detailed Water-Level Data and an Evaluation of Mathematical Approaches for Near-River Monitoring*, Master's Thesis, Montana College of Mineral Science and Technology, Butte, Montana.

Poeter, E. and D. R. Gaylord, 1990, "Influence of Aquifer Heterogeneity on Contaminant Transport at the Hanford Site," *Ground Water*, Vol. 28, No. 6, November-December 1990, pp. 900-909.

Rowe, P. P. 1960, "Equations for Estimating Transmissibility and Coefficient of Storage from River Level Fluctuations," *Journal of Geophysical Research*, October 1960, Vol. 65, No. 10, pp. 3419-3424. [Hantush (1961) contains corrections to these equations]

Spane, F. A., 1991, "Evaluation of Pumping Test Analyses Reported in Schalla et al. (1988)," Letter Report, August 13, 1991, Pacific Northwest Laboratories to Westinghouse Hanford Company.

APPENDIX A

REVIEW OF METHODS FOR ESTIMATING AQUIFER HYDRAULIC PROPERTIES USING AQUIFER/STREAM INTERACTION CHARACTERISTICS

Prepared by

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APPENDIX A

REVIEW OF METHODS FOR ESTIMATING AQUIFER HYDRAULIC PROPERTIES USING AQUIFER/STREAM INTERACTION CHARACTERISTICS

W. J. McMahon
A. G. Law

The importance of aquifer/stream interaction was recognized as early as the seventeenth century (Spiegel 1962 contains a comprehensive list of early references). Only during the last 40 yr have efforts to estimate aquifer hydraulic properties from aquifer/stream interaction data gained momentum. The primary equations used in this research were developed either for flow in a confined aquifer or for flow through an unconfined aquifer utilizing the Dupuit-Forcheimer assumptions (Jacob 1950).

A.1 CONFINED AQUIFERS

For confined aquifers with a fully penetrating stream, the basic differential equation for one-dimensional flow is:

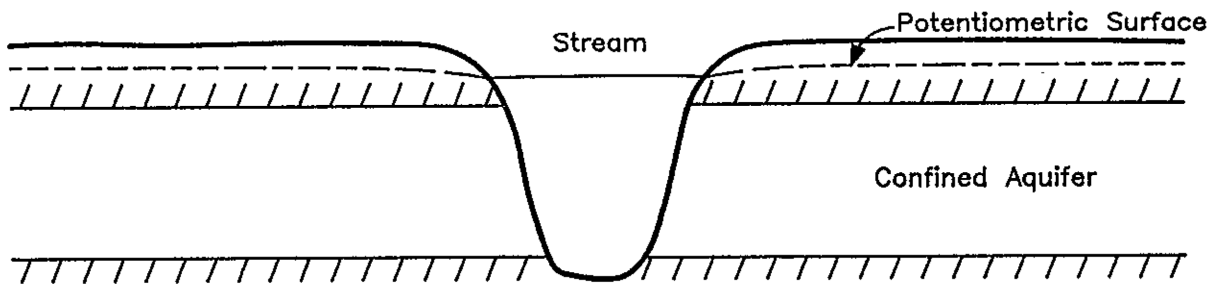
$$\frac{\partial^2 h}{\partial x^2} = \frac{S}{T}\frac{\partial h}{\partial t} \qquad (1)$$

where
- h = hydraulic head
- x = distance
- S = storativity
- T = transmissivity
- t = time.

This equation is linear with the resulting advantage that the principle of superposition holds, allowing for the linear combination of solutions to different boundary conditions. A physical analog for this differential equation is shown in Figure A-1. If the stream were only partially penetrating, the flow would no longer be one-dimensional and another term, $\partial^2 h/\partial z^2$, would need to be added to the left side of equation (1) to account for the resulting vertical component of flow upward into the bottom of the stream.

Many of the solutions presented in the literature for estimating aquifer parameters are solutions to equation (1) for particular boundary conditions. Ferris (1952; also 1951 and 1963) developed equations for estimating aquifer diffusivity that are based on a simple harmonic model of surface water stage elevation (diffusivity is defined as transmissivity divided by storativity). Cooper and Rorabaugh (1963) developed similar equations for more general flood-wave stage oscillations. Rowe (1960; correction by Hantush 1961) determined an equation for aquifer diffusivity for a linear change in the water surface elevation of a stream. Hantush (1961) also presented a solution

Figure A-1. Physical Analog for One-Dimensional Flow Through a Confined Aquifer Having a Fully Penetrating Stream.



for the general case of a fluctuating stream, using the technique of convolution. Note that the results given by these authors are all solutions to the one-dimensional equation (1), where a confined aquifer is fully penetrated by a stream.

Using an approach similar to that used for aquifer tests of a partially penetrating well, Neuman (1974) applied equation (1) to a well located far enough from the stream so that the effects of partial penetration are not apparent. Pinder et al. (1969) developed a curve-matching technique that involved creating discrete time intervals in the stage hydrograph, and summing the influence of each increment. The solution for transmissivity comes from the equation for the best-fitting theoretical response curve to the observed data.

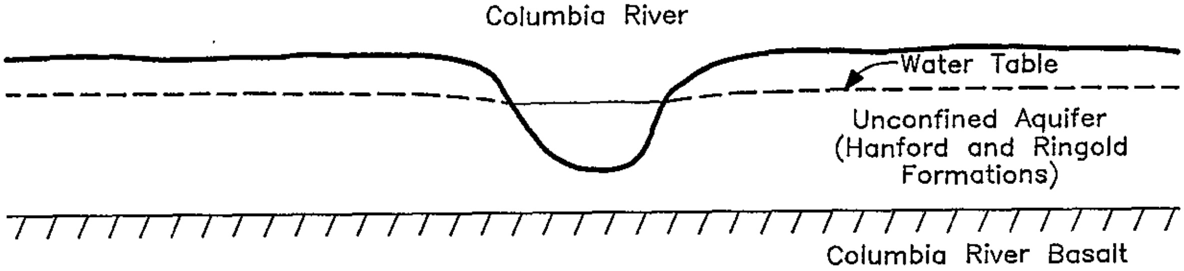
A.2 UNCONFINED AQUIFERS

Along the Columbia River shoreline on the Hanford Site, aquifer/river interaction involves an unconfined aquifer that is partially penetrated by a stream (Figure A-2). Previous research for flow in this type of interaction has utilized the Dupuit-Forcheimer assumptions (Jacob 1950). These assumptions include (1) a homogeneous and isotropic unconfined aquifer, (2) horizontal flow toward the stream, and (3) that through a vertical plane oriented perpendicular to the direction of groundwater flow, the hydraulic gradient is uniform from top to bottom of the aquifer, and equal to the slope of the water table. The saturated thickness of the aquifer is assumed to remain constant over time and space. The storage term is considered to be interchangeable for both the confined and unconfined flow cases. The Dupuit-Forcheimer assumptions are based on small differences in the saturated thickness of the aquifer and thus do not address the matter of partial penetration of the stream.

For an unconfined aquifer, an equation corresponding to equation (1) is presented in Jacob (1950), as follows:

$$\frac{\partial^2 h^2}{\partial x^2} = \frac{2S_y}{K}\frac{\partial h}{\partial t} \qquad (2)$$

Figure A-2. Physical Analog for the Flow Regime Along the Hanford Site Shoreline of the Columbia River.



where S_y is specific yield, K is hydraulic conductivity, and the remaining variables are the same as equation (1). This equation is often referred to as the Boussinesq equation, and its development includes the Dupuit-Forcheimer approximations, which introduce some degree of uncertainty into the reliability of the equation. The equation is nonlinear, making it more difficult to obtain solutions for specified boundary conditions. An additional complication is that S_y (specific yield of an unconfined aquifer) in equation (2) is known to vary with time. In contrast, the variable S (storativity of a confined aquifer) in equation (1) may be assumed to remain constant with time. Equation (2) is also a one-dimensional equation, which limits its use to fully penetrating streams.

Hornberger et al. (1970) recognized this formidable problem and developed a finite difference, predictor-corrector technique for evaluating equation (2). The technique was validated by comparing the results to analytical solutions of the Boussinesq equation. It was also compared to the results using the method of Cooper and Rorabaugh (1963), and reasonable agreement was achieved for confined aquifer discharge to a fully penetrating stream when a sinusoidal flood peak was 1.5 times greater than the initial stage of the stream. However, when the flood peak was 2 or 3 times the initial stage, the results were not as good. An additional limitation is noted by Hornberger et al. (1970, referencing Muskat 1937) in that for the steady-state analog of equation (2), the assumptions necessary for derivation of the equation do not permit an accurate determination of the elevation of the water table.

Zitta and Wiggert (1971) found a simultaneous solution for the Boussinesq equation and an equation that describes the volume of water displaced in the aquifer by fluctuations in the water table. Erskine (1991) concluded that analytical solutions for groundwater surfaces are difficult to obtain. He also noted that variations in storage coefficients calculated from time lag and tidal efficiency data result from the inadequate representation of the unconfined aquifer in the mathematical equations. Spiegel (1962) presented a modification of the differential equation for the unconfined flow, although no solutions were presented for any specified boundary conditions.

Reynolds (1987) applied the method of Pinder et al. (1969) to three well sites along the Tioughnioga River near Cortland, New York. He reasoned that the method worked better for the confined than the unconfined aquifer because the saturated thickness (and transmissivity) of the confined aquifer remained unchanged by either a flood wave or areal recharge from precipitation.

A.3 SUMMARY

Review of the literature indicates that confined aquifer parameters can be estimated reasonably well from the aquifer response to hydraulically connected flood waves or varying surface water levels. However, application of these confined aquifer equations to unconfined aquifers is much less reliable, since the assumptions in the derivation of the basic flow equations are not fulfilled. Uncertainty resulting from the effect that specific yield has on the solution of the flow equation also requires further investigation.

Very little of the published research explicitly addresses calculating unconfined aquifer properties using methods related to stream interaction. Almost all methods described are derived for confined aquifers and utilize the Boussinesq equation, or include the assumptions inherent in that equation.

The Ferris method is an analytical solution to the differential equation that describes flow through a confined aquifer, with potential application to an unconfined aquifer, due to sinusoidal elevation changes in a nearby stream. The extensive water level data collection effort underway at the Hanford Site creates a unique opportunity to further test the Ferris method.

A.4 REFERENCES

Cooper, H. H., Jr. and M. I. Rorabaugh, 1963, "Ground-Water Movements and Bank Storage Due to Flood Stages in Surface Streams," *Geological Survey Water-Supply Paper 1536-J*, pp. 343-366.

Erskine, A. D., 1991, "The Effect of Tidal Fluctuation on a Coastal Aquifer in the UK," *Ground Water*, Vol. 29, No. 4, pp. 556-562.

Ferris, J. G., 1951, "Cyclic Fluctuations of Water Level as a Basis for Determining Aquifer Transmissibility," *International Association of Scientific Hydrology*, Publication 33, pp. 148-155.

Ferris, J. G., 1952, "Cyclic Fluctuations of Water Level as a Basis for Determining Aquifer Transmissibility," *U.S. Geological Survey, Ground-Water Note*, No. 1, April 1952.

Ferris, J. G., 1963, "Cyclic Water-Level Fluctuations as a Basis for Determining Aquifer Transmissibility," in R. Bentall (compiler), *"Methods of Determining Permeability, Transmissibility, and Drawdown," U.S. Geological Survey Water-Supply Paper 1536-I*, Washington, D.C., pp. 305-318.

Hantush, M. S., 1961, "Discussion of Paper by P.P. Rowe, 'An Equation for Estimating Transmissibility and Coefficient of Storage From River Level Fluctuations,'" *Journal of Geophysical Research*, April 1961, Vol. 66, No. 2, pp. 1310-1311.

Hornberger, G. M., J. Ebert, and I. Remson, 1970, "Numerical Solution of the Boussinesq Equation for Aquifer-Stream Interaction," *Water Resources Research*, Vol. 6, No. 2, pp. 601-608.

Jacob, E. C., 1950, "Flow of Ground Water," in H. Rouse, (ed.), *Engineering Hydraulics*, John Wiley, New York, pp. 321-386.

Muskat, M. K., 1937, *Flow of Homogeneous Fluids Through Porous Media*, McGraw-Hill, New York.

Neuman, S. P., 1974, "Effect of Partial Penetration on Flow in Unconfined Aquifers Considering Delayed Gravity Response," *Water Resources Research*, Vol. 10, pp. 303-312.

Pinder, G. F., J. D. Bredehoeft, and H. H. Cooper, 1969, "Determination of Aquifer Diffusivity from Aquifer Response to Fluctuations in River Stage," *Water Resources Research*, Vol. 5, No. 4, pp. 850-855.

Reynolds, R. J., 1987, "Diffusivity of a Glacial-Outwash Aquifer by the WYLW (Floodwave) Response Technique," *Ground Water*, Vol. 25, No. 3, pp. 290-299.

Rowe, P. P., 1960, "Equations for Estimating Transmissibility and Coefficient of Storage from River Level Fluctuations," *Journal of Geophysical Research*, October 1960, Vol. 65, No. 10, pp. 3419-3424. [Hantush (1961) contains a correction for these equations]

Spiegel, Z., 1962, *Hydraulics of Certain Stream-Connected Aquifer Systems*, New Mexico State Engineer, Special Report, Santa Fe, New Mexico.

Zitta, V. L. and J. M. Wiggert, 1971, "Flood Routing in Channels With Bank Seepage," *Water Resources Research*, Vol. 7, No. 5, pp. 1341-1345.

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APPENDIX B

100 AGGREGATE AREA DATA LOGGER PROGRAM

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APPENDIX B

100 AGGREGATE AREA DATA LOGGER PROGRAM

Pressure transducers and data loggers are installed in wells in the 100-B, 100-H, and 100-F Areas. River stage is also recorded at these areas using pressure transducers and data loggers. This equipment has been in operation since late 1991 and records water levels at 1-hr intervals. Digital data are sent from the field installations via radiotelemetry to a computer located in North Richland, where the data are entered into a database maintained by the Westinghouse Hanford Company Geosciences Group. A description of the entire system is presented in Campbell and Newcomer (1992).[1] Similar equipment is installed in numerous wells in the 300 Area. A manual, analog river stage recorder is in operation at 100-N Area; these records are not currently included in the electronic database.

Water level data produced by these installations provide information on the landward extent of water table fluctuations caused by the daily and seasonal rise and fall of the Columbia River. The data also help describe the elevation range of the soil column that is alternately wetted and drained. The original purpose for these installations included obtaining a data set that could be analyzed to infer aquifer hydraulic properties. All of these data objectives pertain to the interaction between Hanford Site groundwater and the Columbia River, an important topic related to environmental restoration decisions for the Hanford Site.

Location maps, historical water levels, and example data logger records for 100-B, 100-H, and 100-F areas are shown in Figures B-1 through B-9.

[1]Campbell, M. D. and D. R. Newcomer, 1992, *Automatic Measurement of Water Levels Within the 300-FF-5 Boundary*, PNL-7874, April 1992, Pacific Northwest Laboratory, Richland, Washington.

Figure B-1. Location Map for Water Level Recorders in the 100-B Area.

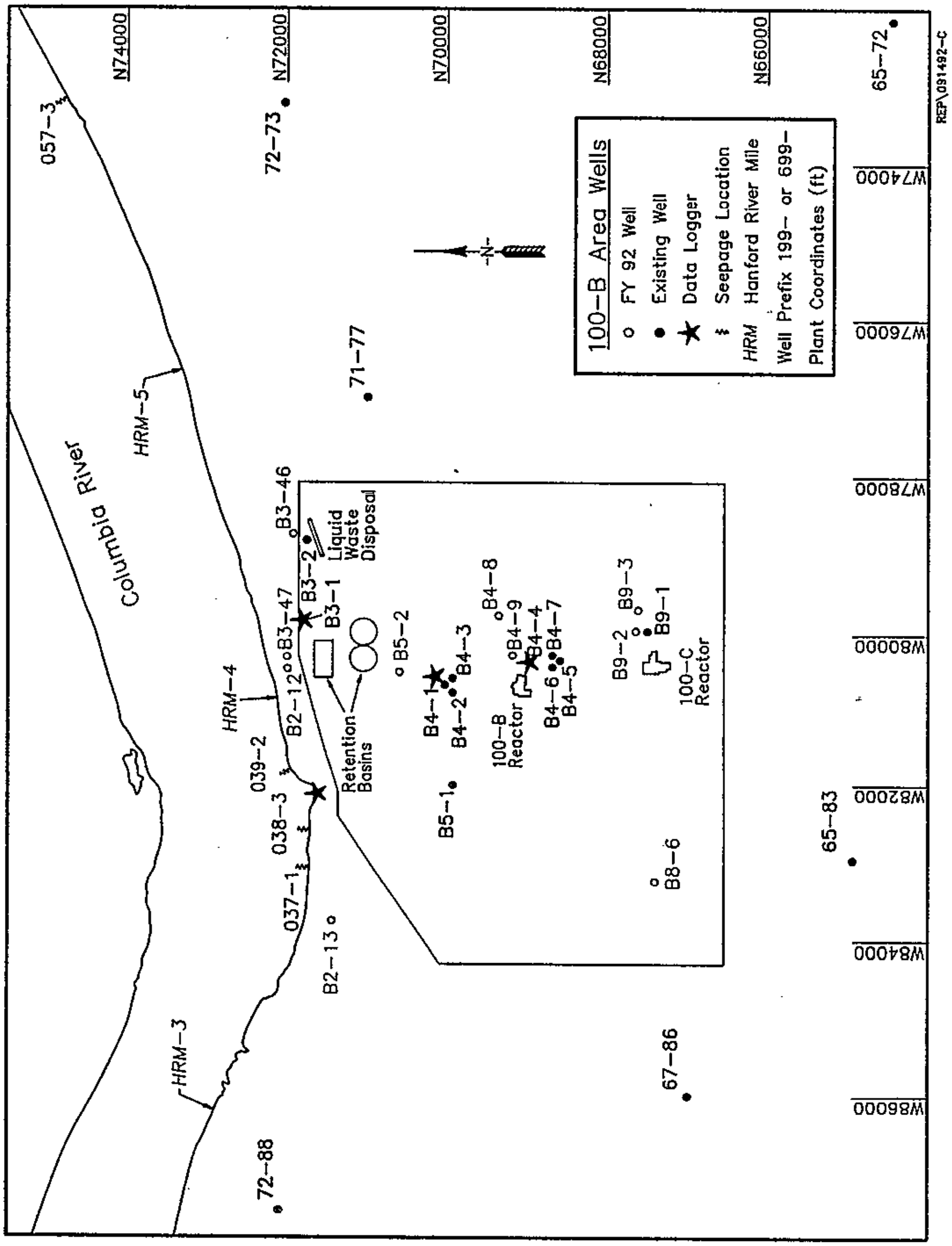

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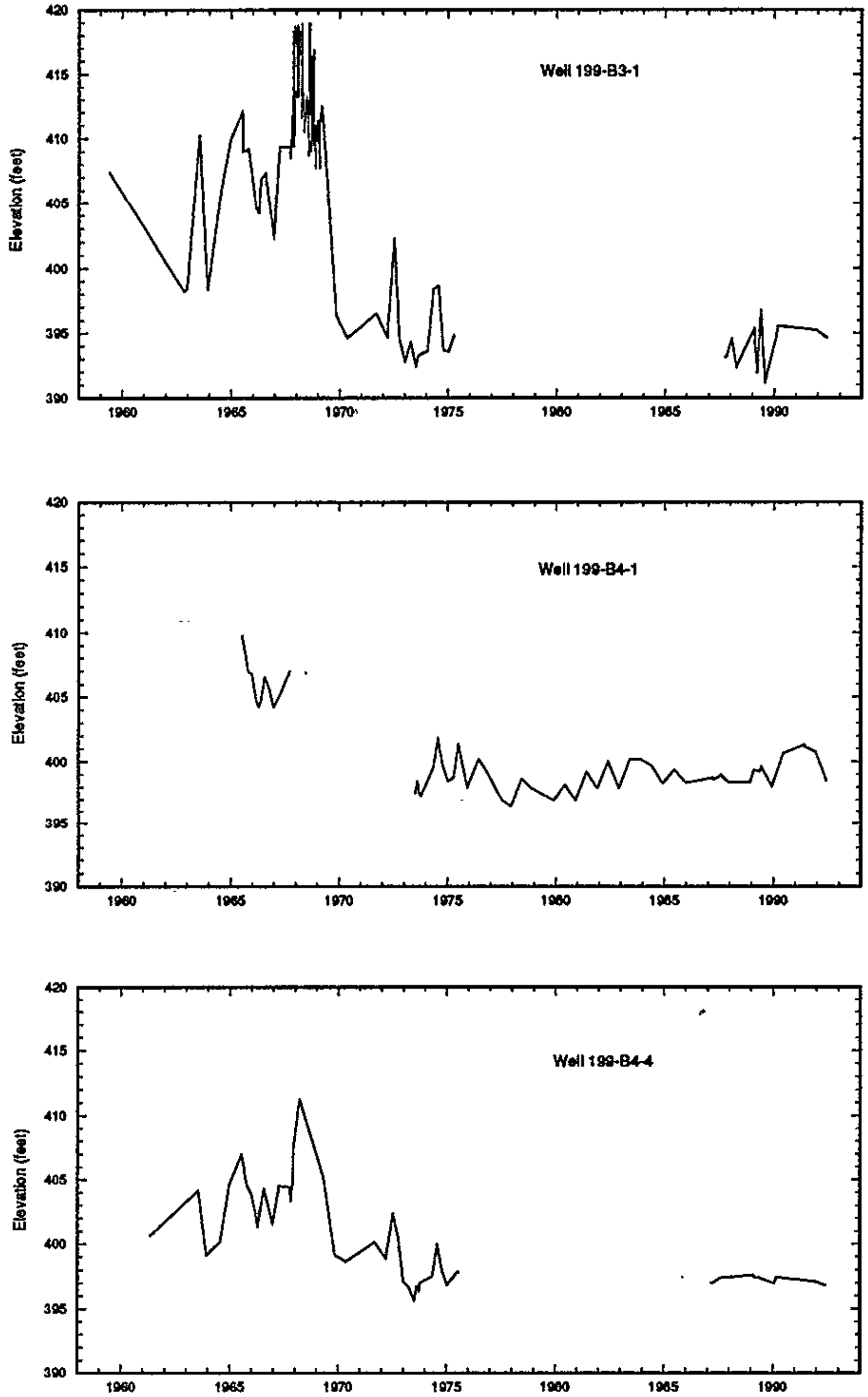
Figure B-2. Manually Collected Water Level Data for Wells Containing Water Level Recorders in the 100-B Area.



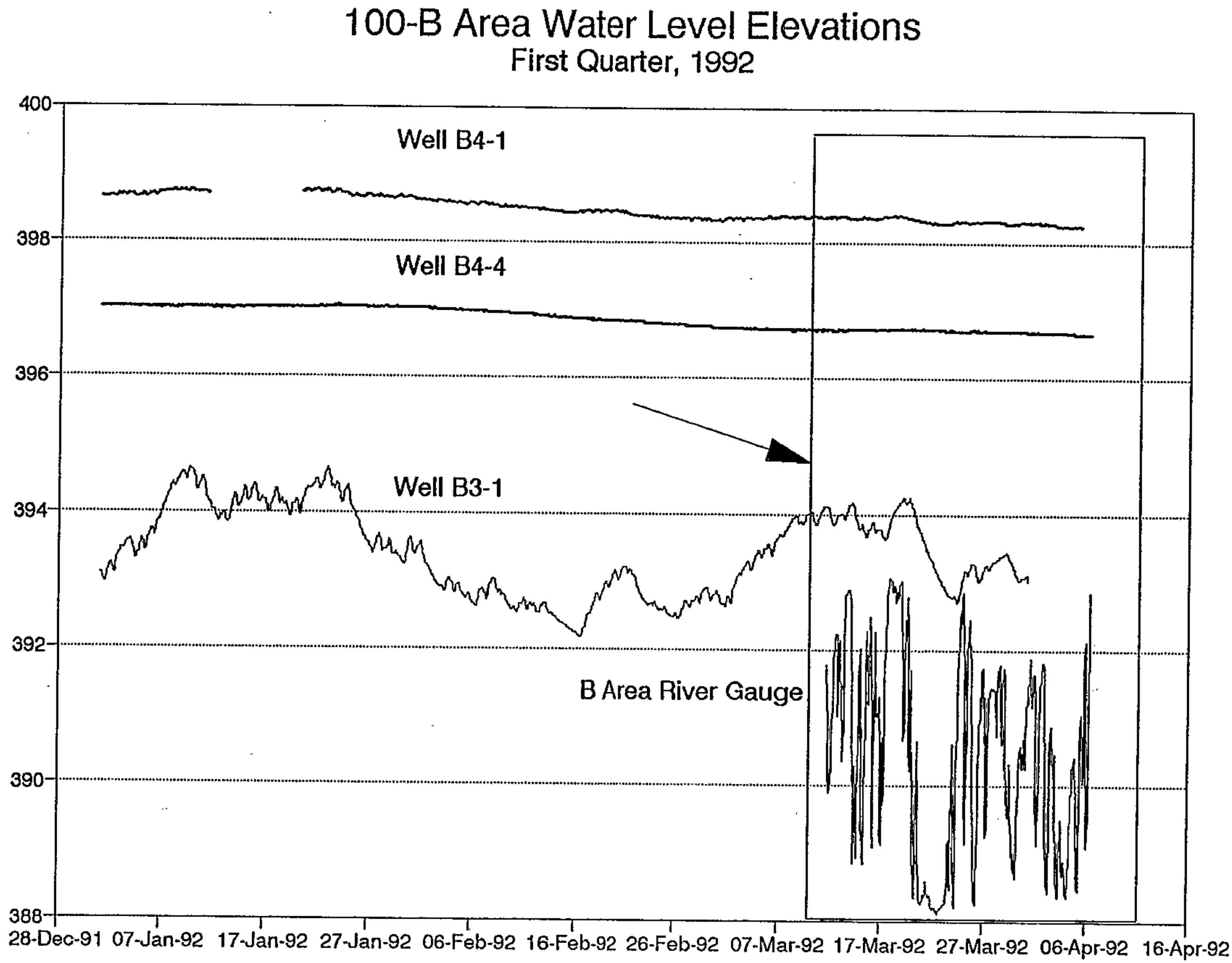


Figure B-3. Data Logger Water Level Records for River and Wells in the 100-B Area for the First Quarter 1992. (sheet 1 of 4)

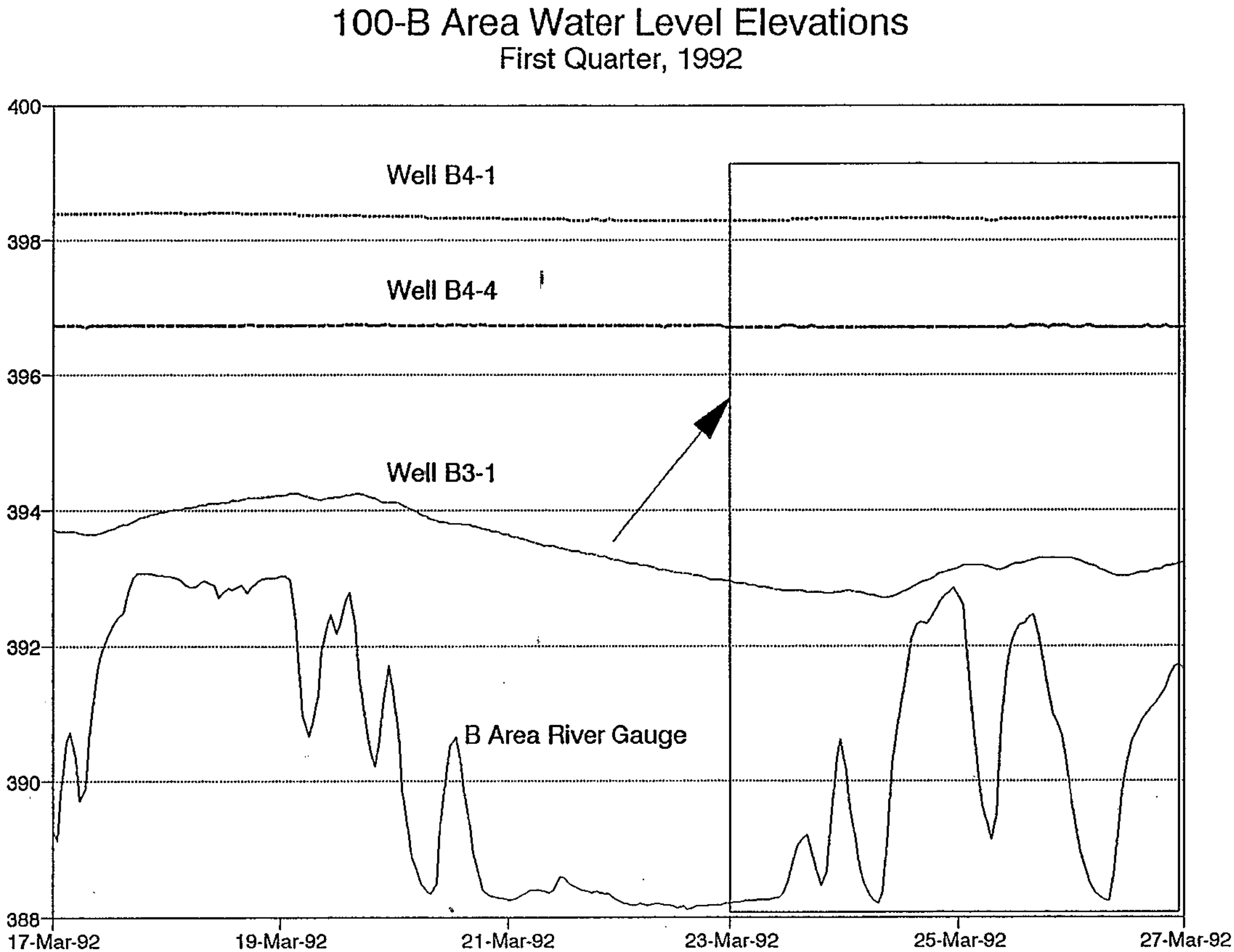


Figure B-3. Data Logger Water Level Records for River and Wells in the 100-B Area for the First Quarter 1992. (sheet 2 of 4)

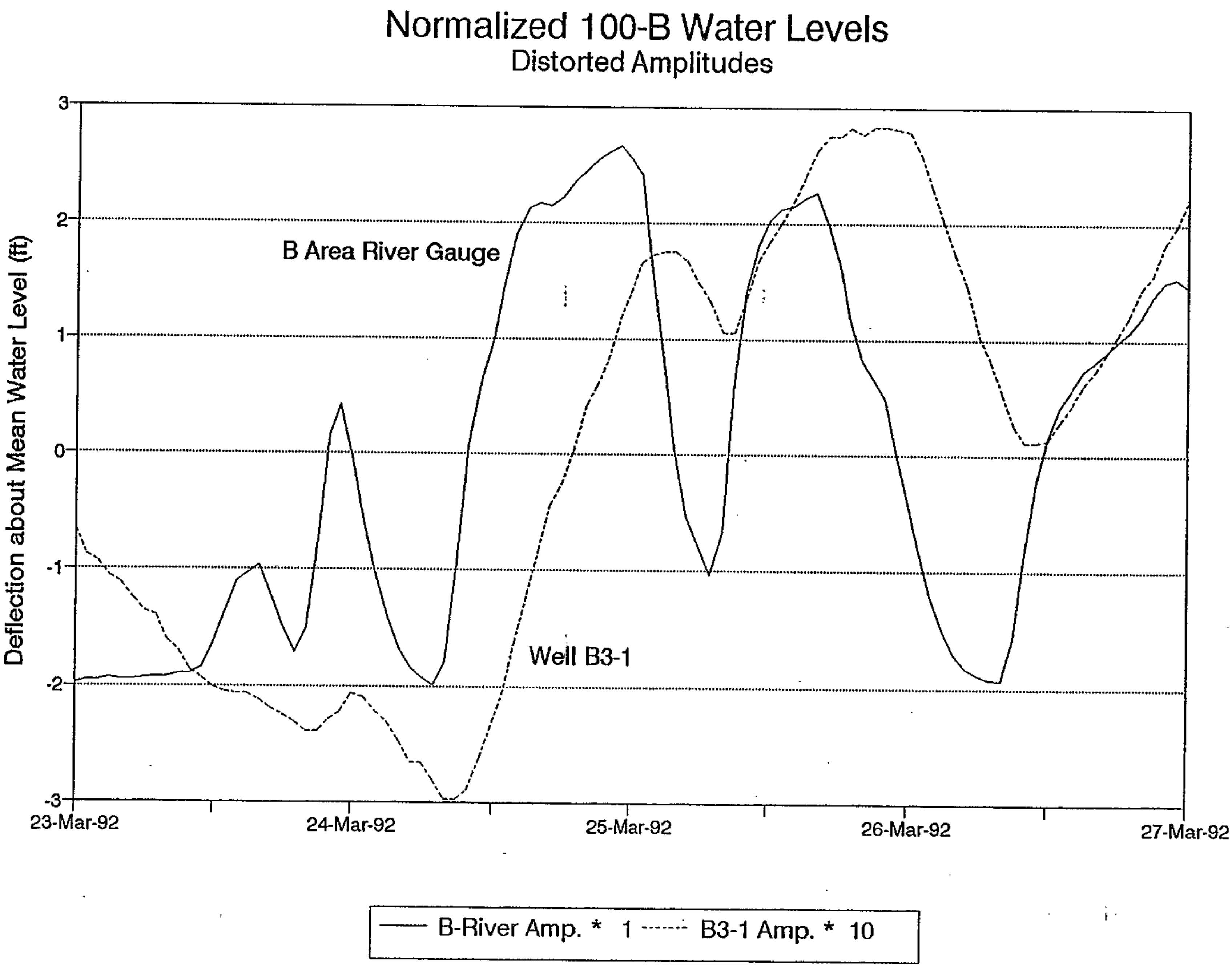
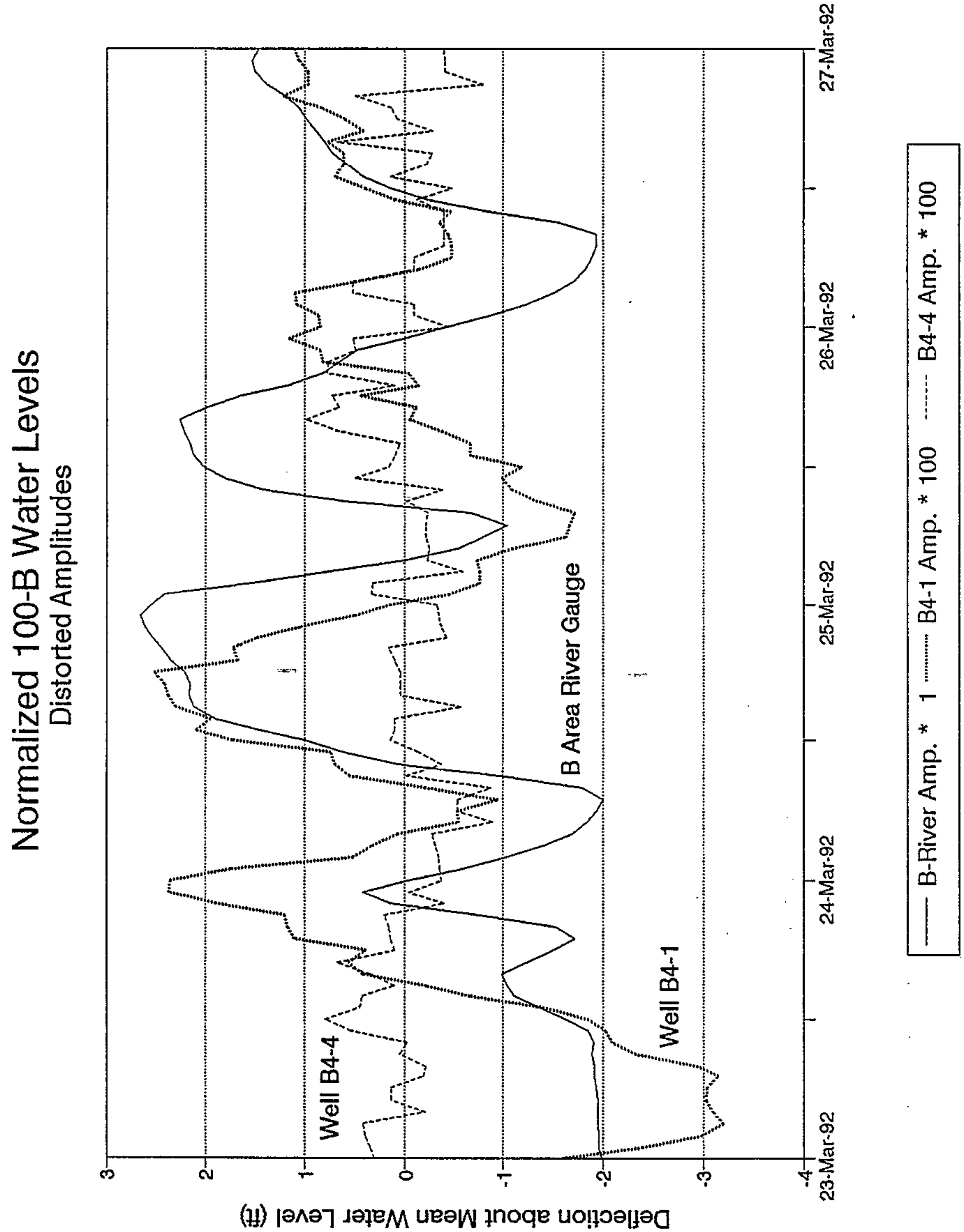


Figure B-3. Data Logger Water Level Records for River and Wells in the 100-B Area for the First Quarter 1992. (sheet 3 of 4)

Figure B-3. Data Logger Water Level Records for River and Wells in the 100-B Area for the First Quarter 1992. (sheet 4 of 4)



9 3 1 2 7 5 2 0 9 1 7

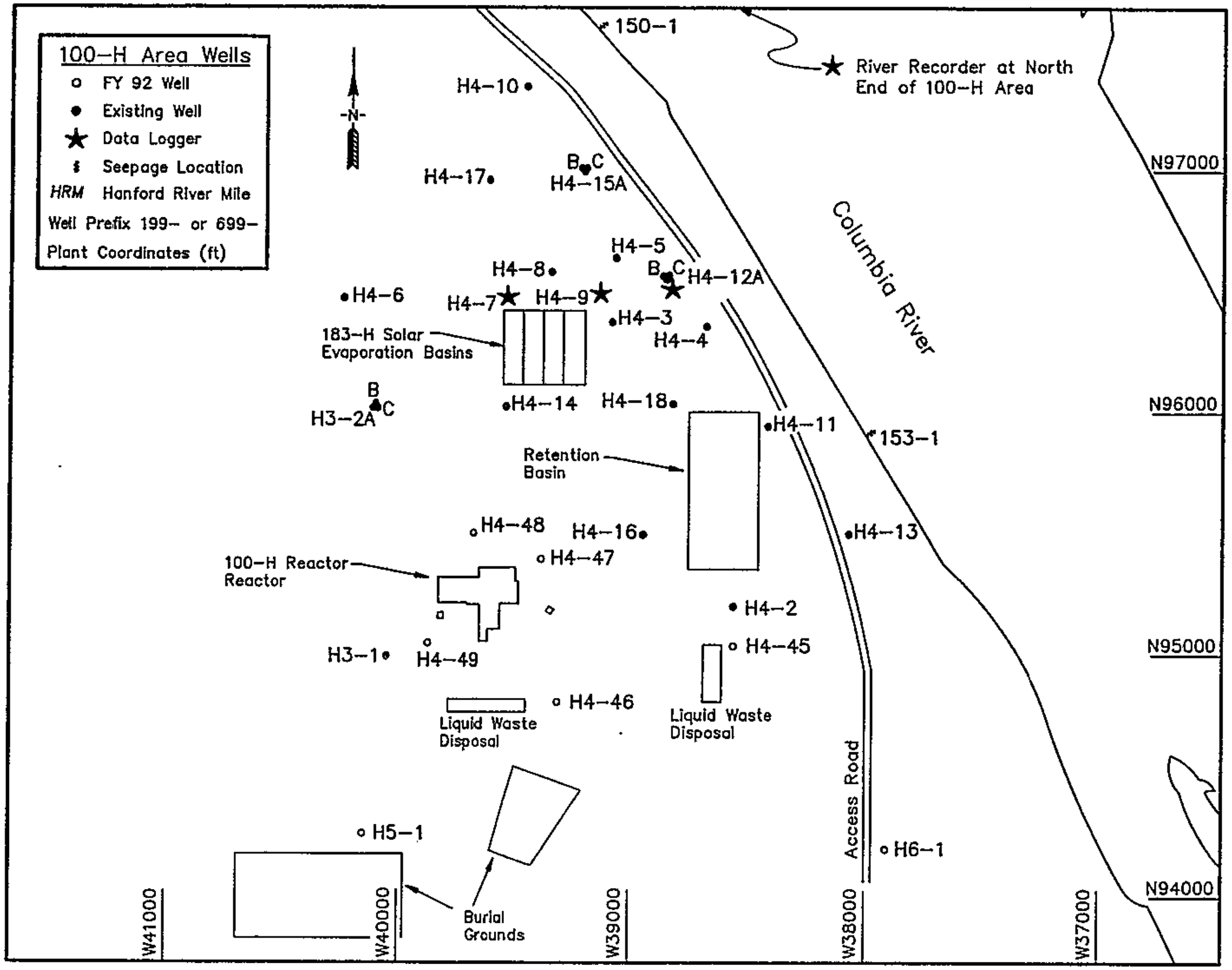


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Figure B-4. Location Map for Water Level Recorders in 100-H Area.

Figure B-5. Manually Collected Water Level Data for Wells Containing Water Level Recorders in the 100-H Area.

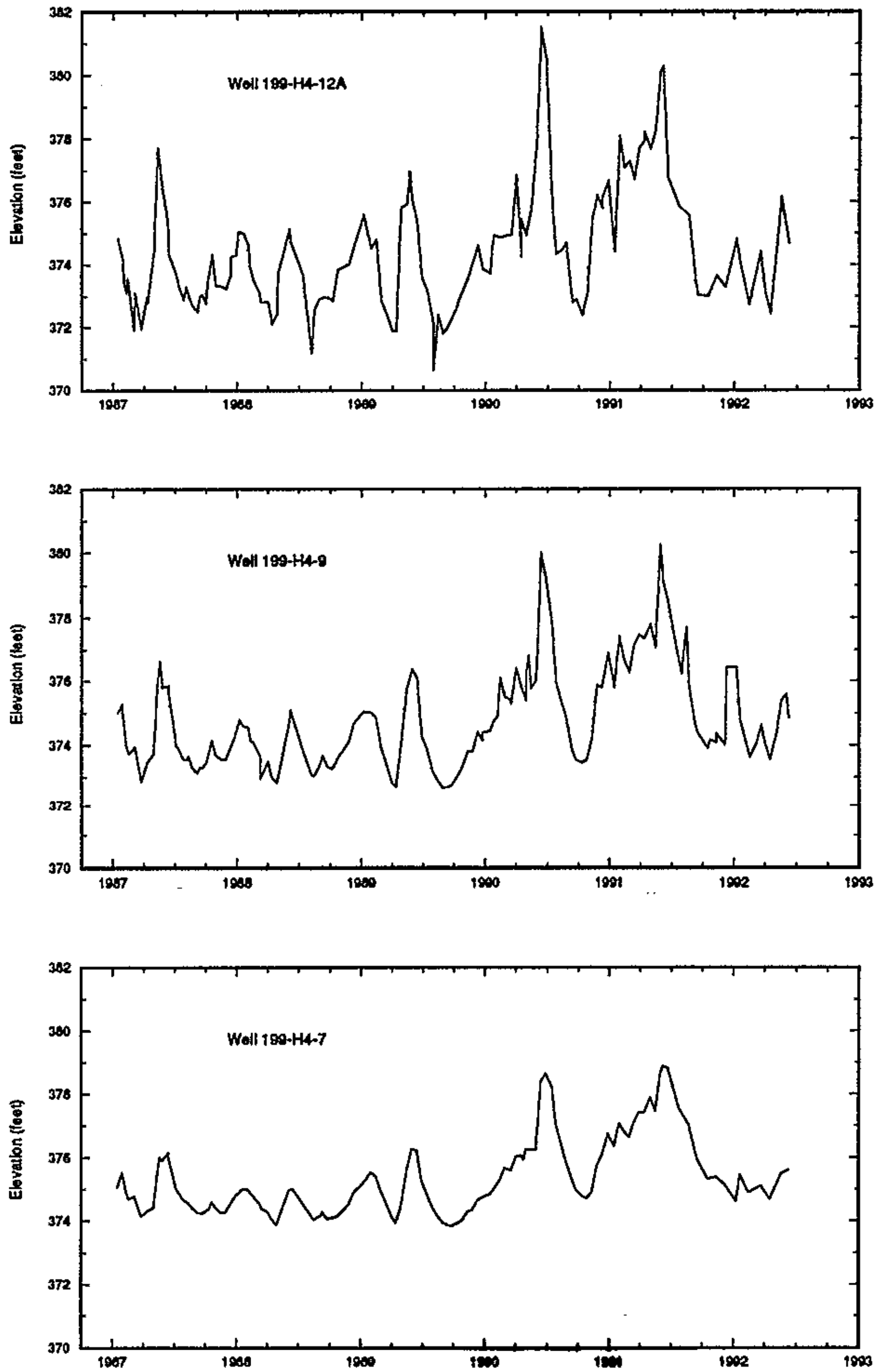

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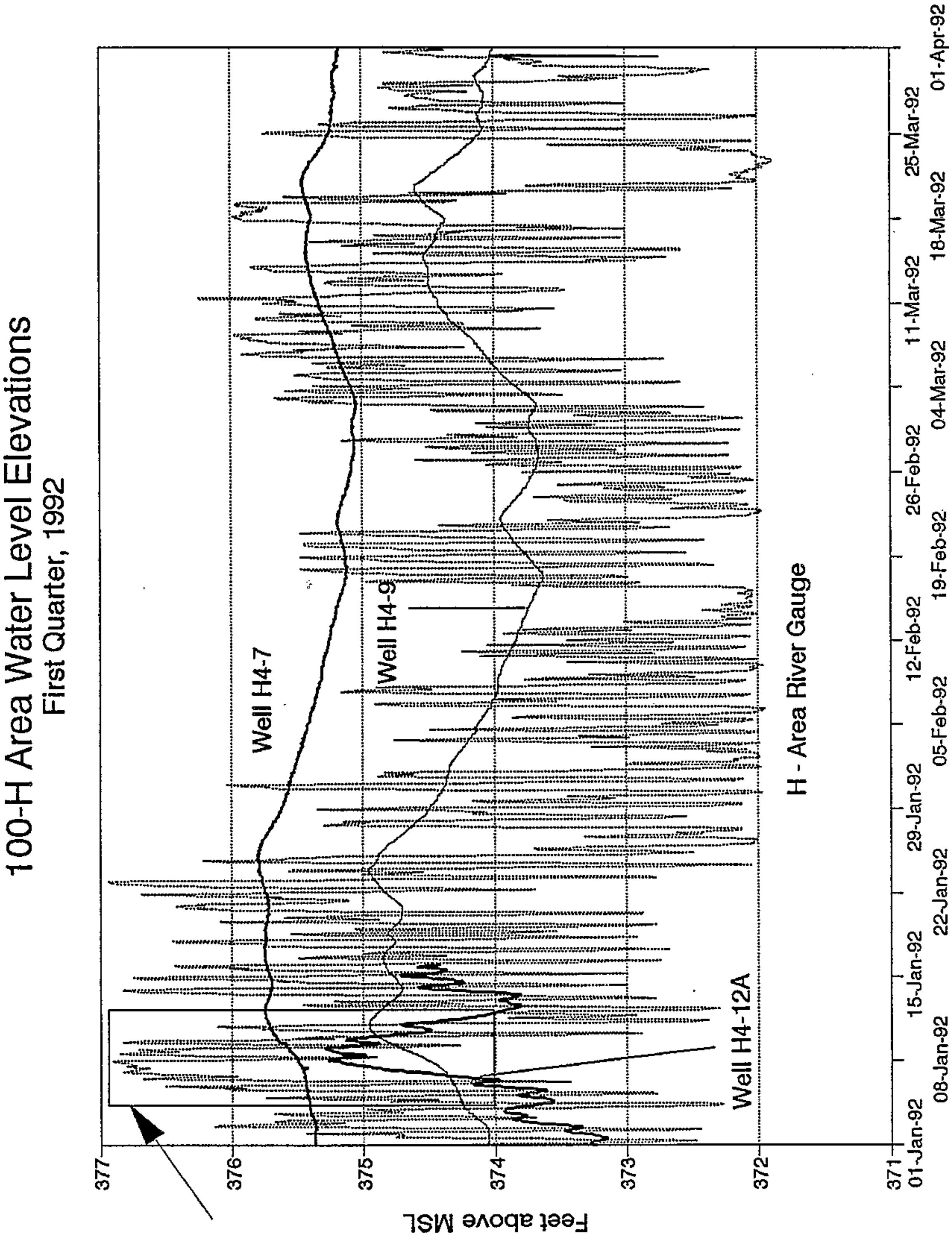
Figure B-6. Data Logger Water Level Records for River and Wells in the 100-H Area for the First Quarter 1992. (sheet 1 of 2)

9 3 1 2 7 5 2 0 9 4 9



9 3 1 2 7 5 2 0 9 5 0

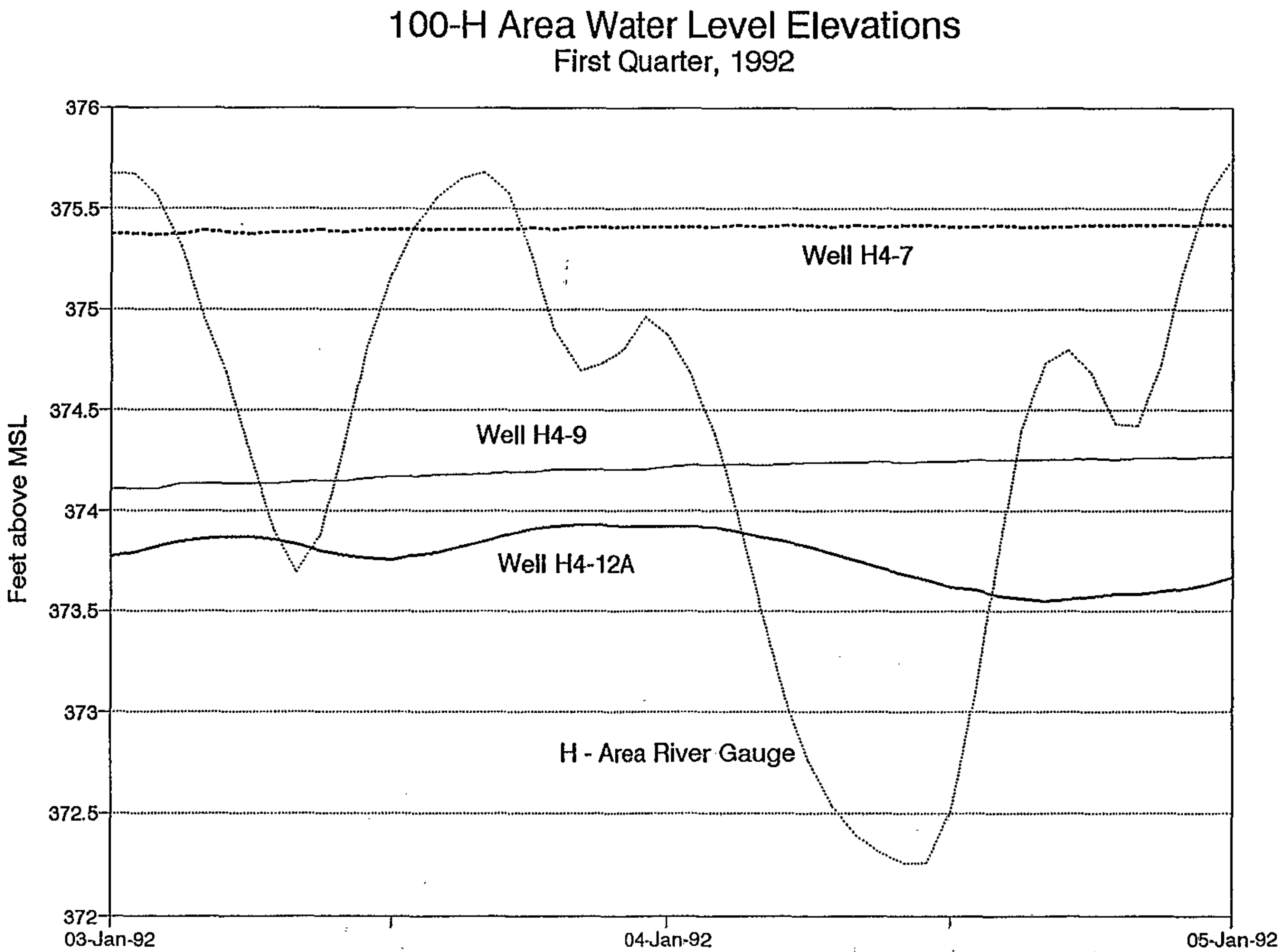


Figure B-6. Data Logger Water Level Records for River and Wells in the 100-H Area for the First Quarter 1992. (sheet 2 of 2)

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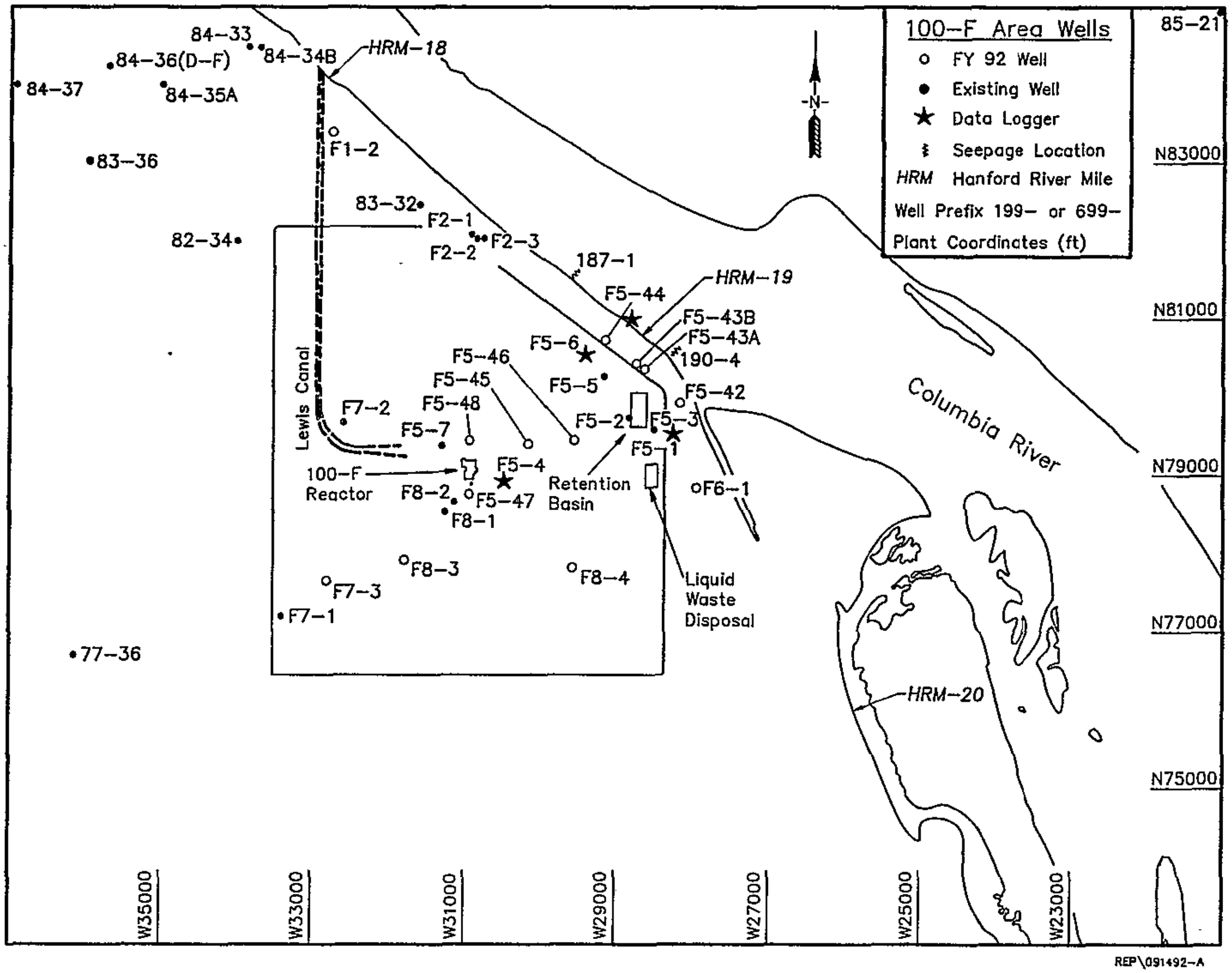
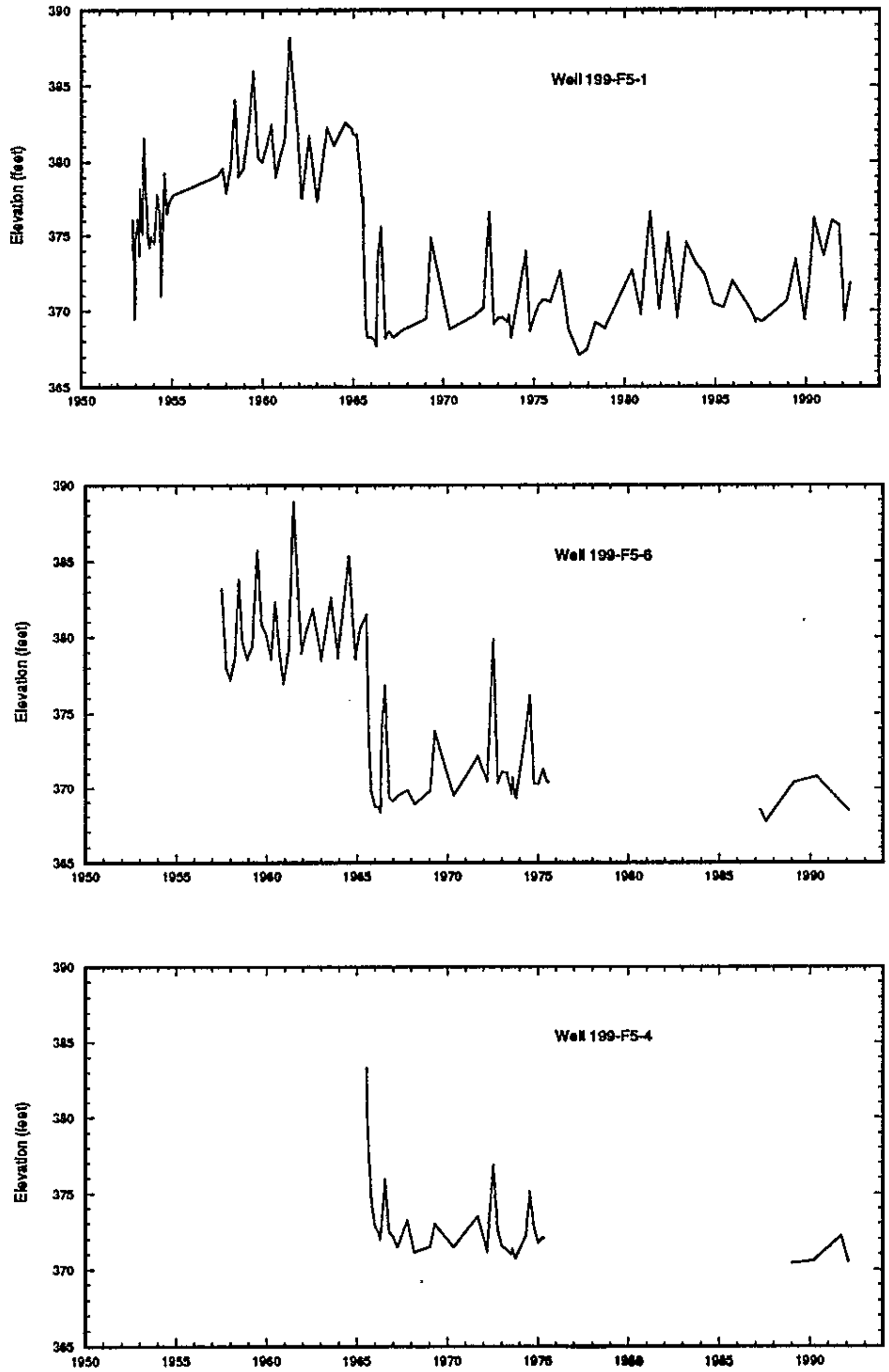


Figure B-7. Location Map for Water Level Recorders in 100-F Area.

Figure B-8. Manually Collected Water Level Data for Wells Containing Water Level Recorders in the 100-F Area.



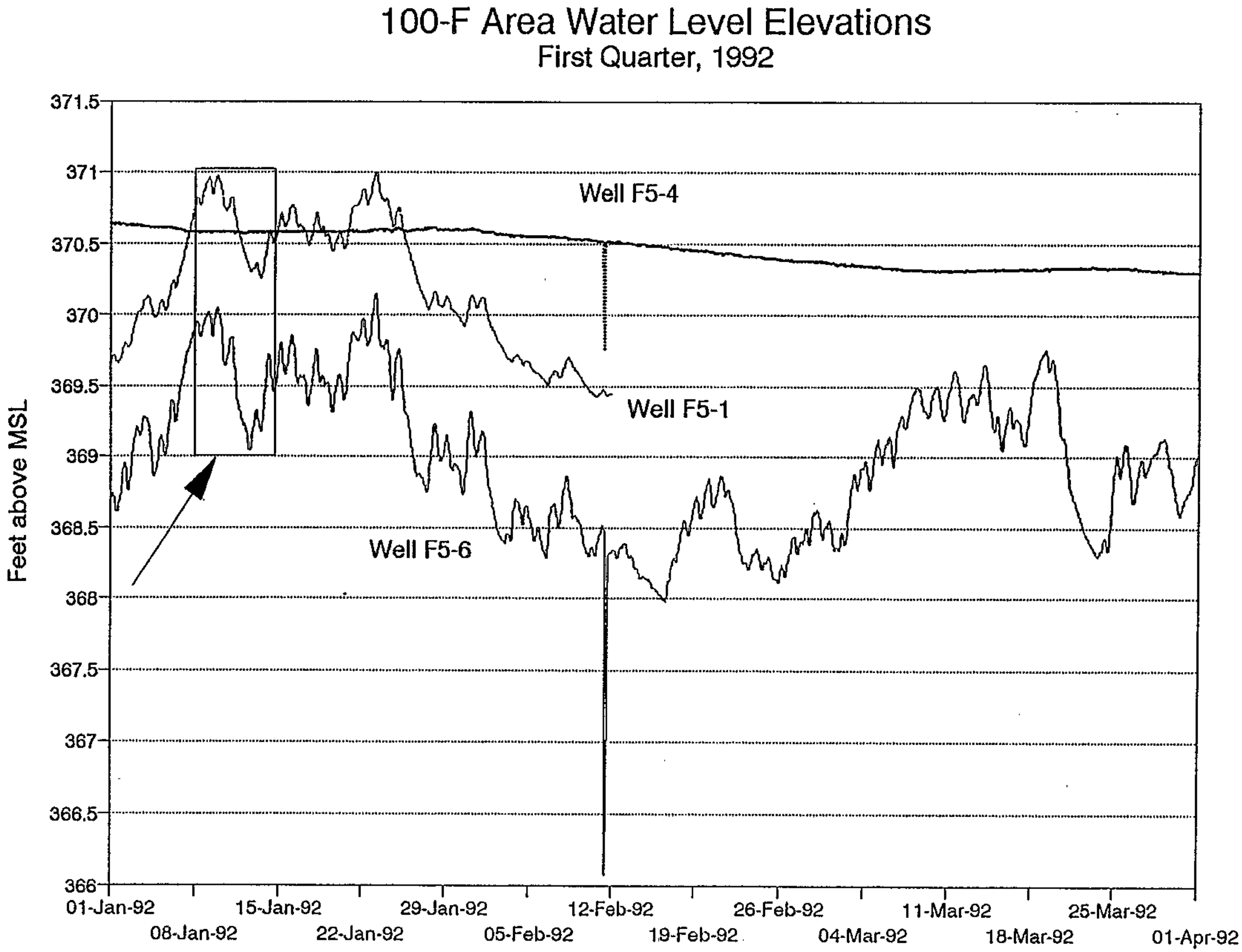


Figure B-9. Data Logger Water Level Records for River and Wells in the 100-F Area for the First Quarter 1992. (sheet 1 of 2)

100-F Area Water Level Elevations
First Quarter, 1992

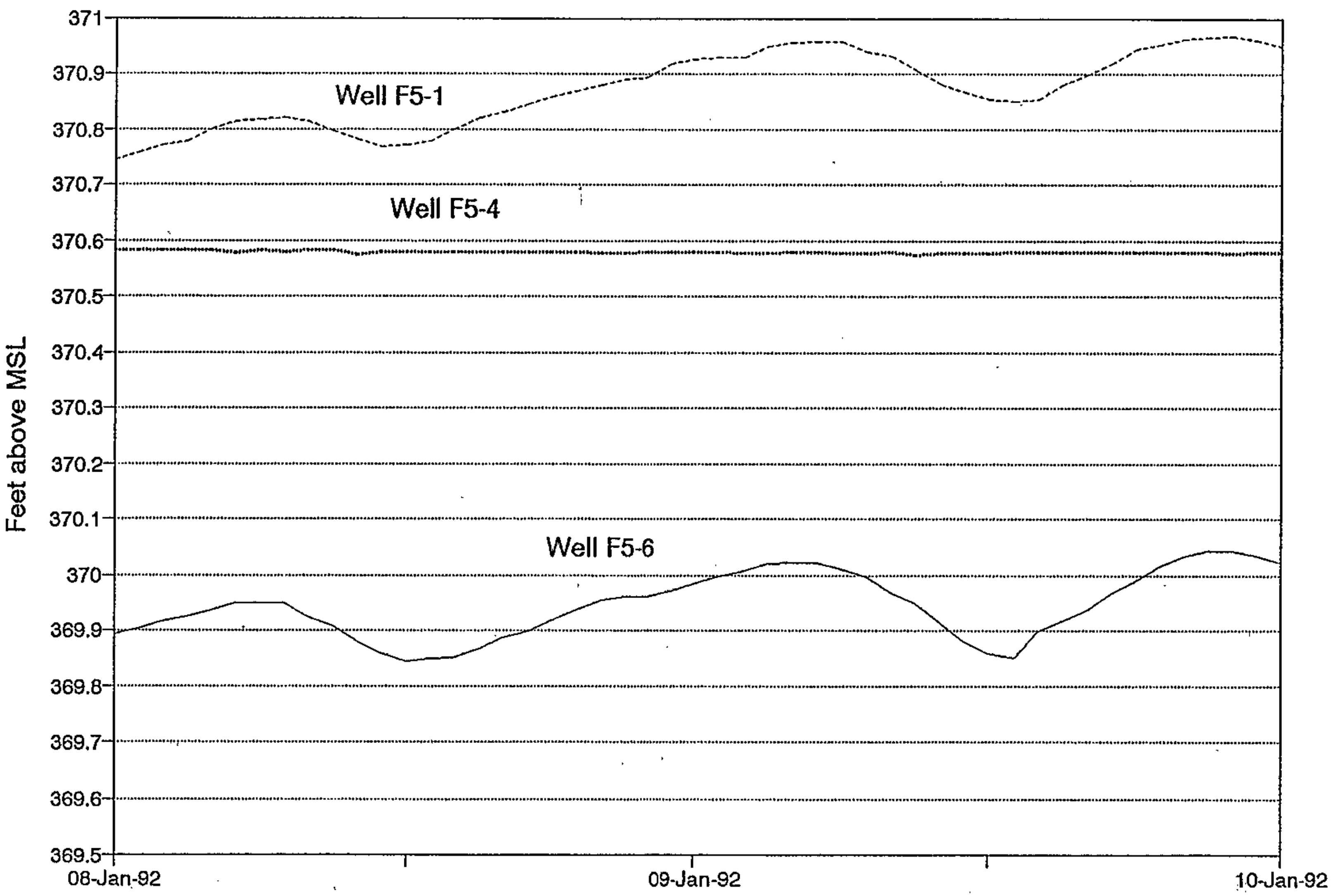


Figure B-9. Data Logger Water Level Records for River and Wells in the 100-F Area for the First Quarter 1992. (sheet 2 of 2)

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APPENDIX C

DATA USED IN FEASIBILITY STUDY OF FERRIS METHOD

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APPENDIX C

DATA USED IN FEASIBILITY STUDY OF FERRIS METHOD

The water level data, stage ratio and lag time measurements, and linear regression plots used in the cyclic fluctuation analysis (Chapter 3) are presented in the following Figures C-1 through C-10 and Tables C-1 through C-6. An index map to the wells used and river stage recorder location is included as Figure C-11.

Figure C-1. Data Logger Records for Wells 399-1-7 and 399-1-2 for (a) May 17-21 and (b) May 25-29, 1992.

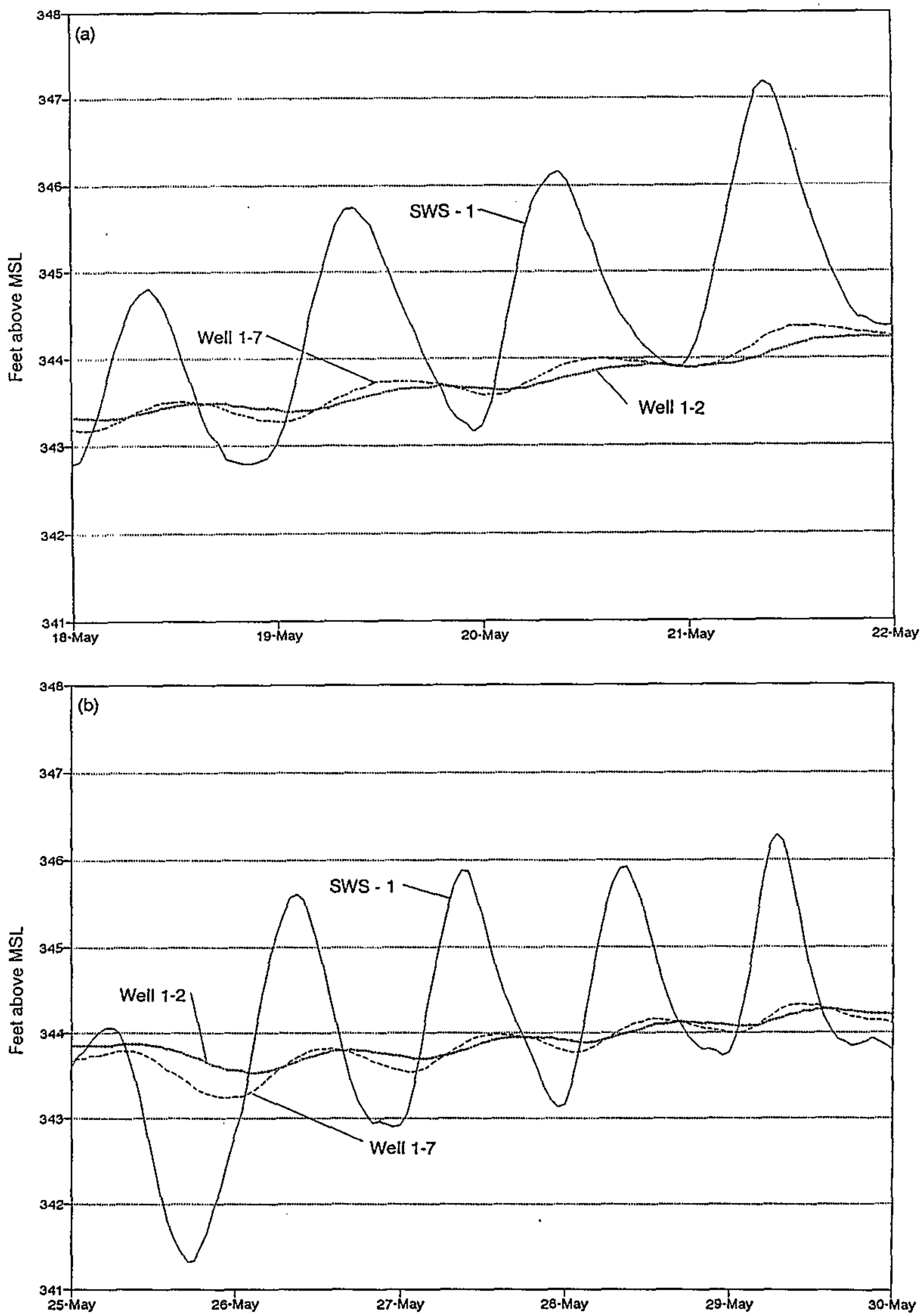


Table C-1. Data for Wells 399-1-7 and 399-1-2 for May 17-21, 1992.

May 17-21

	SWS - 1 Elevation (feet)	Change (feet)	Well 399-1-7 Elevation (feet)	Change (feet)	Change Ratio	Well 399-1-2 Elevation (feet)	Change (feet)	Change Ratio (feet)
Minimum	342.64		343.17			343.3		
Maximum	344.8	2.16	343.51	0.34	0.157407	343.47	0.17	0.078704
Minimum	342.8	-2	343.27	-0.24	0.12	343.39	-0.08	0.04
Maximum	345.74	2.94	343.74	0.47	0.159864	343.68	0.29	0.098639
Minimum	343.16	-2.58	343.57	-0.17	0.065891	343.63	-0.05	0.01938
Maximum	346.15	2.99	343.99	0.42	0.140468	343.92	0.29	0.09699
Minimum	343.88	-2.27	343.88	-0.11	0.048458	343.9	-0.02	0.008811
Maximum	347.18	3.3	344.36	0.48	0.145455	344.24	0.34	0.10303
Avg. Rising Limbs					0.150799			0.094341
Avg. Falling Limbs					0.078117			0.02273
Overall Avg.					0.119649			0.063651

	SWS - 1 Date	Hour	Well 399-1-7 Hour	Lag Time (days)	Well 399-1-2 Hour	Lag Time (days)	Lag Time Between Wells (days)
Minimum	May 17	1800	2600	0.333333	2700	0.375	0.041667
Maximum	May 18	900	1300	0.166667	1600	0.291667	0.125
Minimum	May 18	2000	2400	0.166667	2600	0.25	0.083333
Maximum	May 19	900	1500	0.25	2200	0.541667	0.291667
Minimum	May 19	2300	2500	0.083333	2600	0.125	0.041667
Maximum	May 20	900	1300	0.166667	1900	0.416667	0.25
Minimum	May 20	2300	2400	0.041667	2400	0.041667	0
Maximum	May 21	900	1300	0.166667	2200	0.541667	0.375
Avg. Lag Time (days)				0.171875		0.322917	0.151042

Period = ~ 1 day

Distance (River to 399-1-7) 700 feet
Distance (River to 399-1-2) 1400 feet
Distance (399-1-7 to 399-1-2) 700 feet

Table C-2. Data for Wells 399-1-7 and 399-1-2 for May 25-29, 1992.

May 25-29

	SWS - 1 Elevation (feet)	Change (feet)	Well 399-1-7 Elevation (feet)	Change (feet)	Change Ratio	Well 399-1-2 Elevation (feet)	Change (feet)	Change Ratio (feet)
Minimum	341.31		343.24			343.53		
Maximum	345.61	4.3	343.82	0.58	0.134884	343.8	0.27	0.062791
Minimum	342.9	-2.71	343.54	-0.28	0.103321	343.7	-0.1	0.0369
Maximum	345.87	2.97	343.98	0.44	0.148148	343.95	0.25	0.084175
Minimum	343.13	-2.74	343.77	-0.21	0.076642	343.88	-0.07	0.025547
Maximum	345.92	2.79	344.15	0.38	0.136201	344.12	0.24	0.086022
Minimum	343.73	-2.19	343.99	-0.16	0.073059	344.07	-0.05	0.022831
Maximum	346.28	2.55	344.33	0.34	0.133333	344.27	0.2	0.078431
Avg. Rising Limbs					0.435			0.077855
Avg. Falling Limbs					-0.21667			0.028426
Overall Avg.					0.155714			0.056671

	SWS - 1 Date	Hour	Well 399-1-7 Hour	Lag Time (days)	Well 399-1-2 Hour	Lag Time (days)	Lag Time Between Wells (days)
Minimum	May 25	1700	2400	0.291667	2600	0.375	0.083333
Maximum	May 26	900	1400	0.208333	1700	0.333333	0.125
Minimum	May 26	2300	2600	0.125	2700	0.166667	0.041667
Maximum	May 27	900	1400	0.208333	1800	0.375	0.166667
Minimum	May 27	2300	2600	0.125	2700	0.166667	0.041667
Maximum	May 28	900	1400	0.208333	1700	0.333333	0.125
Minimum	May 28	2300	2500	0.083333	2600	0.125	0.041667
Maximum	May 29	700	1100	0.166667	1400	0.291667	0.125
				0.177083		0.270833	0.09375

Period =~ 1 day

Figure C-2. Data Logger Records for Wells 399-3-9 and 399-3-12 for (a) May 17-21 and (b) May 25-29, 1992.

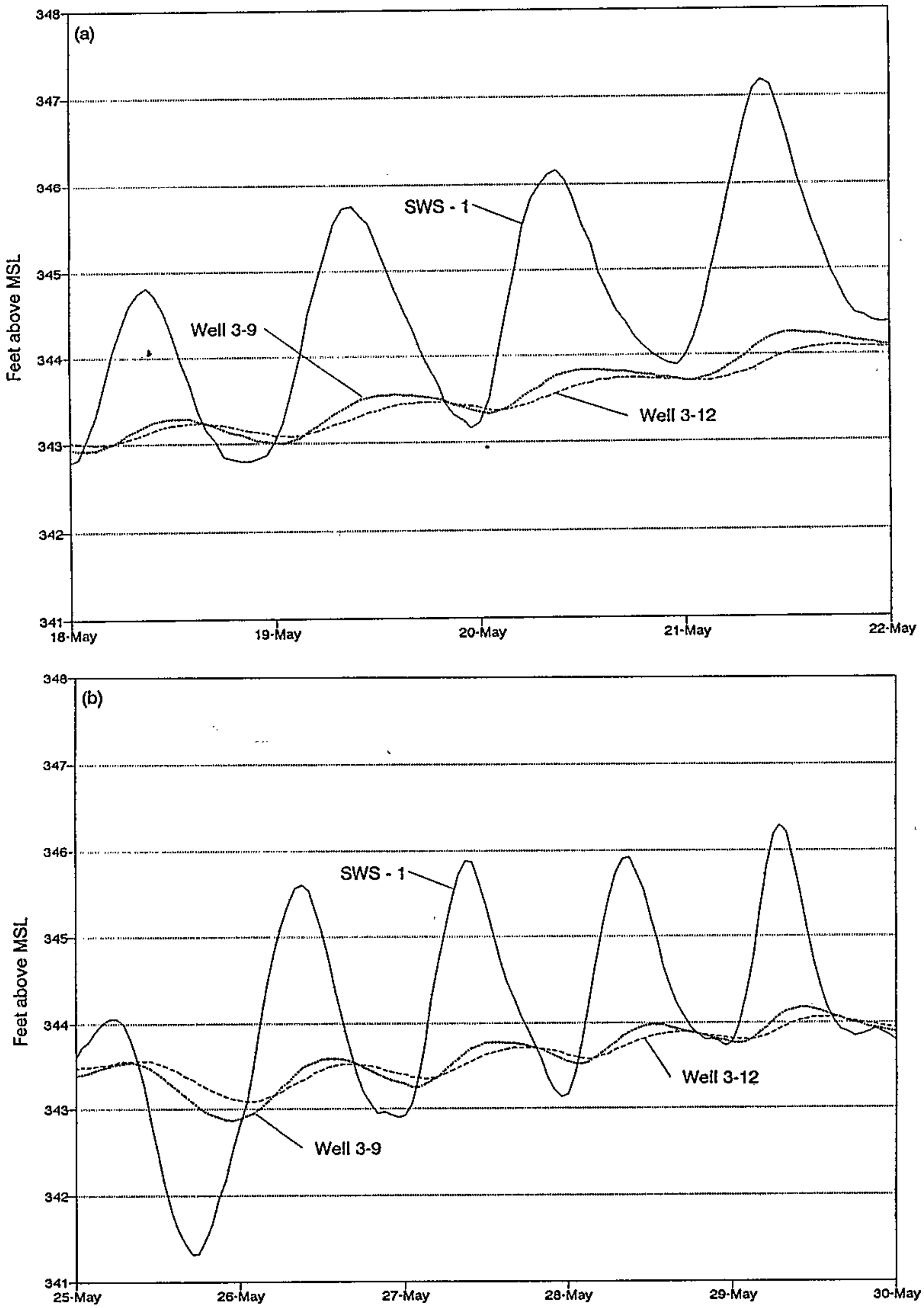


Table C-3. Data for Wells 399-3-9 and 399-3-12 for May 17-21, 1992.

May 17-21

	SWS - 1 Elevation (feet)	Change (feet)	Well 399-3-9 Elevation (feet)	Change (feet)	Change Ratio	Well 399-3-12 Elevation (feet)	Change (feet)	Change Ratio (feet)
Minimum	342.64		342.92			342.99		
Maximum	344.8	2.16	343.29	0.37	0.171296	343.23	0.24	0.111111
Minimum	342.8	-2	343	-0.29	0.145	343.08	-0.15	0.075
Maximum	345.74	2.94	343.54	0.54	0.183673	343.46	0.38	0.129252
Minimum	343.16	-2.58	343.33	-0.21	0.081395	343.37	-0.09	0.034884
Maximum	346.15	2.99	343.83	0.5	0.167224	343.73	0.36	0.120401
Minimum	343.88	-2.27	343.7	-0.13	0.057269	343.69	-0.04	0.017621
Maximum	347.18	3.3	344.25	0.55	0.166667	344.1	0.41	0.124242
Avg. Rising Limbs					0.172215			0.121252
Avg. Falling Limbs					0.094555			0.042502
Overall Avg.					0.138932			0.087502

	SWS - 1 Date	Hour	Well 399-3-9 Hour	Lag Time (days)	Well 399-3-12 Hour	Lag Time (days)	Lag Time Between Wells (days)
Minimum	May 17	1800	2500	0.291667	2700	0.375	0.083333
Maximum	May 18	900	1200	0.125	1400	0.208333	0.083333
Minimum	May 18	2000	2500	0.208333	2700	0.291667	0.083333
Maximum	May 19	900	1400	0.208333	1900	0.416667	0.208333
Minimum	May 19	2300	2500	0.083333	2700	0.166667	0.083333
Maximum	May 20	900	1400	0.208333	1800	0.375	0.166667
Minimum	May 20	2300	2400	0.041667	2500	0.083333	0.041667
Maximum	May 21	900	1300	0.166667	1900	0.416667	0.25
Average Lag Time (days)				0.166667		0.291667	0.125

Period = 1 day

Distance (River to 399-3-9) 200 feet
Distance (River to 399-3-12) 1200 feet
Distance (399-1-9 to 399-3-12) 1000 feet

Table C-4. Data for Wells 399-3-9 and 399-1-12 for May 25-29, 1992.

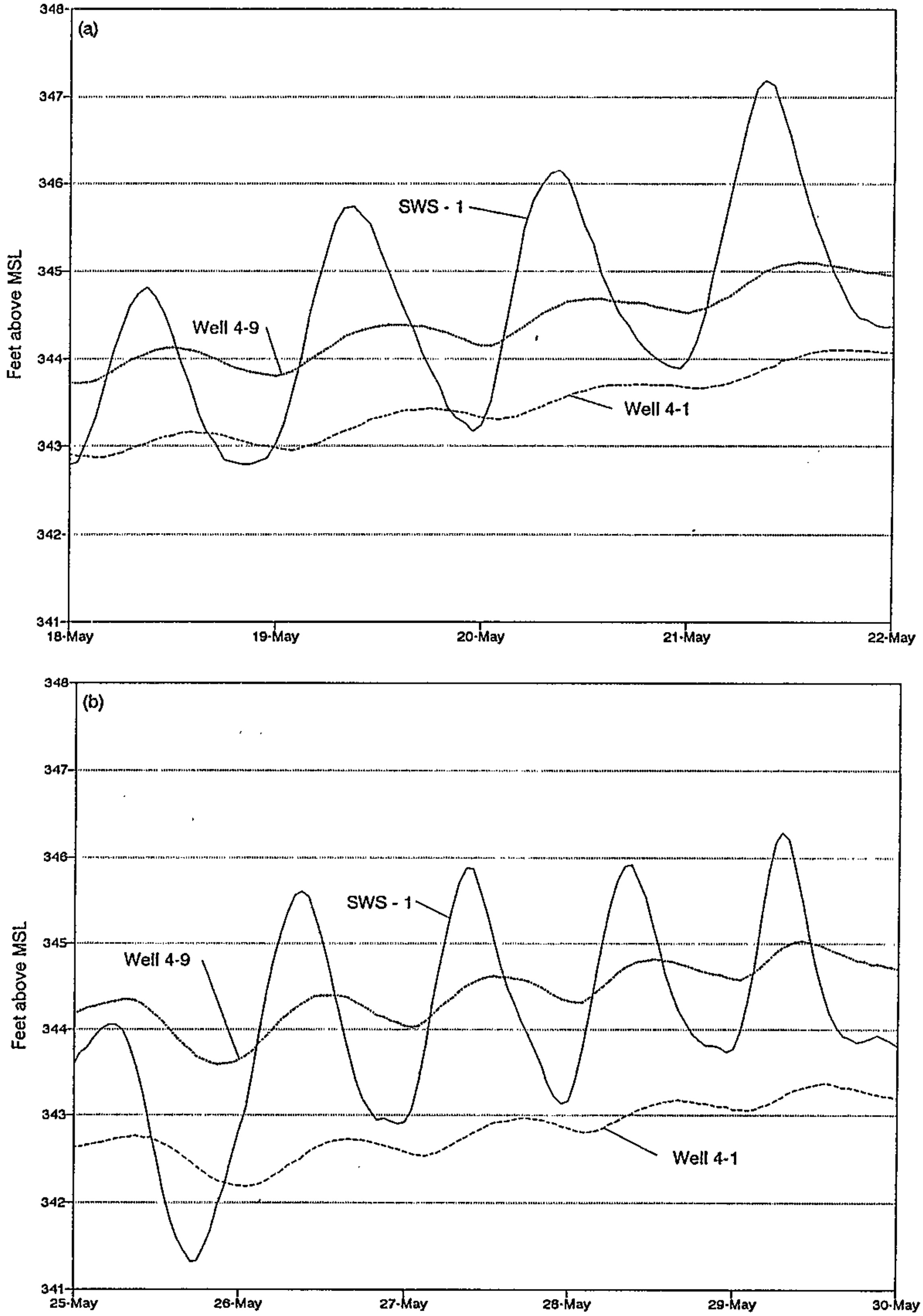
May 25-29

	SWS - 1 Elevation (feet)	Change (feet)	Well 399-3-9 Elevation (feet)	Change (feet)	Change Ratio	Well 399-3-12 Elevation (feet)	Change (feet)	Change Ratio (feet)
Minimum	341.31		342.87			343.08		
Maximum	345.61	4.3	343.59	0.72	0.167442	343.52	0.44	0.102326
Minimum	342.9	-2.71	343.25	-0.34	0.125461	343.34	-0.18	0.066421
Maximum	345.87	2.97	343.78	0.53	0.178451	343.71	0.37	0.124579
Minimum	343.13	-2.74	343.52	-0.26	0.094891	343.58	-0.13	0.047445
Maximum	345.92	2.79	343.97	0.45	0.16129	343.89	0.31	0.111111
Minimum	343.73	-2.19	343.76	-0.21	0.09589	343.8	-0.09	0.041096
Maximum	346.28	2.55	344.17	0.41	0.160784	344.07	0.27	0.105882
Avg. Rising Limbs					0.166992			0.110975
Avg. Falling Limbs					0.105414			0.051654
Overall Avg.					0.140601			0.085551

	SWS - 1 Date	Hour	Well 399-3-9 Hour	Lag Time (days)	Well 399-3-12 Hour	Lag Time (days)	Lag Time Between Wells (days)
Minimum	May 25	1700	2300	0.25	2500	0.333333	0.083333
Maximum	May 26	900	1300	0.166667	1600	0.291667	0.125
Minimum	May 26	2300	2500	0.083333	2700	0.166667	0.083333
Maximum	May 27	900	1400	0.208333	1800	0.375	0.166667
Minimum	May 27	2300	2600	0.125	2700	0.166667	0.041667
Maximum	May 28	900	1200	0.125	1700	0.333333	0.208333
Minimum	May 28	2300	2500	0.083333	2600	0.125	0.041667
Maximum	May 29	700	1000	0.125	1300	0.25	0.125
Average Lag Time (days)				0.145833		0.255208	0.109375

Period = 1 day

Figure C-3. Data Logger Records for Wells 399-4-9 and 399-4-1 for (a) May 17-21 and (b) May 25-29, 1992.



9 3 1 2 7 5 2 0 9 5 5

Table C-5. Data for Wells 399-4-9 and 399-4-1 for May 17-21, 1992.

May 17-21

	SWS - 1 Elevation (feet)	Change (feet)	Well 399-4-9 Elevation (feet)	Change (feet)	Change Ratio	Well 399-4-1 Elevation (feet)	Change (feet)	Change Ratio (feet)
Minimum	342.64		343.71			342.87		
Maximum	344.8	2.16	344.12	0.41	0.189815	343.15	0.28	0.12963
Minimum	342.8	-2	343.8	-0.32	0.16	342.95	-0.2	0.1
Maximum	345.74	2.94	344.39	0.59	0.20068	343.41	0.46	0.156463
Minimum	343.16	-2.58	344.15	-0.24	0.093023	343.3	-0.11	0.042636
Maximum	346.15	2.99	344.69	0.54	0.180602	343.69	0.39	0.130435
Minimum	343.88	-2.27	344.53	-0.16	0.070485	343.65	-0.04	0.017621
Maximum	347.18	3.3	345.1	0.57	0.172727	344.1	0.45	0.136364
Avg. Rising Limbs					0.185956			0.138223
Avg. Falling Limbs					0.107836			0.053419
Overall Avg.					0.152476			0.101878

	SWS - 1 Date	Hour	Well 399-4-9 Hour	Lag Time (days)	Well 399-4-1 Hour	Lag Time (days)	Lag Time Between Wells (days)
Minimum	May 17	1800	2500	0.291667	2700	0.375	0.083333
Maximum	May 18	900	1200	0.125	1400	0.208333	0.083333
Minimum	May 18	2000	2400	0.166667	2600	0.25	0.083333
Maximum	May 19	900	1300	0.166667	1800	0.375	0.208333
Minimum	May 19	2300	2400	0.041667	2600	0.125	0.083333
Maximum	May 20	900	1300	0.166667	1900	0.416667	0.25
Minimum	May 20	2300	2400	0.041667	2600	0.125	0.083333
Maximum	May 21	900	1300	0.166667	1700	0.333333	0.166667
Average Lag Time (days)				0.145833		0.276042	0.130208

Period = 1 day

Distance (River to 399-4-9) 300 feet
Distance (River to 399-4-1) 1400 feet
Distance (399-4-9 to 399-4-1) 1100 feet

Table C-6. Data for Wells 399-4-9 and 399-4-1 for May 25-29, 1992.

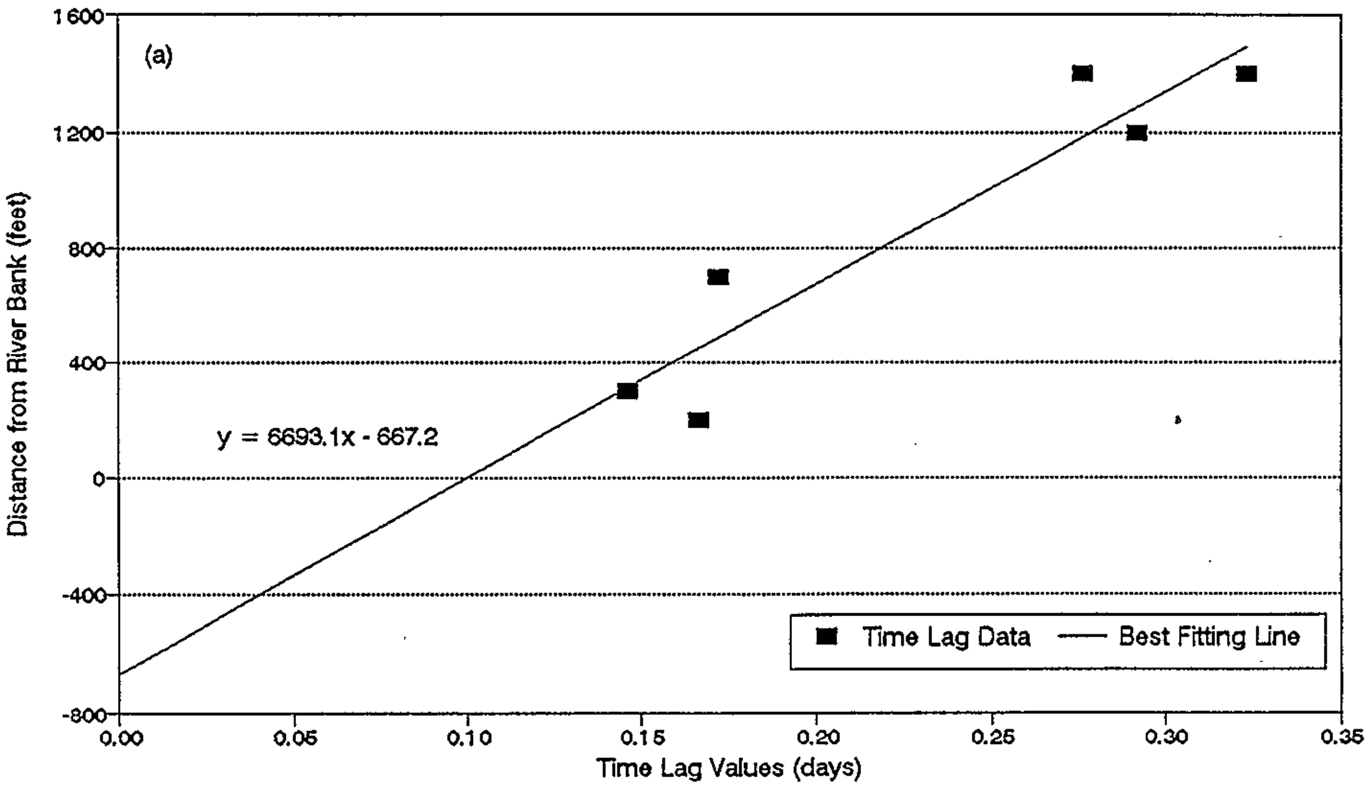
May 25-29

	SWS - 1 Elevation (feet)	Change (feet)	Well 399-4-9 Elevation (feet)	Change (feet)	Change Ratio	Well 399-4-1 Elevation (feet)	Change (feet)	Change Ratio (feet)
Minimum	341.31		343.59			342.87		
Maximum	345.61	4.3	344.4	0.81	0.188372	343.41	0.54	0.125581
Minimum	342.9	-2.71	344.03	-0.37	0.136531	343.21	-0.2	0.073801
Maximum	345.87	2.97	344.62	0.59	0.198653	343.64	0.43	0.144781
Minimum	343.13	-2.74	344.31	-0.31	0.113139	343.49	-0.15	0.054745
Maximum	345.92	2.79	344.81	0.5	0.179211	343.85	0.36	0.129032
Minimum	343.73	-2.19	344.58	-0.23	0.105023	343.73	-0.12	0.054795
Maximum	346.28	2.55	345.02	0.44	0.172549	344.04	0.31	0.121569
Avg. Rising Limbs					0.184696			0.130241
Avg. Falling Limbs					0.118231			0.061113
Overall Avg.					0.156211			0.100615

	SWS - 1 Date	Hour	Well 399-4-9 Hour	Lag Time (days)	Well 399-4-1 Hour	Lag Time (days)	Lag Time Between Wells (days)
Minimum	May 25	1700	2100	0.166667	2500	0.333333	0.166667
Maximum	May 26	900	1300	0.166667	1600	0.291667	0.125
Minimum	May 26	2300	2500	0.083333	2700	0.166667	0.083333
Maximum	May 27	900	1300	0.166667	1800	0.375	0.208333
Minimum	May 27	2300	2500	0.083333	2600	0.125	0.041667
Maximum	May 28	900	1200	0.125	1600	0.291667	0.166667
Minimum	May 28	2300	2500	0.083333	2600	0.125	0.041667
Maximum	May 29	700	1000	0.125	1300	0.25	0.125
Average Lag Time (days)				0.125		0.244792	0.119792

Period = 1 day

Figure C-4. Time Lag Method Regression Plots Using SWS-1 as the Source for (a) May 17-21 and (b) May 25-29, 1992.



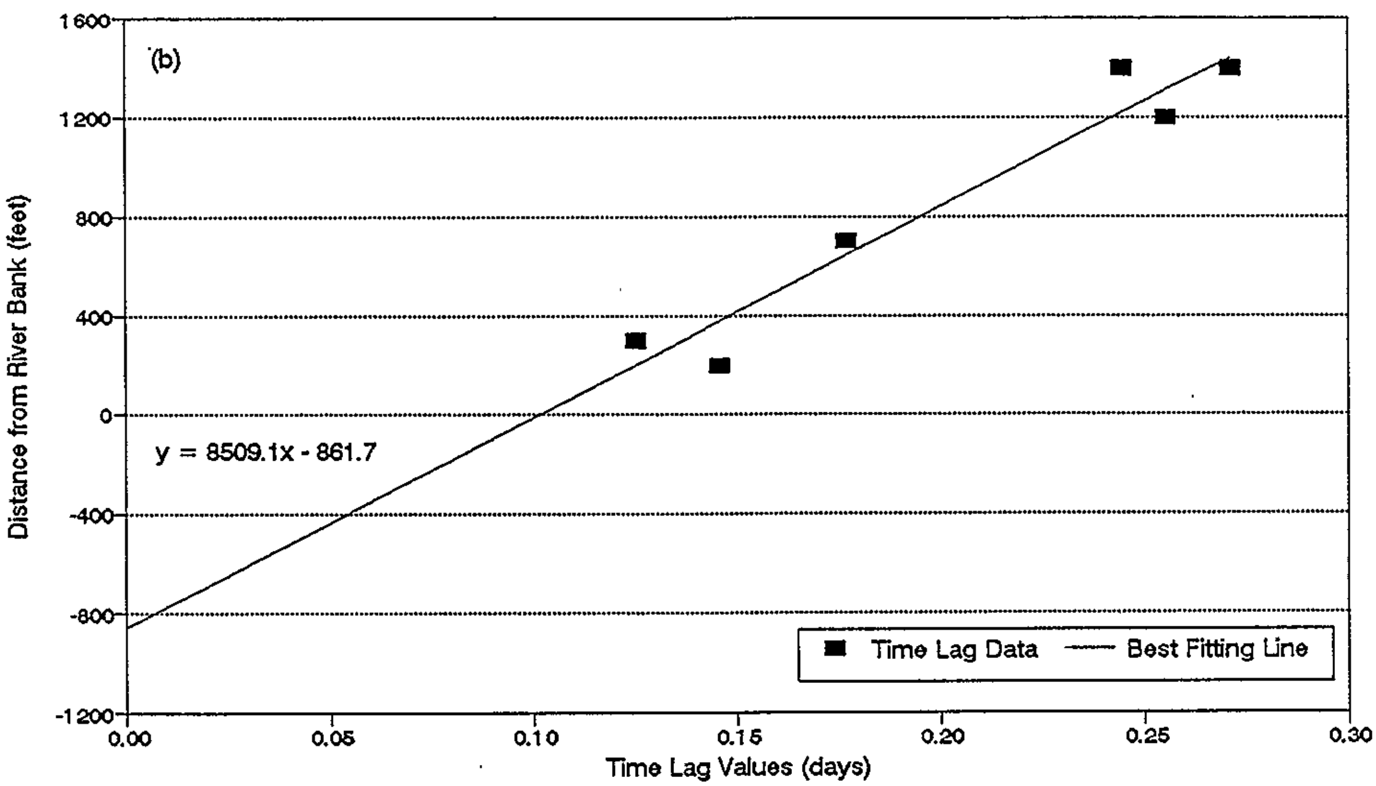


Figure C-5. Time Lag Method Regression Plots Using 399-4-9 as the Source for (a) May 17-21 and (b) May 25-29, 1992.

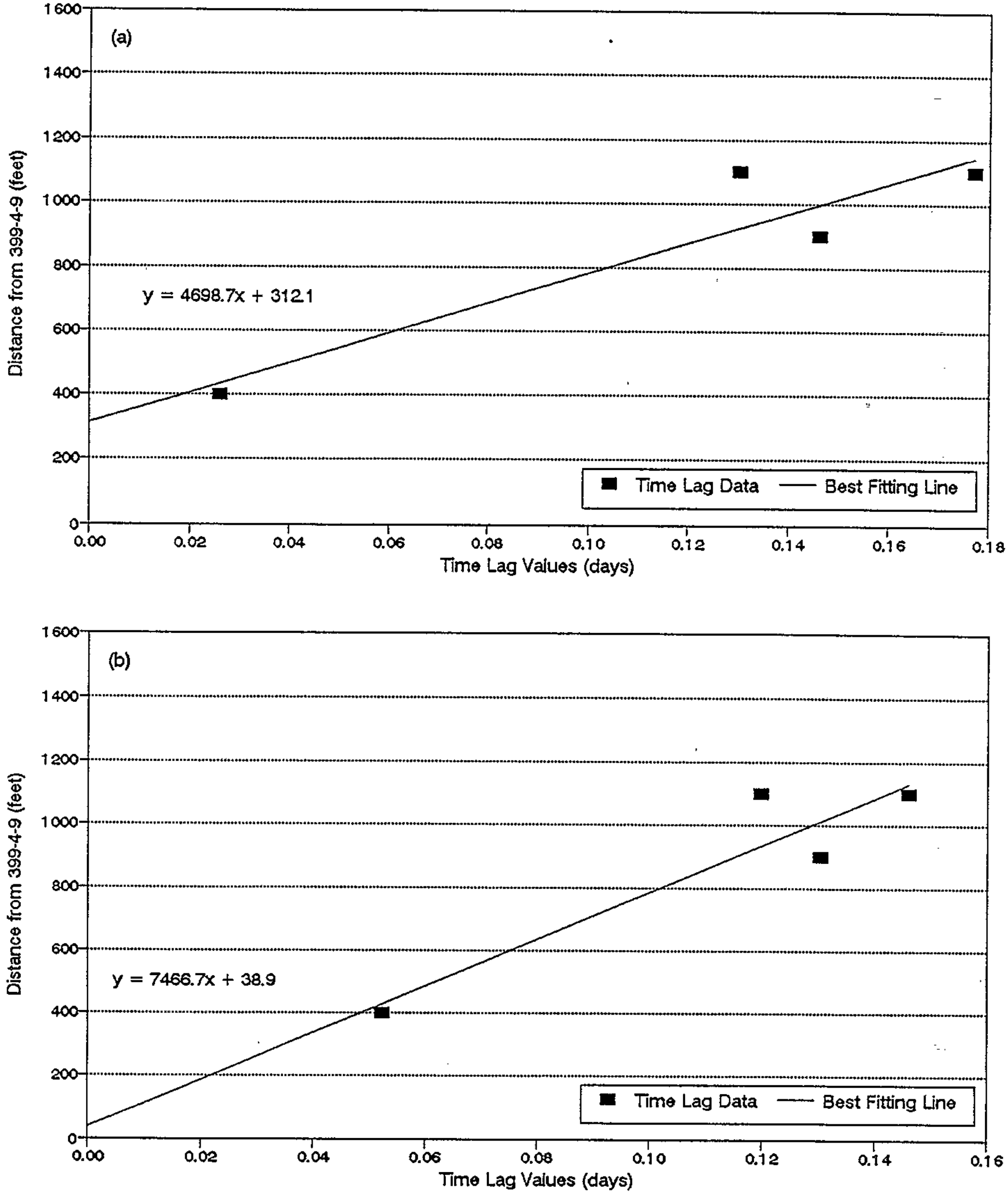


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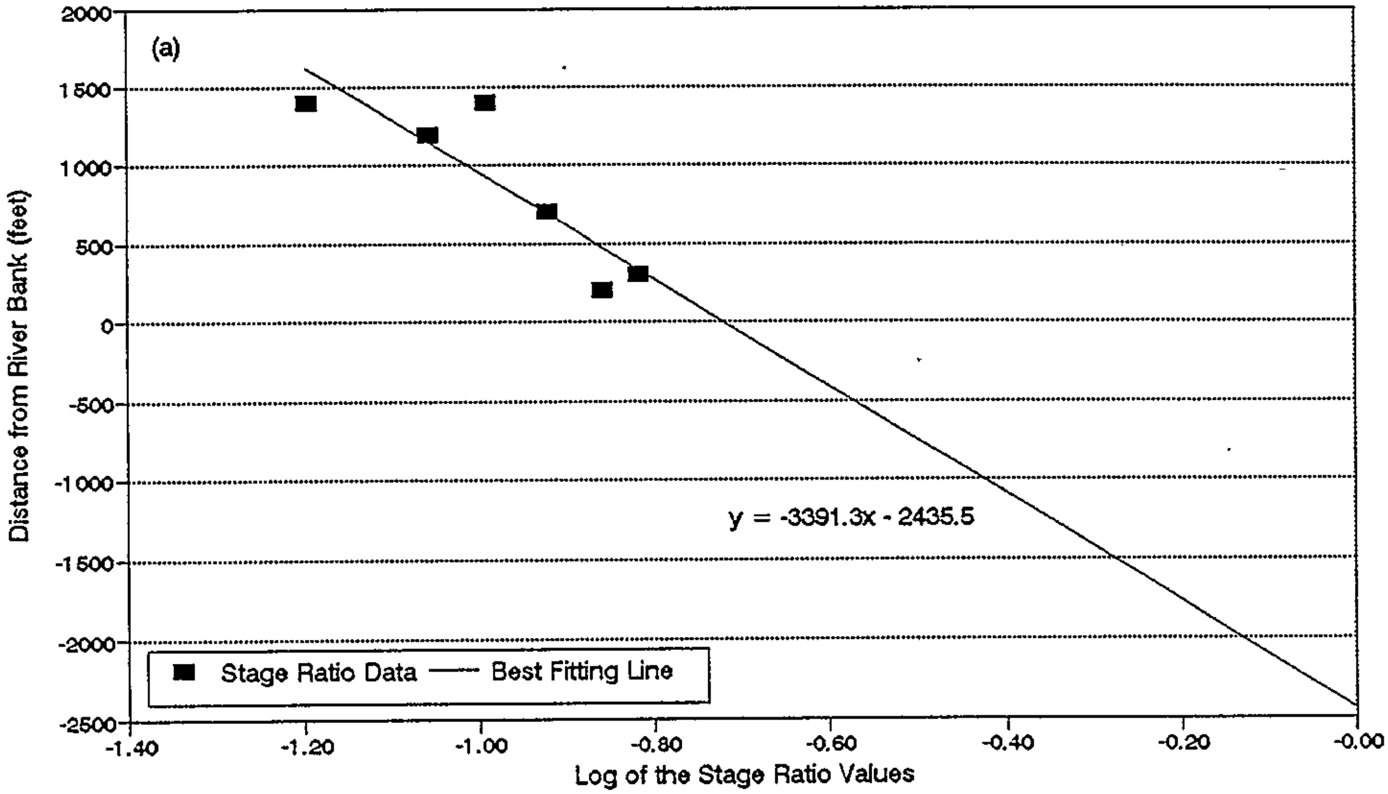
Figure C-6. Stage Ratio Method Regression Plots Using SWS-1 as the Source for (a) May 17-21 and (b) May 25-29, 1992.



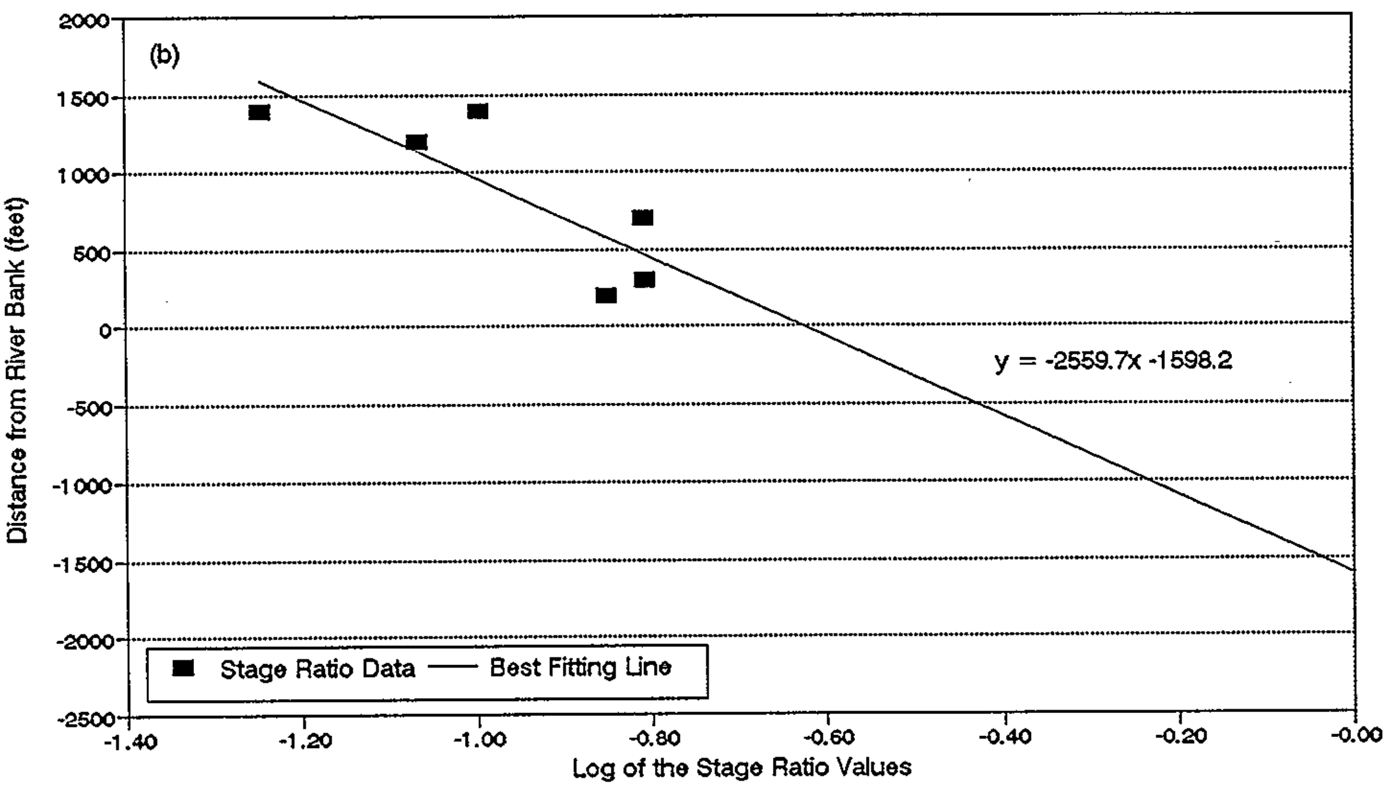


Figure C-7. Stage Ratio Method Regression Plots Using 399-4-9 as the Source for (a) May 17-21 and (b) May 25-29, 1992.

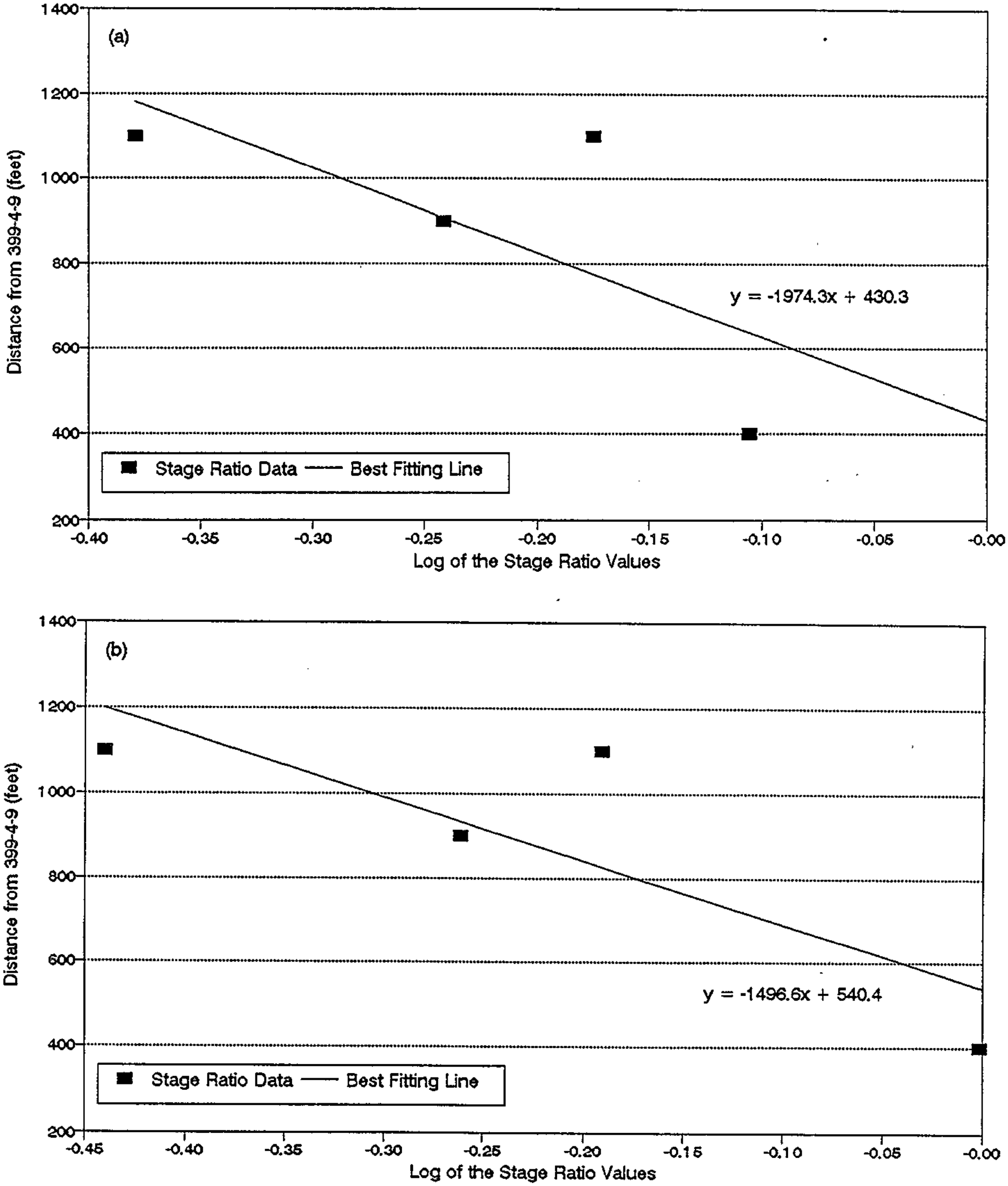


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Figure C-8. Apparent Tidal Efficiency Method Regression Plots Using SWS-1 as the Source for (a) Near-River Wells and (b) Inland Wells.

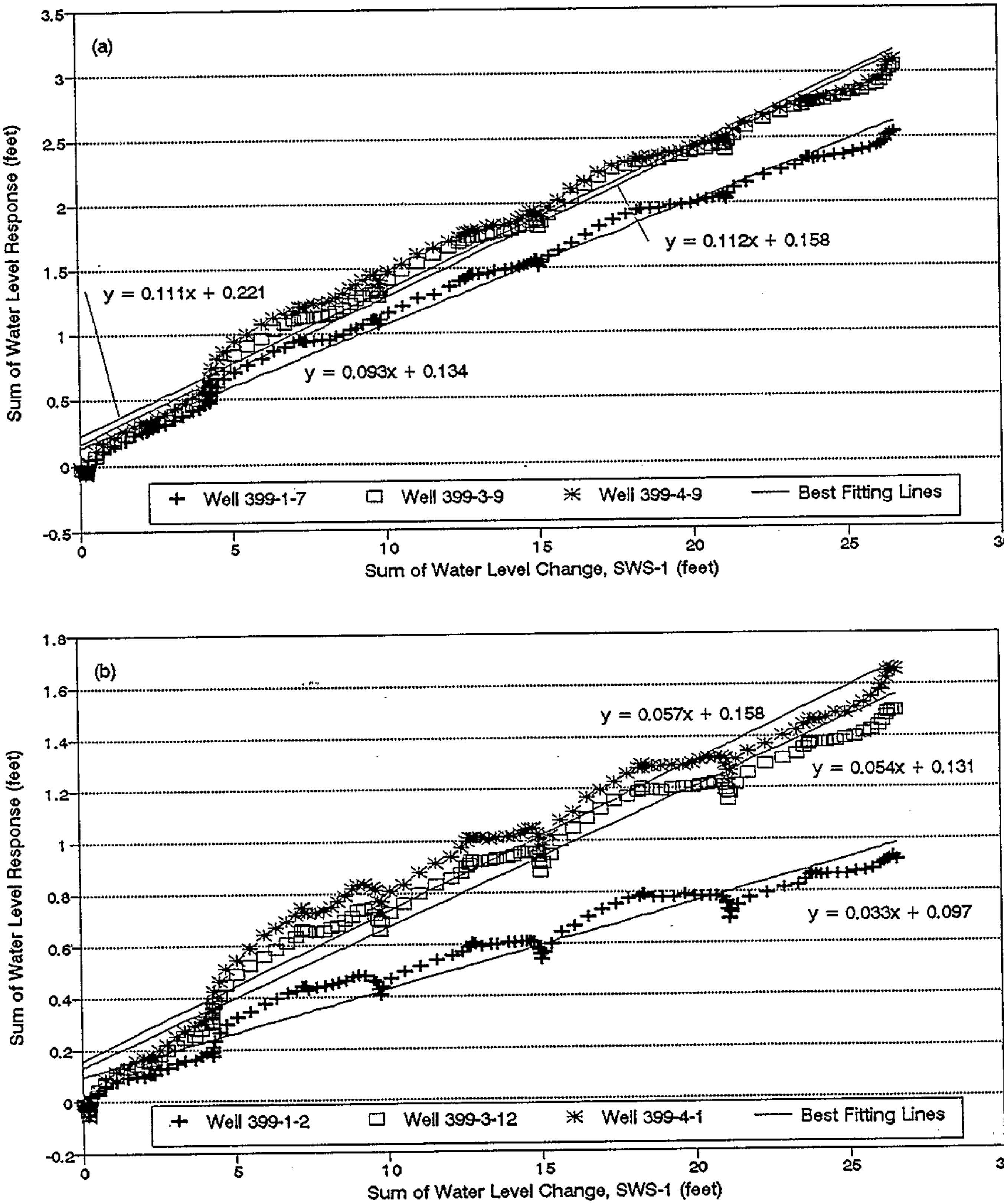


Figure C-9. Apparent Tidal Efficiency Method Regression Plots Using 399-4-9 as the Source for (a) Near-Source Wells and (b) Distant Wells.

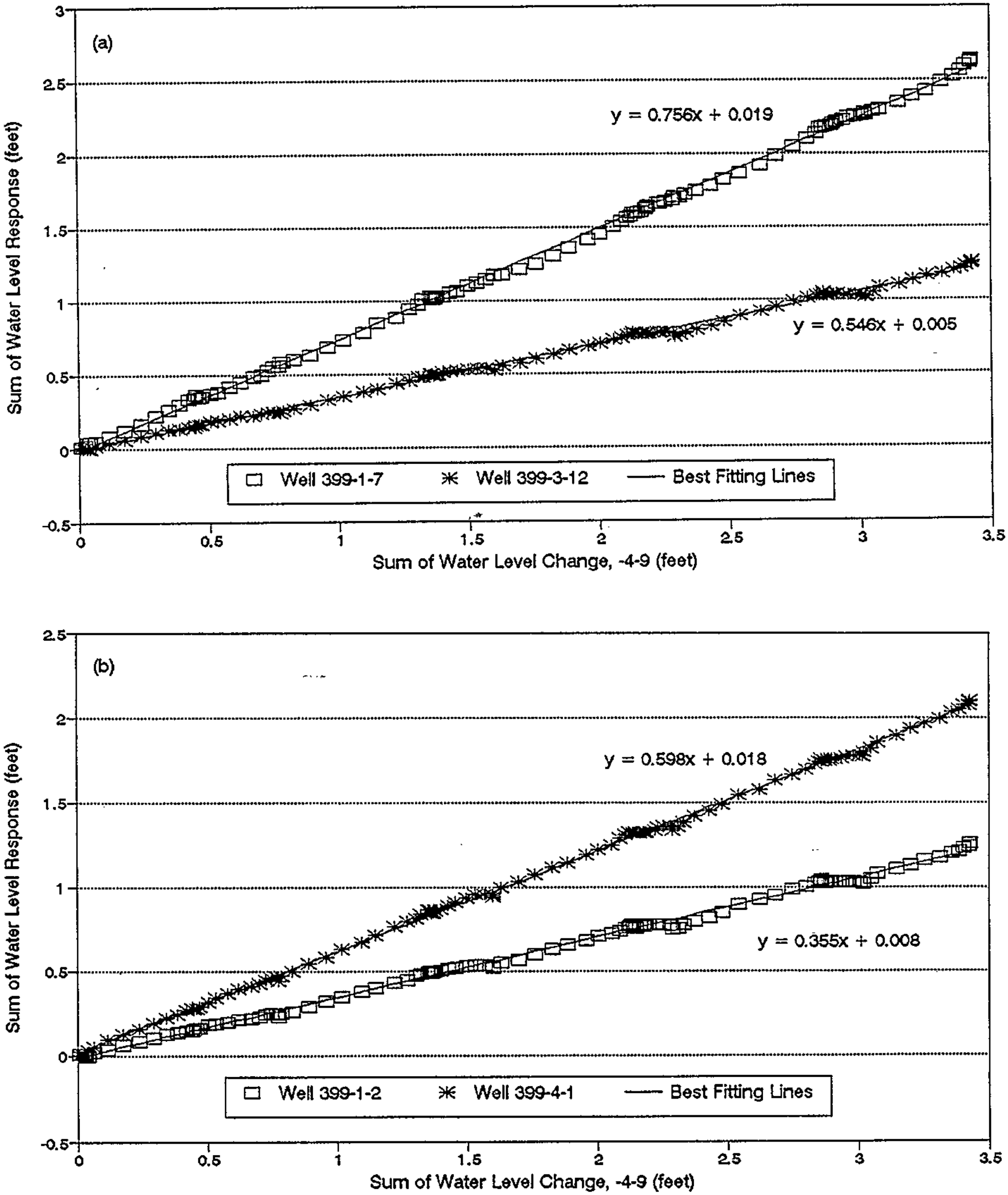
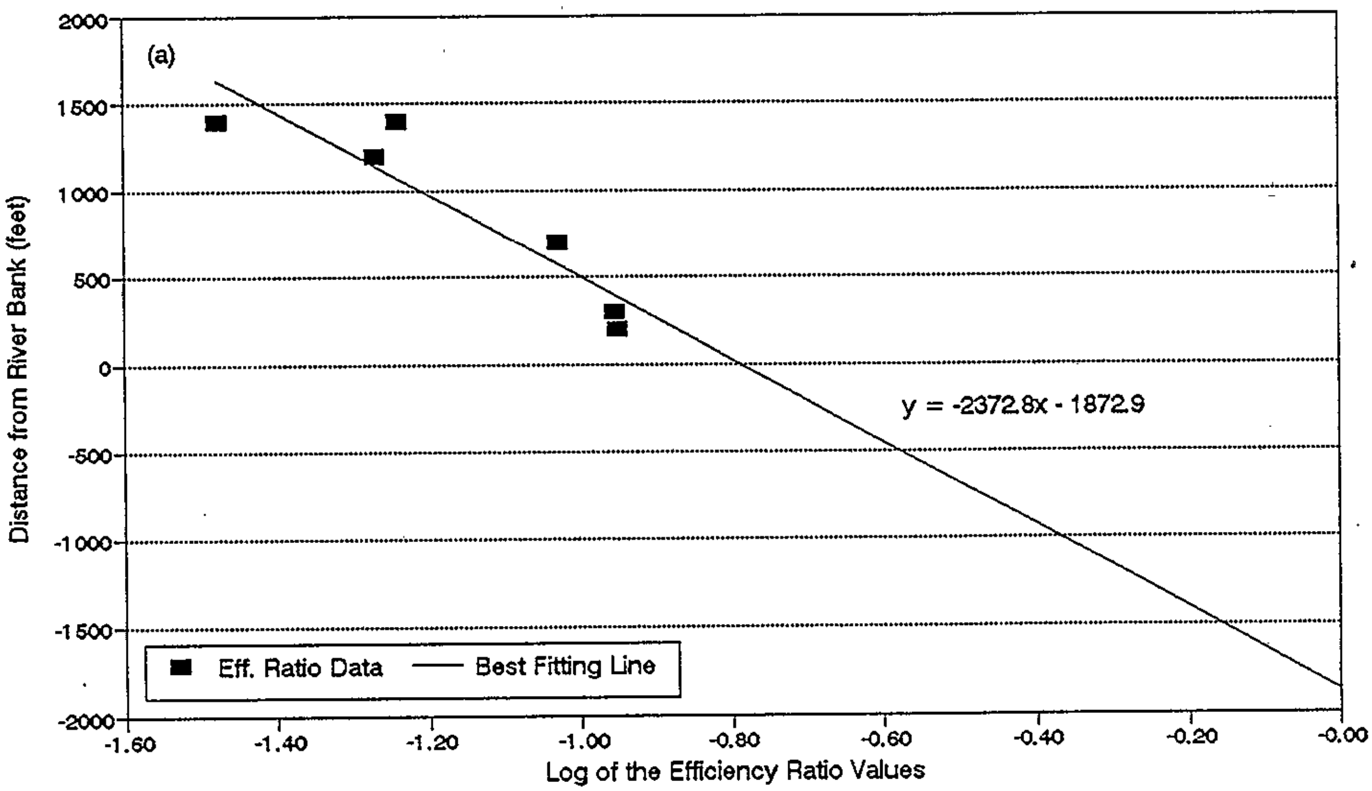


Figure C-10. Efficiency Ratio Regression Plots For May 17-21, 1992 Using (a) SWS-1 as the Source and (b) 399-4-9 as the Source.



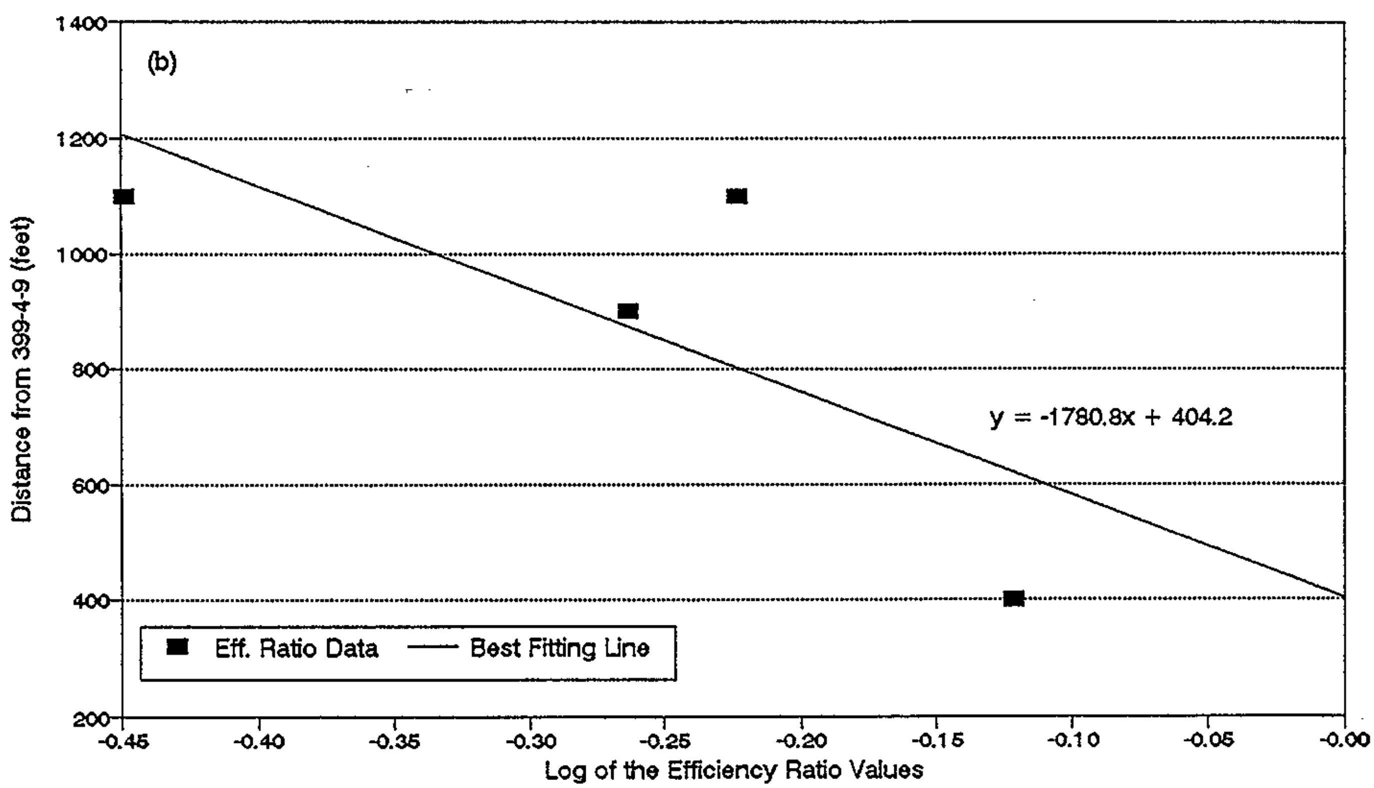


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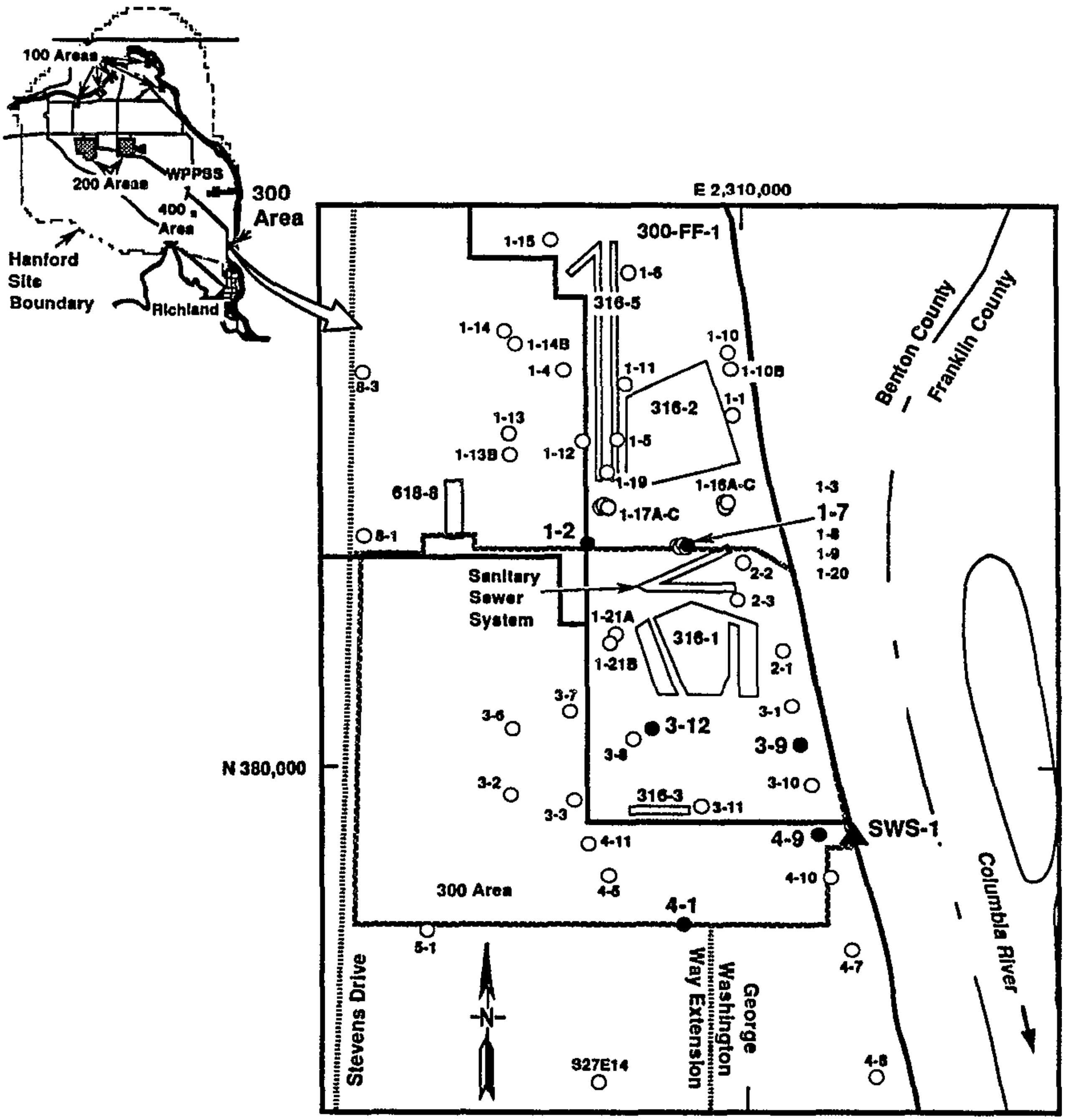
Figure C-11. Location Map for 300 Area Wells and River Stage Recorders.



o 1-12 Well Location and Number (Wells Prefixed by 399-, Except Those Beginning with S are Prefixed with 699-)

△ SWS-1 Surface-Water Monitoring Station

Roads

H9209027.1

APPENDIX D

BIERSCHENK (1959) ANALYSIS OF CYCLIC FLUCTUATIONS

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APPENDIX D

BIERSCHENK (1959) ANALYSIS OF CYCLIC FLUCTUATIONS

The following five pages are reproduced from Bierschenk (1959).[1] His work is the first known use of the Ferris method (Ferris 1952; 1963)[2,3] on the Hanford Site to infer aquifer hydraulic properties. A location map for the wells that Bierschenk used is included.

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[1] Bierschenk, W. H., 1959, *Aquifer Characteristics and Ground-Water Movement at Hanford*, HW-60601, June 1959, Hanford Atomic Products Operation, General Electric Company, Richland, Washington.

[2] Ferris, J.G., 1952, "Cyclic Fluctuations of Water Level as a Basis for Determining Aquifer Transmissibility," *U.S. Geological Survey, Ground-Water Note*, No. 1, April 1952.

[3] Ferris, J.G., 1963, "Cyclic Water Level Fluctuations as a Basis for Determining Aquifer Transmissibility," in R. Bentall (compiler), *"Methods of Determining Permeability, Transmissibility, and Drawdown," U.S. Geological Survey Water-Supply Paper 1536-I*, Washington, D.C., pp. 305-318.

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APPENDIX VI

ESTIMATING TRANSMISSIBILITY FROM CYCLIC FLUCTUATION DATA

Ferris[32] has shown that the equation for the range of ground-water fluctuation in an observation well of known distance from the aquifer contact with the surface-water body, whose stage changes sinusoidally, has the non-dimensional form:

$$s_r = 2\, s_o\, e^{-4.8\, X \sqrt{\frac{S}{t_o T}}} \qquad (17)$$

where

s_r = range in ground-water stage, in feet,
s_o = amplitude or half range of river stage, in feet,
X = distance from the observation well to the surface-water contact with the aquifer ("suboutcrop"), in feet,
t_o = period of the stage fluctuation, in days,
S = coefficient of storage,
T = coefficient of transmissibility, gpd/ft.

For convenience equation (17) can be written:

$$2.1 \sqrt{\frac{S}{t_o T}} = \frac{-\log_{10}\left(\frac{s_r}{2\, s_o}\right)}{X} \qquad (18)$$

The right-hand member of equation (18) may be represented as a slope by plotting on semilog paper the logarithm of the average range ratio ($s_r/2s_o$) for each well against the respective distance (X) of each well from the river. If the change in logarithm of the range ratio is selected over one log cycle, the numerator of this slope expression reduces to unity. Thus, equation (18) may be reduced to $T = 4.4\, (\Delta X)^2\, S/t_o$. Figure 16 roughly illustrates this method, and shows the plotted points for five wells which are located in the eastward trending glaciofluviatile channel north of Gable

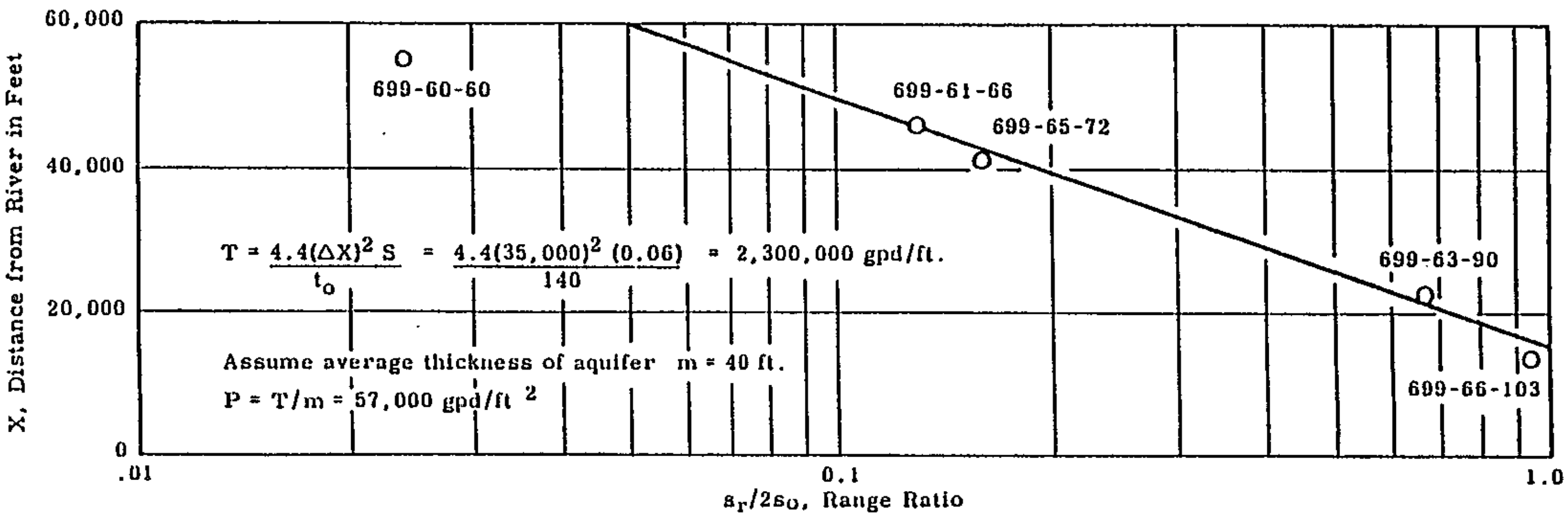

FIGURE 16

Semilog Plot of the Ratio of Ground-Water Stage to Stream Stage vs. Distance from Well to Cyclic Stream

Butte (for well location, see map, Figure 1). The data for well 699-60-60 were discounted because the hydrograph of the well indicates that the water level is influenced also by artificial recharge which masks the ground-water range due to the influence of the river.

Table XIII includes the data from which estimates of transmissibility were made for the aquifers penetrated by 15 wells in which the water level fluctuated in response to changes in Columbia River stage. The range in ground-water stage (s_r) was averaged for the period of record as was the range of river stage ($2\,s_o$). Inasmuch as the river fluctuation is not strictly sinusoidal but generally occurs as a single sharp crest each year, the period of the river fluctuation (t_o) was taken as an average of 140 days. As indicated by preceding equations, it is necessary that the coefficient of storage S be known in order to evaluate T. Only a few data are available giving values for S at Hanford, but where it has been calculated,(6, 12, 18) a range within 0. 06 to 0. 10 appears reasonable.

The indicated values (Table XIII) of the coefficient of transmissibility should be considered tentative. However, these data serve to demonstrate the applicability of the method described for analyzing cyclic fluctuations of ground-water level. The results, except for several inordinately large values, appear to be within the correct order of magnitude of transmissibility as derived previously for sites elsewhere on the project. The estimates of permeability were made assuming various effective thicknesses for the aquifers.

TABLE XIII

TRANSMISSIBILITY COEFFICIENTS ESTIMATED FROM CYCLIC FLUCTUATION DATA

Well Number	Years of Record	Average Ground-Water Range s_r (feet)	Distance from River (feet)	Average Range Ratio ($s_r/2s_o$)	Transmissibility (gpd/ft) when S = 0.06	Transmissibility (gpd/ft) when S = 0.10	Estimated Permeability (gpd/ft^2) when S = 0.06	Estimated Permeability (gpd/ft^2) when S = 0.10
699-60-60	6	0.45	55,000	0.024				
-61-66	3	2.40	46,000	0.13	2,300,000	3,800,000	57,000	95,000
-65-72	8	3.00	41,000	0.16				
-63-90	9	12.75	23,000	0.67				
-66-103	6	17.80	14,000	0.94				
-57-29	12	4.44	10,000	0.28	610,000	1,000,000	17,000	29,000
-62-32	12	4.62	11,000	0.29	780,000	1,300,000	23,000	37,000
-63-25	3	3.70	4,000	0.24	80,000	130,000	1,100	1,700
-67-77	12	3.79	5,500	0.20	115,000	190,000	960	1,600
-70-68	4	2.99	7,000	0.16	145,000	240,000	1,200	2,000
-71-84	12	6.62	1,000	0.35	9,000	15,000	90	150
-72-88	12	10.79	1,000	0.57	31,000	51,000	310	510
-92-38	9	6.52	1,200	0.42	19,000	32,000	190	320
-97-48	12	3.82	4,000	0.23	74,000	120,000	500	800
-HAN-23	12	4.50	5,000	0.28	155,000	260,000	3,700	6,200
	3	(from ref. 18)			520,000	860,000	6,100	10,000

Columbia River at Area	Years of Record	Average River-Stage Range $2s_o$ (feet)
B	12	19.1
D	12	15.9
H	12	15.6
F	12	16.0

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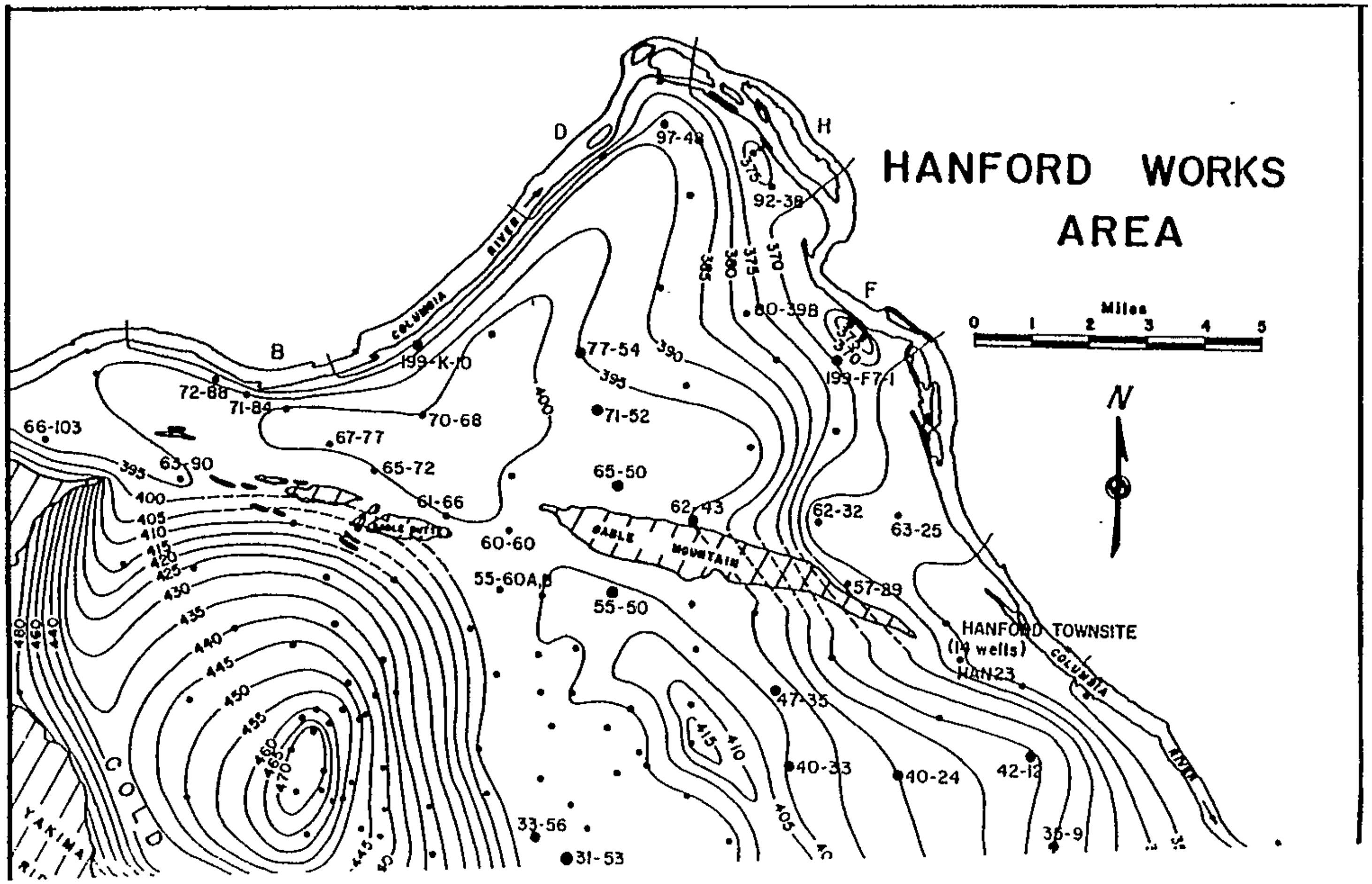

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APPENDIX E

BIBLIOGRAPHY

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APPENDIX E

BIBLIOGRAPHY

The following bibliography includes all the references cited earlier in this report, as well as additional references that pertain to the general topic of the report.

Aimo, N. J., 1987, *Preliminary Numerical Study of the Effects of Fluctuating River Level on Solute Transport Within Riverbank Aquifers*, Master's Thesis, Washington State University, Pullman, Washington.

Aimo, N. J., 1991, *Stochastic Analysis of River Stage Fluctuations and River Bank Contaminant Concentrations*, Unpublished Report, Department of Hydrology and Water Resources, University of Arizona, Tucson, Arizona.

Bear, J., 1972, *Dynamics of Fluids in Porous Media*, American Elsevier, New York, New York.

Bierschenk, W. H., 1959, *Aquifer Characteristics and Ground-Water Movement at Hanford*, HW-60601, June 1959, Hanford Atomic Products Operation, General Electric Company, Richland, Washington, pp. 29-31 and Appendix VI.

Brown, D. J. and P. P. Rowe, 1960, *100-N Aquifer Evaluation*, HW-67326, November 1960, Hanford Atomic Products Operation, General Electric Company, Richland, Washington, 14 pp. [Note: Published after Rowe, 1960; does not contain corrections noted by Hantush, 1961].

Campbell, M. D. and D. R. Newcomer, 1992, *Automatic Measurement of Water Levels Within the 300-FF-5 Boundary*, PNL-7874, April 1992, Pacific Northwest Laboratory, Richland, Washington.

Carr, P. A. and G. S. Van Der Kamp, 1969, "Determining Aquifer Characteristics by the Tidal Method," *Water Resources Research*, Vol. 5, No. 5, pp. 1023-1031.

Carslaw, H. S. and J. C. Jaeger, 1959, *Conduction of Heat in Solids*, 2nd Edition, Oxford University Press, London and New York, New York.

Clark, W. E., 1967, "Computing the Barometric Efficiency of a Well," *American Society of Civil Engineering, Journal of the Hydraulics Division*, Vol. HY 4, pp. 93-98.

Cooper, H. H., Jr. and M. I. Rorabaugh, 1963, "Ground-Water Movements and Bank Storage Due to Flood Stages in Surface Streams," *Geological Survey Water-Supply Paper 1536-J*, pp. 343-366.

DOE-RL, 1990, *Remedial Investigation/Feasibility Study Work Plan for the 300-FF-5 Operable Unit, Hanford Site, Richland, Washington*, DOE/RL-89-14, June 1990, U.S. Department of Energy, Richland Field Office, Richland, Washington.

DOE-RL, 1992, *Remedial Investigation/Feasibility Study Work Plan for the 100-BC-5 Operable Unit, Hanford Site, Richland, Washington*, DOE/RL-90-08, Rev. 0, July 1992, U.S. Department of Energy, Richland Field Office, Richland, Washington.

Erskine, A. D., 1991, "The Effect of Tidal Fluctuation on a Coastal Aquifer in the UK," *Ground Water*, Vol. 29, No. 4, pp. 556-562.

Ferris, J. G., 1951, "Cyclic Fluctuations of Water Level as a Basis for Determining Aquifer Transmissibility," *International Association of Scientific Hydrology*, Publication 33, pp. 148-155.

Ferris, J. G., 1952, "Cyclic Fluctuations of Water Level as a Basis for Determining Aquifer Transmissibility," *U.S. Geological Survey, Ground-Water Note*, No. 1, April 1952.

Ferris, J. G., 1963, "Cyclic Water-Level Fluctuations as a Basis for Determining Aquifer Transmissibility," in R. Bentall (compiler), *"Methods of Determining Permeability, Transmissibility, and Drawdown," U.S. Geological Survey Water-Supply Paper 1536-I*, Washington, D.C., pp. 305-318.

Ferris, J. G., D. B. Knowles, R. H. Brown, and R. W. Stallman, 1962, "Theory of Aquifer Tests," *Geological Survey Water-Supply Paper 1536-E*, pp. 132-135.

Gambolati, G., F. Toffolo, and F. Uliana, 1984, "Groundwater Response Under an Electronuclear Plant to a River Flood Wave Analyzed by a Nonlinear Finite Element Model," *Water Resources Research*, Vol. 20, No. 7, pp. 903-913.

Gilmore, T. J., J. V. Borghese, H. P. McDonald, and D. R. Newcomer, 1990, *Evaluation of the Effects of the Columbia River on the Unconfined Aquifer Beneath the 1301-N Liquid Waste Disposal Facility*, PNL-7341, May 1990, Pacific Northwest Laboratory, Richland, Washington.

Gilmore, T. J., J. V. Borghese, and D. R. Newcomer, 1991, *Evaluation of the Effects of the Columbia River on the Unconfined Aquifer Beneath the 100-N Area*, PNL-7646, May 1991, Pacific Northwest Laboratory, Richland, Washington.

Gilmore, T. J., F. A. Spane, Jr., D. R. Newcomer, and C.R. Sherwood, 1992, *Application of Three Aquifer Test Methods for Estimating Hydraulic Properties Within the 100-N Area*, PNL-8335, September 1992, Pacific Northwest Laboratory, Richland, Washington.

Gregg, D. O., 1966, "An Analysis of Ground-Water Fluctuations Caused by Ocean Tides in Glynn County, Georgia," *Ground Water*, Vol. 4, No. 3, pp. 24-32.

Grubb, H. F. and H. H. Zehner, 1973, "Aquifer Diffusivity of the Ohio River Alluvial Aquifer by the WYLW (Floodwave) Response Method," *U.S. Geological Survey, Journal of Research*, Vol. 1, No. 5, pp. 597-601.

Hall, F. R. and A. F. Moench, 1972, "Application of Convolution Equations to Stream/Aquifer Relationships," *Water Resources Research*, Vol. 8, No. 2, pp. 487-493.

Hanson, R. L. and S. G. Brown, 1972, "Subsurface Hydraulics in the Area of the Gila River Phreatophyte Project, Graham County, Arizona," *Geological Survey Professional Paper 655-F*, U.S. Geological Survey, Washington, D.C., pp. F14-F15.

Hantush, M. S., 1961, "Discussion of Paper by P.P. Rowe, 'An Equation for Estimating Transmissibility and Coefficient of Storage From River Level Fluctuations,'" *Journal of Geophysical Research*, April 1961, Vol. 66, No. 2, pp. 1310-1311.

Hornberger, G. M., J. Ebert, and I. Remson, 1970, "Numerical Solution of the Boussinesq Equation for Aquifer-Stream Interaction," *Water Resources Research*, Vol. 6, No. 2, pp. 601-608.

Hvorslev, M. J., 1951, *Time Lag and Soil Permeability in Ground-Water Observations*, Bulletin No. 36, U.S. Waterways Experimental Station, Vicksburg, Mississippi.

Jacob, E. C., 1950, "Flow of Ground Water," in H. Rouse, (ed.), *Engineering Hydraulics*, John Wiley, New York, pp. 321-386.

Kelly, W. E., 1976, "Representing Aquifer-River Interaction," *Advances in Groundwater Hydrology*, American Water Resources Association, pp. 262-270.

Kraeger-Rovey, C. E., 1975, *Numerical Model of Flow in Aquifer-River System*, Hydrology Paper 74, Colorado State University, Fort Collins, Colorado.

Morel-Seytoux, H. J., 1975, "A Combined Model of Water Table and River Stage Evolution," *Water Resources Research*, Vol. 11, No. 6, pp. 968-972.

Muskat, M. K., 1937, *Flow of Homogeneous Fluids Through Porous Media*, McGraw-Hill, New York, New York.

Neuman, S. P., 1974, "Effect of Partial Penetration on Flow in Unconfined Aquifers Considering Delayed Gravity Response," *Water Resources Research*, Vol. 10, pp. 303-312.

Newcomb, R. C. and S. G. Brown, 1961, "Evaluation of Bank Storage Along the Columbia River Between Richland and China Bar, Washington," *Geological Survey Water-Supply Paper 1539-I*, pp. I-1 to I-13.

Newcomer, D. R., 1988, *Detailed Water-Level Data and an Evaluation of Mathematical Approaches for Near-River Monitoring*, Master's Thesis, Montana College of Mineral Science and Technology, Butte, Montana.

Pinder, G. F., J. D. Bredehoeft, and H. H. Cooper, 1969, "Determination of Aquifer Diffusivity from Aquifer Response to Fluctuations in River Stage," *Water Resources Research*, Vol. 5, No. 4, pp. 850-855.

Pinder, G. F. and S. P. Sauer, 1971, "Numerical Simulation of Flood Wave Modification Due to Bank Storage Effects," *Water Resources Research*, Vol. 7, No. 1, pp. 63-70.

Poeter, E. and D. R. Gaylord, 1990, "Influence of Aquifer Heterogeneity on Contaminant Transport at the Hanford Site," *Ground Water*, Vol. 28, No. 6, November-December 1990, pp. 900-909.

Reynolds, R. J., 1987, "Diffusivity of a Glacial-Outwash Aquifer by the WYLW (Floodwave) Response Technique," *Ground Water*, Vol. 25, No. 3, pp. 290-299.

Rowe, P. P. 1960, "Equations for Estimating Transmissibility and Coefficient of Storage from River Level Fluctuations," *Journal of Geophysical Research*, October 1960, Vol. 65, No. 10, pp. 3419-3424. [Hantush (1961) contains corrections for these equations]

Schalla, R., (ed.), 1988, *Interim Characterization Report for the 300 Areas Process Trenches*, PNL-6716, September 1988, Pacific Northwest Laboratory, Richland, Washington.

Serfes, M. E., 1987, "Interpretation of Tidally Affected Ground-Water Flow Systems in Pollution Studies," in *Proceedings, Petroleum Hydrocarbons and Organic Chemicals in Ground Water: Prevention, Detection, and Restoration -- A Conference and Exposition*, National Water Well Association, Dublin, Ohio, pp. 55-73.

Serfes, M. E., 1991, "Determining the Mean Hydraulic Gradient of Ground Water Affected by Tidal Fluctuations," *Ground Water*, Vol. 29, No. 4, pp. 549-555.

Singh, S. R. and B. Sagar, 1977, "Estimation of Aquifer Diffusivity in Aquifer - River Systems," *American Society Civil Engineering, Journal Hydrology Division*, Vol. 103, No. HY11, pp. 1293-1302.

Spane, F. A., 1991, "Evaluation of Pumping Test Analyses Reported in Schalla et al. (1988)," Letter Report, August 13, 1991, Pacific Northwest Laboratories to Westinghouse Hanford Company.

Spane, F. A., 1992, "Barometric and River Stage Fluctuation Removal From Hydraulic Characterization Test Data," Letter Report, July 14, 1992, Pacific Northwest Laboratories to Westinghouse Hanford Company.

Spiegel, Z., 1962, *Hydraulics of Certain Stream-Connected Aquifer Systems*, New Mexico State Engineer, Special Report, Santa Fe, New Mexico.

Todd, D. K., 1955, "Ground-Water Flow in Relation to a Flooding River," *American Society of Civil Engineering Proceedings*, Vol. 81, Separate No. 628, pp. 1-20.

Walton, W. C., 1970, "Use of Stream-Stage Changes in Estimating Hydrogeologic Properties of Aquifers," in *Groundwater Resource Evaluation*, McGraw-Hill Book Company, New York, New York, pp. 176-181.

Weeks, E. P., 1979, "Aquifer Tests--The State of the Art in Hydrology," *Proceedings of the Invitational Well-Testing Symposium, October 19-21, 1977, Berkeley, California*, LBL-7027, Lawrence Berkeley Laboratory, Berkeley, California.

Werner, P. W. and D. Nóren, 1951, "Progressive Waves in Nonartesian Aquifers," *Transactions of the American Geophysical Union*, Vol. 32, pp. 238-244.

Zitta, V. L. and J. M. Wiggert, 1971, "Flood Routing in Channels With Bank Seepage," *Water Resources Research*, Vol. 7, No. 5, pp. 1341-1345.

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