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Interactions Between Surface Water and Groundwater in a Virginia Coastal Plain Watershed

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Diana L. Weigmann
Assistant Director

Abstract

A field study of teractions of groundwater and surface water during both baseflow and stormflow conditions was performed at the Reedy Creek watershed in the Virginia Coastal Plain. Several watershed and hillslope scale experiments (including an investigation of the watershed's hydrogeological setting) indicated that surface-water baseflow is supported entirely by drainage of shallow groundwater from a relatively thin (1-6 m thick) water-table aquifer. This relatively permeable surficial aquifer was found to be underlain by dark, olive-gray, clay-silt and diatomaceous Miocene deposits of low permeability known as the Calvert Formation, which is believed to function as an aquiclude in the area.

Estimates of the mean saturated hydraulic conductivity (K_s) of the water-table aquifer using three different methods were in reasonable agreement (within about a factor of three of 10^{-2} cm/sec), suggesting that Reedy Creek baseflow can be attributed solely to shallow groundwater discharge. A chemical hydrograph separation technique was used to resolve the contributions of old and new water to stormflow. Results from six rainfall events indicated that old water dominated the stormflow response of the watershed, although the new-water contribution often approached 40% at the hydrograph peak. Computed new-water flow contributing areas were in reasonable agreement with estimated areas of likely saturation (channels, ponds, floodplains) based on digitized topographic maps.

These results suggested that stormflow at Reedy Creek is primarily generated by a saturation overland flow mechanism from variable source areas that usually include the stream channels and some significant proportion of the riparian wetland area. Additional field studies of runoff generation, including observations of surface saturation behavior and water-table responses within riparian zones and topographically convergent sideslopes, would be useful in further testing of this hypothesis.

Key words: groundwater, groundwater-surface water interactions, nontidal wetlands, watershed hydrology, hillslope hydrology, riparian zone, streamflow, water balance.

1. Introduction

There are few watersheds for which hydrological flowpaths and streamflow generation mechanisms have been quantitatively elucidated. However, assessing the impacts of human activities on water quality is dependent upon an understanding of the interactions between groundwater and surface water and on the physical mechanisms of storm runoff generation.

Modern theories of streamflow generation are challenging the classical Hortonian theory of infiltration-excess overland flow, particularly in vegetated watersheds in humid, temperate regions. These relatively new theories recognize several mechanisms (subsurface stormflow and saturation overland flow) whereby groundwater is actually the dominant source of stormflow (Freeze, 1974) and emphasize the hydrological importance of riparian wetlands in streamflow generation (Dunne et al., 1975; Beven, 1978; Taylor, 1982).

The question of whether stormflow is comprised primarily of water flowing over the ground, as opposed to flowing through the ground, is significant because the former flowpath allows for little soil contact, while the latter permits soil contact. If soil contact occurs, the residence time of water in the ground and the extent of contact will largely determine the degree of assimilation of inorganic nutrients, degradation of organic pollutants, and neutralization of acidic deposition. Hydraulic residence time then will be largely determined by the specific hydrogeological setting of a particular watershed.

1.1 Objectives

The overall goal of this project was to elucidate watershed and hillslope-scale, groundwater-surface water interactions, and storm runoff generation processes within a Mid-Atlantic Coastal Plain watershed through an intensive field study involving hydrometric techniques and the use of natural tracers. The fundamental hypothesis of the research was that shallow groundwater and surface water are tightly coupled in this region; conceptually, it was hypothesized that baseflow is sustained entirely by shallow groundwater discharge, while saturation overland flow (including both return flow of old water and direct precipitation of new water onto stream channels and saturated areas) is the dominant source of stormflow in streams in this region.

It was hypothesized that riparian wetlands are the immediate source of both return flow and direct precipitation in Mid-Atlantic Coastal Plain watersheds during stormflow periods, and that groundwater discharge through these wetlands is the major source of stream baseflow. Wetlands, therefore, play an important hydrological role in influencing both the quantity and quality of water discharged into surface waters. Further, it was hypothesized that upland hillslopes with convergent topography and shallow aquicludes also can contribute groundwater during baseflow periods, and return flow and direct precipitation during stormflow events.

The specific research objectives of this one-year project were to:

- establish a gaged watershed in the Virginia Coastal Plain for conducting watershed- and hillslope-scale hydrological and hydrochemical research,
- describe the watershed's hydrogeological setting and qualitatively describe shallow groundwater-surface water interactions,
- quantify spatial variability of baseflow within the watershed,
- compare three independent, quantitative estimates of the saturated hydraulic conductivity of the water-table aquifer, and
- quantify contributions of old and new water to streamflow during several stormflow events using a classical chemical hydrograph separation technique.

1.2 Study Site Description

The Reedy Creek watershed in Caroline County, Virginia, (Figure 1) is a tributary of the Mattaponi River, which is part of the York River basin. The watershed is located in a rural area 42 km north of Richmond just east of Interstate 95. Land use in the watershed is mixed, with large tracts subjected to agricultural and silvicultural practices; currently, the watershed is influenced minimally by suburban, commercial, and industrial development, although the region surrounding the I-95 corridor is one of the fastest growing areas in the eastern U.S. (Chesapeake Executive Council, 1988). The total watershed area at a gaging station just upstream of U.S. Rte. 301 is 4506 ha, with ground surface elevations of 100-220 ft above mean sea level.

Reedy Creek was selected randomly and sampled by the National Stream Survey (NSS) as part of a comprehensive regional study of stream acidification in the eastern U.S. (Kaufmann et al., 1988; Sale et al., 1988). The site was selected as appropriate for this study based on chemical and physical data collected during the NSS and on its location within two hours driving time from Charlottesville. In particular, it was noted that the alkalinity of Reedy Creek during spring baseflow conditions was low (< 25 μeq/L), suggesting that water that ultimately becomes streamflow has a relatively short subsurface residence time within the watershed.

2. Methodology

2.1 Hydrogeological Characterization

The hydrogeological setting of the watershed was characterized on the basis of stratigraphic strip logs compiled from sediment samples taken from 12 hand-augured well borings. The well borings were located along two transects that crossed an upland sideslope and adjacent riparian floodplain along Reedy Creek (Figures 1, 2). Each transect dissected a topographic divide, the direction of which was roughly perpendicular to the creek. Transect T1 was approximately 200 m long; transect T2 was about 150 m long. Each well boring was approximately 12 cm in diameter—large enough to accommodate a 2-in PVC well casing and well screen. Depths from which the boring samples were taken were recorded in the field. Samples were transported in plastic bags to the laboratory, where they were correlated and described. Sediment columns were interpreted using 1:24,000-scale geologic maps of the Ashland (Weems, 1986) and Ruther Glen (Rader, in preparation) quadrangles and a 1:250,000-scale map of the Virginia Coastal Plain and parts of the adjacent Piedmont (Mixon et al., 1989).

2.2 Surface Water Gaging and Water Balance

A continuously recording stream gage was installed in early January 1990 near the bridge where Rte. 662 crosses Reedy Creek (site RC2); the watershed area associated with this location is 4506 ha (Figure 1). A Stevens Type A recorder equipped with a 6-in float and counterweight was used to provide a continuous record of creek stage over the duration of the study. The recording device was placed in an instrument shelter mounted above a stilling well, which consisted of a 5-ft length of corrugated, galvanized steel (21-in diameter, 16-gauge) culvert pipe. A well was dug manually into the creek bank to the lowest depth of the stream channel bed, and the culvert pipe was set in place. A 2-in hole cut in the culvert at the stream bed level, fitted with a 10-ft length of PVC pipe, served as a conduit for water exchange between the creek and the well through a hand-dug trench. After completing the installation, the well and trench were backfilled with sediment and stabilized with gravel.

A 1-m Stevens porcelain staff plate was mounted onto a pier along the southwest side of the bridge at site RC2 and used as a nonrecording gage. Reference stage was determined on each visit to the site. Discharge measurements made on a regular basis were used to define the rating for the gage; all measurements were made using the midsection method of the U.S. Geological Survey (USGS) (Rantz et al., 1982). All velocity measurements were made using a Marsh-McBirney Model 201M portable electromagnetic current meter mounted to a steel wading rod. Rating curves relating stage to discharge were developed for the gage, although it was necessary to use two different equations to account for a shift resulting from channel scouring and filling that occurred during a major flood in late May 1990. In addition, it was necessary to adjust stage records for March through September 1991 to account for shifts attributed to backwater effects resulting from construction of a beaver dam downstream of the gage.

Continuous stage records were digitized using EZDIGIT (Porter, 1987), and the digital data and appropriate rating equations were used to generate continuous hourly and daily flow records for Reedy Creek. A regression equation relating daily discharge at Reedy Creek to daily discharge at nearby Totopotomoy Creek, a stream continuously monitored by the Virginia Water Control Board, was used to supplement the Reedy Creek record for periods of missing data (October 1, 1989 - January 16, 1990 and March 28 - April 26, 1990). Annual water balances for the Reedy Creek watershed were computed using measured discharge data and precipitation data obtained from the National Oceanic and Atmospheric Administration (NOAA) for the Ashland, Virginia, station (19 km southwest of the Reedy Creek watershed).

2.3 Baseflow Variability Studies

Ten synoptic discharge experiments were performed during the study at various locations within the watershed to quantify the spatial variability of creek baseflow. In addition, computation of lateral, diffuse groundwater inflows/outflows from sideslope areas was attempted using synoptic discharge data from upstream, downstream, and tributary locations. Eleven stations along Reedy Creek (including tributaries and the main creek) were identified within the watershed (Figure 1). Drainage area divides for tributary subwatersheds and sideslope drainages were drawn on USGS quadrangles on the basis of surface topography; drainage divides were digitized, and corresponding subwatershed areas were computed using Earth Resources Data Analysis Systems (ERDAS). Discharge measurements at 9 of the 11 stations were made on various dates using the same USGS technique used at site RC2 (see section 2.2).

2.4 Hillslope/Wetland Experiments

A series of experiments were performed within the well transect area (see section 2.1) to determine the hydraulic properties of the phreatic aquifer and to examine the hydraulic linkage between the upland sideslope, the riparian wetland, and the creek. Twelve combination piezometer/sampling wells were installed along the two transects (Figure 2) using 2-in PVC casing and 2-in PVC commercial well screen. All wells were hand-augured to a depth at which an extremely low-permeability stratum was encountered. Each well consisted of an end cap, a 1-ft section of well screen, a coupling, an appropriate length of PVC pipe, and a screw-on top cap. During installation, each well was backfilled with a bed of well-sorted coarse sand, sealed using a thick slurry made from native-clay borings and water, and backfilled with excavated materials. Each well was developed by bailing for several hours after installation. Relative well elevations subsequently were determined by surveying the well transect area.

Static water levels were monitored within the well/piezometers at approximately triweekly intervals over the course of the study using a YSI field temperature-level-conductivity (T-L-C) meter. Estimates of the saturated hydraulic conductivity (K_s) of the water-table aquifer were computed using Darcy's Law for each monitoring date, assuming steady flow conditions, and making the Dupuit approximation of horizontal, one-dimensional flow normal to the creek (Freeze and Cherry, 1979):

$$Q_{lat} = -K_s A_s (dh/dl) \qquad [1]$$

where A_s is the seepage area (calculated as the product of the mean height of the water table above the low-permeability stratum at the two creekbank wells [1C and 2C] and the distance between the two transects [241 m]); dh/dl is the hydraulic head gradient across the wetland (mean computed from the two transects); Q_{lat} is the lateral groundwater discharge to the creek from the sideslope drainage area within which the transects were located; and the minus sign denotes that flow is in the direction of decreasing hydraulic head. Because of an inability to obtain reliable direct measurements of Q_{lat} from synoptic discharge measurements (see section 2.3), Q_{lat} was estimated for each date as the product of the discharge at downstream site RC2 (Q_{RC2}) and the ratio of sideslope drainage area (A_{SDA}) to the total watershed area at site RC2 (A_{RC2}):

$$Q_{lat} = Q_{RC2} A_{SDA} / A_{RC2} \qquad [2]$$

Therefore, K_s was determined by algebraically combining equations 1 and 2:

$$K_s = -Q_{lat}/A_s(dh/dl) = -[Q_{RC2} A_{SDA} / A_{RC2} A_s (dh/dl)] \qquad [3]$$

A series of bailtests (rising head slug tests) were conducted on all wells (except wells 2SM and 2SH, which had insufficient recharge) during May 1991 using the following method. First, the static water level within each well was measured; second, the well water was evacuated using a bailer (2.7-in dia. x 1.5 m long); third, the water level was recorded at appropriate time intervals (usually 30-60 seconds) during recharge. Data were analyzed according to the method of Hvorslev (Hvorslev, 1951): a plot of the logarithm of the relative increase in head vs. time theoretically should yield a linear curve. Deviations from necessary assumptions and variations in well construction, however, sometimes resulted in nonlinear curves. Thus, it was necessary to subjectively assess such plots and interpret a linear relationship to enable computation of K_s.

In addition to the 2-in observation wells, a 4-in PVC well was installed in alluvial sands in the wetland along transect T2 at a position midway between wells 2WN and 2WS (site 2WX; Figure 2). This well was installed in January 1991 and was equipped with a Stevens Type F recorder, which operated continuously from February 4 through March 31, 1991, a period in which several significant rainfall events occurred. The water-level recorder was removed from the well when clearcutting of the hillslope was initiated by the landowner in early April 1991. Analog water-level records from the well were digitized using EZDIGIT.

At each triweekly sampling date, a 500-ml sample was bailed from each well along transects T1 and T2 and taken to the laboratory for chloride analysis. Chloride concentrations were measured by ion chromatography using a Dionex 4000i instrument equipped with a fast anion (AS-4) separator column, a micromembrane (AMMS) suppressor column, and a conductivity cell. Samples were injected through an inline, 0.45-μm filter, and the resulting chromatograms were analyzed using a Spectra-Physics computing integrator. Standard curves were developed by injecting aliquots of working standards, which were made by volumetric dilution of a NaCl stock solution. The stock solution (5.000×10^{-2} M) was made from analytical grade NaCl salt. The relative precision of Cl^- analysis was estimated by analyzing duplicates in each batch of samples. The mean relative standard deviation (%) from 26 sets of duplicates was 1.9%, with all values less than 6%.

A subsurface tracer experiment was performed within the well transect area in the spring of 1991 to provide an independent estimate of K_s (Figure 2). The configuration of wells used in the tracer experiment is shown in Figure 3. The injection well (TW1) was located near the base of the hillslope (valley wall) and the monitoring wells (TW2-TW6) were located within the wetland (valley bottom). The distance between TW1 and each monitoring well was 11.4 m. The transect was oriented such that the path intersecting TW1 and TW4 is approximately normal to the stream channel at that point (Figure 2). Other paths (from TW1 to the other wells) represent 15° departures from the normal direction. Wells were installed in a manner similar to those along transects T1 and T2. On May 30, the tracer (a solution of 3600 g NaCl and 1000 g KBr dissolved in 7.6 l of deionized water) was introduced into well TW1. At each subsequent sampling over a period of three weeks, static water levels in the tracer wells were monitored and used to compute hydraulic gradients (dh/dl). Each sampling well then was bailed 10 bailer volumes (approximately 250 ml each) or until the well was dry; after fully recharging, the conductivity (specific conductance in μmhos/cm) was determined in situ using the T-L-C meter. The mean seepage velocity (v) was computed from the elapsed time from the start of the experiment (t_0) to the peak in specific conductance (t_p) and the distance between the wells (L) as follows:

$$v = L/(t_p - t_0) \qquad [4]$$

The Darcian velocity (q) was computed using the following equation, assuming a porosity (n) of 0.40:

$$q = vn \qquad [5]$$

K_s was computed using Darcy's Law and the mean hydraulic head gradient (dh/dl) over the course of the experiment.

2.5 Chemical Separations of Stormflow Hydrographs

Contributions of new and old water to storm discharge in Reedy Creek were determined using a classic chemical hydrograph separation technique (Sklash et al., 1976; Freeze and Cherry, 1979) based on the following equations for conservation of water and solute mass:

$$Q_t = Q_n + Q_o \qquad [6]$$

$$Q_t C_t = Q_n C_n + Q_o C_o \qquad [7]$$

where Q is discharge, C is concentration of a conservative natural solute tracer, and the subscripts t, n, and o represent total, new, and old, respectively. Equations 6 and 7 can be combined algebraically to yield two simple equations that can be solved for Q_n/Q_t and Q_o/Q_t, the instantaneous (proportional) contributions of new and old water to the stormflow hydrograph, respectively:

$$Q_o/Q_t = (C_t - C_n)/(C_o - C_n) \qquad [8]$$

$$Q_n/Q_t = 1 - Q_o/Q_t \qquad [9]$$

Naturally occurring chloride (Cl^-) was used to separate several storm hydrographs for which data were obtained throughout the project. Q_t was measured for each event at site RC2. A portable ISCO automatic water sampler was deployed at a location about 50 m upstream from site RC2 to provide creek-water samples for subsequent chloride analysis (determination of C_t). The automatic sampler was programmed to sample at six-hour intervals and thus could sample continuously for seven days (total capacity of 28 samples). C_o was based on analyses of creek samples taken immediately prior to the initiation of each rainfall event, and C_n was determined from precipitation samples collected in a funnel-type bulk precipitation sampler (Likens et al., 1977) that was deployed in an open field near site RC2.

3. Results

3.1 Hydrogeological Characterization

The Reedy Creek watershed was found to be underlain by dense, dark, olive-gray, clay-silt deposits known as the Calvert Formation, which is comprised of middle Miocene strata of the Lower Chesapeake Group. This group of strata was described by Weems (1986) as a medium-gray, sulfurous, sparsely diatomaceous clay that weathers to pale-gray blocky chips. The layer has low permeability and apparently functions as an aquiclude in the area. The Calvert Formation contrasts with the overlying Upper Chesapeake Group, which is comprised of fine- to medium-grained sands and clayey sands; the Upper Chesapeake cycle is described as an excellent aquifer where clay content is low. At higher elevations in the watershed, Pliocene terrace sand and gravel deposits lie above the Upper Chesapeake deposits. These materials typically cap watershed divides in the area at elevations above 50 m (Weems, 1986). Within the floodplains, deposits are mostly alluvial sands and gravels of Holocene age. These descriptions from published sources are generally consistent with those of the sediments logged in the field along transects T1 and T2.

The presence of extensive wetland reaches within the Reedy Creek basin may be explained by the creek bed's position at or near the contact between the Upper and Lower Chesapeake Groups (Rader, in preparation). This conjecture was supported by field verification of the presence of the dense, hard, impermeable Calvert Formation approximately 1-2 m below the surface of the wetlands and 2-6 m below the sandier hillslopes. The riparian wetlands within the watershed generally were found to be comprised of a layer of medium- to coarse-grained alluvial sands and gravels (1-2 m thick), underlain by the Calvert Formation; the top 10-20 cm of sand appeared high in organic matter, while the lower sand and gravels ordinarily were found to be extremely clean.

3.2 Surface Water Gaging and Annual Water Balance

Because of the importance of the discharge record obtained from the Reedy Creek gage, annual water balances were computed for water years 1990 and 1991 (October 1989 - September 1991), which overlapped the project period. Rating curves for the two periods of record indicate a significant shift in the channel control, and discharge measurements during early 1991 (shown as outliers) indicate a significant backwater effect due to the construction of a beaver dam (Figure 4). The annual hydrograph for Reedy Creek for water year 1990 indicated a mean daily flow range of 0-70 cfs and a mean annual discharge of 15.0 cfs (Figure 5); corresponding values for water year 1991 were 0-60 cfs and 6.7 cfs, respectively (Figure 6). The 1991 water year ended in a major drought, which persisted throughout the summer of 1991 in central and eastern Virginia. Using annual precipitation data for Ashland (101.3 and 90.8 cm), total annual evapotranspiration was computed as

71.6 and 77.5 cm for water years 1990 and 1991, respectively. These values are within about 10% of the long-term annual evapotranspiration computed for the Mattaponi River basin (80.2 cm) based on long-term precipitation (Richmond) and streamflow records. The similarity of estimated Reedy Creek annual evapotranspiration to the long-term mean annual value suggests that the Reedy Creek gage was calibrated with reasonable accuracy and precision.

3.3 Baseflow Variability Studies

Baseflow variability within the Reedy Creek watershed was determined using data from 10 synoptic discharge experiments. All baseflow discharge values were computed on a per-drainage-area basis in units of mm/day (Table 1). These data then were grouped into comparable subsets based on date of measurement and normalized by the mean discharge measured at all sites for each date (Tables 2a-2c). The data showed little consistent variability among sites. Analysis of variance was used to test the null hypothesis that there were no significant differences in normalized mean discharges ($p=0.05$). No significant differences were found for the sites in Tables 2(a) and 2(c): F-values were 1.12 and 0.36 (critical levels were 3.06 and 3.35), respectively. However, the null hypothesis was rejected for the sites in Table 2(b) at the $p=0.05$ level ($F = 3.14$, compared to a critical level of 3.10). Subsequently, a one-tailed t-test revealed that the normalized mean discharge at site TR1 (a small tributary draining a 40-ha watershed and small pond) was significantly less than the normalized mean discharge at site RC3 and site RC2 ($p=0.05$); no other differences among the sites were found to be significant at the $p=0.05$ level, however.

An attempt also was made to quantify the lateral groundwater inflow into Reedy Creek from an ungaged area (114 ha) between upstream site RC3 on the main creek, upstream site TR2 on a major tributary, and downstream site RC2 (Figure 1). The lateral groundwater inflow (Q_{lat}) was computed using the following water balance equation (neglecting the time of travel of water within the drainage network and evaporative losses):

$$Q_{lat} = Q_{RC2} - Q_{RC3} - Q_{TR2} + Q_{TR1} \qquad [10]$$

where Q represents the measured discharge and the subscripts RC2, RC3, TR2, and TR1 represent the corresponding sites, respectively. Computed values for Q_{lat}, expressed both in units of cfs and mm/day, were compared with estimates of Q_{lat} made on the basis of measured flows at site RC2 for each of the 10 sampling dates, assuming discharges were proportional to drainage areas (i.e., equation 2). The results indicated little agreement between the 2 estimates of Q_{lat}, although the mean estimates for the 10 dates using the 2 methods were of the same order of magnitude and very close to zero (Table 3). The high degree of variability in the estimates suggests that the precision and accuracy of stream gaging is insufficient to estimate lateral groundwater inflows from this small ungaged area, which represents only 2.5% of the total watershed area. Duplicate precision in surface-water gaging using the best technique generally is considered to be in the 5-15% range (Winter, 1981). Given the similar size of this ungaged hillslope to the hillslope containing the well transects, it was concluded that the estimate of Q_{lat} using equation 2 is more reliable than the estimate based on synoptic discharge measurements.

3.4 Hillslope/Wetland Experiments

Results from piezometric monitoring of the water-table aquifer within the small sideslope area just upstream from site RC3 (Figure 2) indicated an expected seasonal pattern of groundwater recharge during the fall and winter, followed by groundwater discharge during the spring and summer. Cross-sectional profiles of the hillslope sediments and the water table along transects T1 and T2 at four times of the year are shown in Figures 7-10. Along transect T1, the cross-sectional profiles confirm that the water table was drawn down substantially in late summer of 1990 due to evapotranspirative water demand (Figure 7), but was partially recharged by late fall (Figure 8); the saturated zone at this time was still confined to the wetland and lower hillslope. By early spring, saturated conditions were found along the entire transect (Figure 9), although the water table was drawn down substantially again by late spring (Figure 10). Few differences were observed between transects T1 and T2, although a saturated zone was never observed in the midslope position on transect T2 (well 2SM). These data clearly indicate that measured hydraulic gradients were always in the direction of the stream under baseflow conditions through June 1991, suggesting that shallow groundwater was discharging from the hillslope through the riparian wetlands at a rate sufficient to maintain baseflow in the creek.

During the extreme drought conditions that persisted into the late summer of 1991, however, reversed head gradients between the wetland and creek were observed (Figure 11). Under the conditions measured on September 17, 1991, it is important to point out that there was no measurable creek discharge from the watershed at site RC2. Therefore, creek water levels actually were being drawn down to recharge groundwater levels near the stream (Figure 11).

Because of its conservative nature, naturally occurring chloride was used as a solute tracer to confirm that shallow upland groundwater, wetland groundwater, and stream water were linked hydrologically. It was hypothesized that large differences in concentrations would suggest that surface-water baseflow was not largely originated from shallow groundwater. Measurements of chloride concentrations in wells along both transects showed levels that were comparable in magnitude to measurements made in the creek proper. In addition, a dramatic increase in stream chloride concentration beginning in October 1990 was clearly indicated in well samples collected from transect T1, but to a lesser extent in samples from transect T2 and the creek (Figure 12). The results suggest that these three components are hydrologically linked.

To determine whether the magnitude of shallow groundwater discharge is of the same magnitude as stream baseflow, the saturated hydraulic conductivity (K_s) of the phreatic aquifer was estimated using three independent techniques. Each of the methods required data collected at different spatial scales. The use of downstream discharge and Darcy's Law used estimates of lateral groundwater flow based on *watershed-scale* discharge measurements and measured hydraulic gradients at the *hillslope scale*. The tracer experiment was performed at the *hillslope scale;* the slug tests represented data collected at the *point scale*.

Results using downstream discharge (Q_{RC2}) and Darcy's Law for 18 dates on which the hydraulic gradients and stream discharge were measured are shown in Table 4. A mean value for K_s of 0.019 cm/sec ($\pm$ 0.004 cm/sec; p=0.05) was computed, a value that is quite consistent with the sandy materials overlying the Calvert aquiclude. Freeze and Cherry (1979) indicate that a value of about 0.02 cm/sec falls in about the middle of the range given for clean sand materials. While the mean value may provide an adequate estimate of K_s, a better estimate was obtained by regressing the computed K_s value for the 18 dates on Q_{RC2}; a statistically significant linear relationship ($r^2 = 0.62$; $p < 0.001$) with positive slope and intercept was found between the two variables (Figure 13). It was presumed that the y-intercept of the graph in Figure 13 (0.009 cm/sec) is a better estimate of K_s because it is equivalent to the value one expects as the stream discharge approaches zero. Under these conditions, streamflow presumably would be supplied only by direct lateral groundwater discharge to the creek, which is the physical situation that this technique assumes. Under higher flow conditions, however, a significant proportion of baseflow may result from water draining through small wetland channels or from incised subwatersheds, particularly those characterized as topographically convergent. Discharge from these sources would be expected to increase the estimate of K_s using Darcy's Law, an explanation that is clearly consistent with the results shown in Figure 13.

The tracer experiment performed at the base of the hillslope provided an estimate for K_s that was virtually identical to the value estimated on the basis of downstream discharge and Darcy's Law. The value computed was 0.010 cm/sec, based on the arrival of the tracer peak at well TW4 in 7 days (Figure 14), a travel distance of 11.4 m, an estimated porosity of 0.4, and an average measured hydraulic gradient of 0.0736 (actual range was 0.073-0.078).

K_s also was computed based on results from a series of bailtests (rising head slug tests) using the Hvorslev method (Table 5). Two typical bailtest results are shown in Figure 15. Assuming a lognormal distribution, a mean value for K_s of 0.0033 cm/sec (total range 0.00036-0.0132) was computed (mean logK = -2.49 cm/sec). While the mean value was about a factor of three less than the estimate of K_s using the other two methods, the highest bailtest estimates were within less than a factor of two different. A two-tailed t-test indicated no significant difference between the mean K_s computed for the hillslope wells (0.0019) and the mean for the wetland wells (0.0045) at a level of $p=0.05$, again assuming a lognormal distribution.

Results from continuous monitoring of the water level in wetland well 2WX relative to the surface elevation during February and March 1991 indicated the occurrence of eight recharge events due to rainfall (Figure 16). During four of the events in March, the water table in the wetland rose to levels above the ground surface, indicating transient conditions of full saturation. During each of these four events, the corresponding stream hydrograph from site RC2 indicated a significant stormflow response. However, during the other four rainfall events in February, wetland surface saturation did not occur and the stream indicated little (if any) stormflow response. These results support the hypothesis that the stormflow response of the Reedy Creek watershed is closely tied to the recharge status of the riparian wetlands.

3.5 Chemical Separations of Stormflow Hydrographs

Chemical hydrograph separations based on naturally occurring chloride ion were performed as a means of quantifying the absolute and relative contributions of new and old water to storm hydrographs. Chloride concentrations were measured in rainfall and streamwater before and during eight separate events: a major rainstorm in August 1990, two smaller storms in October 1990, three storms in March-April 1991, one storm in July-August 1991, and one storm in September 1991.

Chloride concentrations were always much lower in precipitation than in antecedent streamflow, which is consistent with the observation that rainstorms caused substantial chloride dilution in the creek. During the extreme event monitored in August 1990, in which peak hourly streamflow reached nearly 50 cfs, chloride concentrations were diluted nearly 50% (from 120 μeq/L to 70 μeq/L) at the event peak (Figure 17). Similar dilution patterns were observed during most of the other events, but the magnitudes were generally somewhat less than during the August event (Figures 18-21).

In most cases, the chloride concentration at the end of the recession returned to a value close to the antecedent concentration. This was not the case, however, during the first October 1990 event, in which the chloride concentration rose dramatically near the end of the recession to a value much higher than the antecedent value. A similar phenomenon occurred during the second October 1990 storm, suggesting that the assumption of constant old-water chloride during these two events appeared to be violated. The most likely explanation for this behavior is that chloride stored in the vadose zone may have been mobilized during the groundwater recession. For this reason, the October events were not subjected to chemical hydrograph separation.

Results from the hydrograph separations for the other events are shown in Figures 22-26. (Two sequential storms in early March 1991 that occurred close together were considered one storm for the purpose of this analysis.) For all events, maximum new-water discharge usually occurred near the hydrograph peak and ranged from about 20-40% for the six events with a mean of 31% (Table 6). The percentage contributions of new water determined by integration of the new-water hydrograph were expectedly much less, ranging from a minimum of 7% for the September 1991 storm to a maximum of nearly 25% for the August 1990 event. The mean integrated new-water contribution for the six storms was computed as 15% (Table 6).

4. Discussion

Results from a series of field hydrogeological experiments performed at various spatial scales within the Reedy Creek watershed confirmed that stream baseflow in this system is supplied exclusively by drainage of shallow (< 6 m deep) groundwater from a thin surficial water-table aquifer. Under low-baseflow conditions observed during summer and early autumn, groundwater discharge is supplied only from the highly permeable alluvial sand deposits that comprise the riparian wetlands surrounding Reedy Creek (and perhaps from springs that feed small tributaries draining upland subwatersheds). Under higher-baseflow conditions in late winter and spring, however, groundwater discharge from the base of moderately steep hillslopes within the watershed appears to be important in maintaining partial saturation of the riparian wetlands. In addition, under these higher-flow conditions, deeply incised, topographically convergent sideslope drainages are more likely to contribute to stream baseflow, as the channel network effectively taps into the shallow aquifer. Interestingly, based on our description of the watershed's hydrogeological setting (particularly the presence of the Calvert aquiclude), it is evident that there is probably no significant interaction of deep groundwater with surface waters at Reedy Creek. This supports the overall premise that the low alkalinity of Reedy Creek can at least in part be attributed to a rather limited interaction of water with readily weatherable surficial materials.

The stormflow response of the Reedy Creek watershed is also dependent upon the conditions of the shallow aquifer, particularly in the riparian wetland. The fact that significant stormflow responses occur only when the wetland is fully saturated supports the hypothesis that a saturation overland flow mechanism is the dominant stormflow-generating process. As a result of the high permeability of soils within the watershed, overland flow does not appear to occur as a result of infiltration-excess, but may result from soil profile saturation. To further examine this hypothesis, the new-water contributing area (NWCA) associated with each storm was estimated by dividing the new-water volume for a particular storm (determined by integrating the new-water hydrograph) by the precipitation depth for that storm. These computed areas then were compared with estimates of perennial channel area (3.8 ha), intermittent channel area (1.1 ha), pond area (14.8 ha), and wetland area (191.8 ha) based on a digitized topographic map and field estimates of channel width for streams of various order (Table 6). It is important to note that this method assumes a static NWCA during each event (as opposed to a variable source area) and neglects evaporative losses from interception and depression storage.

For four of the six storms that occurred prior to the summer of 1991 in which rainfall depths ranged from 1.76-3.90 in, the computed NWCAs were much greater than the sum of total channel and pond area (20 ha), but much less than the total channel, pond, and wetland area (210 ha).

This comparison supports the hypothesis that the stormflow response of the Reedy Creek watershed ordinarily cannot be explained based on expansion of the channel network alone, but also must consider the areas of nearly saturated to saturated land that lie within the riparian zone. During moderate rainstorms with wet antecedent conditions (the second March 1991 storm), nearly two-thirds of the mapped riparian wetlands may have contributed direct precipitation to streamflow via a saturation overland flow route. It is interesting to note that the rainfall depth associated with the first March 1991 storm was of the same magnitude as the second, but the NWCA was only one-third as large, due to differences in antecedent moisture conditions. Similarly, the computed NWCA for the large August 1990 storm, with a rainfall depth nearly twice as large, was about 50% less than the second March 1991 event (Table 6).

Interestingly, under the extreme drought conditions that existed in the watershed and throughout the region prior to the July and September 1991 storms (there was water in the channels but no measurable flow at the gage for several weeks prior to September 19), our technique indicated computed NWCAs of 4 and 3 ha, respectively, virtually the same value as the estimated perennial stream area of 3.7 ha (Table 6). During extremely dry conditions, it is reasonable to suspect that only the perennial channels themselves would be in a state of complete saturation.

Results from the chemical stormflow separation suggest that stormflow response in coastal plain watersheds like Reedy Creek can be explained Our results clearly demonstrate that significant rainstorms cause overland flow of direct precipitation from saturated areas. The extent of saturated area can vary seasonally and during the course of an event, although our study did not examine the latter phenomenon. In most cases, the NWCA is a large fraction of the total mapped wetland area (i.e., the floodplain). However, under extremely dry antecedent conditions, the perennial stream area provides a better estimate. Under wetter conditions antecedent to large rainstorm events, we would expect the NWCA to approach or even exceed the riparian wetland area. It is conceivable that topographically convergent headwater watersheds and sideslopes overlying the shallow Calvert aquiclude also may rapidly saturate and contribute new water under such conditions.

Saturation overland flow also occurs by the more dominant return-flow phenomenon, whereby groundwater (and perhaps soil water) is displaced as the water tables rise as a result of (1) direct infiltration and (2) recharge of saturated zones from downslope groundwater flow. Stormflow at the Reedy Creek watershed is comprised predominantly of old water that was in storage in the watershed before a rainstorm event. Even for the extreme August 1990 event, stormflow was more than 75% old water (Table 6).

There are a number of physical mechanisms that have been proposed to explain the rapid groundwater response of systems such as Reedy Creek. In some systems, a groundwater ridging mechanism has been proposed in which a disproportionately large rise in the water table is caused by the conversion of a tension-saturated capillary fringe into phreatic water by infiltrating rain (Sklash and Farvolden, 1979; Gillham, 1984; Sklash et al., 1986). In the Reedy Creek watershed, however, the near-saturated areas are primarily comprised of alluvial sands and gravels in which the capillary fringe presumably would be very thin or insignificant. For this reason, we believe that rapid drainage of groundwater from upslope areas and vertical infiltration in riparian wetland soils may be sufficient to explain the return-flow response of Reedy Creek. This drainage may be augmented by flow from swales and small tributaries, where water has been observed to disappear beneath the ground where the channel reaches the alluvial floodplain deposits. Undoubtedly, an important contributing factor is the shallow depth of the water table within the riparian zone. Another factor may be the riparian microtopography, which may allow for rapid response of low-lying areas.

This explanation also seems to explain the results in Figure 16, whereby only five storms caused measurable stormflow response, while four storms caused no significant stormflow. All of the storms, however, produced a response of the riparian water table. Analysis of the hyetographs for the events indicate that the nonstormflow-producing events were characterized by low peak rainfall intensities (< 0.20 in/day). In contrast, the March events were more intense, with peak rainfall intensities ranging from 0.25-2.34 in/day (Figure 16). From these results, it is reasonable to hypothesize that high-intensity events are more efficient at producing stormflow than are low-intensity events, because they permit saturation to occur over greater areas of the riparian zone. During high-intensity events, saturation conditions occur because vertical-flow velocities are much higher than horizontal-flow velocities, the later causing subsidence of the water-table response. Conversely, during low-intensity events, horizontal velocities allow rapid equilibration and limit the extent of surficial saturation.

We are not aware of any other studies conducted in the coastal plain or elsewhere that have quantified NWCA's using natural chemical tracers and related them to mappable basin characteristics. Several studies have used the quickflow separation technique of Hewlett and Hibbert (1967), which arbitrarily assumes a linear rise in delayed flow beginning at the time of rise. Taylor (1982) used this technique and found large values for the ratio of total quickflow to precipitation (0.6-78%) for a small glaciated watershed in Ontario. Equivalent values for the ratio of the new-water contributing area to the total watershed area (WA) at Reedy Creek ranged from less than 0.1% to about 3% for the four storms described in Table 6. Sklash et al. (1986) used naturally occurring deuterium and the quickflow technique of Hewlett and Hibbert (1967) to separate hydrographs from two humid headwater watersheds in the Maimai experimental area in New Zealand. Their results indicated that the quickflow technique overestimated the new-water contribution by as much as a factor of seven. The use of natural tracers is, therefore, a major step in the direction of assessing the streamflow generation potential of watersheds and of determining the extent of interaction of groundwater and surface waters.

The finding that computed contributing areas for saturation overland flow, based on chemical hydrograph separations of stormflow hydrographs using chloride, were close to expected saturated areas, based on digitized topographic maps for a range of moisture conditions, is significant for several reasons. First, a method for elucidating contributing areas at the watershed scale previously required substantial instrumentation (piezometers) to detect conditions of surface saturation, whereas the method used in this study requires only chemical and discharge data from one point within a drainage basin. Second, the results tend to support the use of a two-component model of stormflow generation for vegetated watersheds in humid, temperate climatic regions. This model and its explicit assumptions, proposed more than 15 years ago by Sklash et al. (1976), have come under considerable scientific scrutiny during the last few years (e.g., Kennedy et al., 1986; DeWalle et al., 1988; Genereux and Hemond, 1990; McDonnell et al., 1990; McDonnell et al., in press). Third, the results are also significant because they should permit the subsequent independent parameterization and validation of both lumped and distributed physical models of stormflow generation based on the saturation overland flow conceptual model (e.g., TOPMODEL: Beven and Kirkby, 1979). It is our opinion that this approach may prove successful at Reedy Creek, although several recent papers have demonstrated difficulties associated with parameterization of hydrological models using chemical data (Hooper et al., 1988; Beck et al., 1990).

5. Conclusions and Research Needs

A field study of the surficial hydrology of a representative watershed in the Virginia Coastal Plain demonstrated that surface waters within the Reedy Creek watershed are derived primarily from drainage of a shallow water-table aquifer and from direct precipitation onto near-channel saturated areas. The saturated hydraulic conductivity of lower hillslope soils and alluvial deposits (estimated using two independent methods) is sufficiently high to explain stream discharge under baseflow conditions. That these methods all provided similar results (within about a factor of three) is encouraging, but not unexpected given the relatively homogeneous surficial sediments within the system. The presence of the Calvert aquiclude at shallow depth results in a water-table aquifer in which groundwater residence times would be expected to be quite short.

Given this situation, it is expected that surface applications of some agricultural chemicals (particularly those that are not easily biodegraded) could rapidly contaminate surface waters within this watershed; however, the probability of extensive contamination of deeper groundwater resources in the basin would be expectedly much lower owing to the presence of the aquiclude. Applications of agricultural fertilizers may behave differently, however. Because the dominant flowpath of water is through the riparian wetlands, there is an opportunity for nutrient removal, both by plant and microbial uptake. In the case of nitrogen and sulfur, known dissimilatory reduction mechanisms (denitrification and sulfate reduction by obligate anaerobes) may substantially reduce the loadings of such pollutants to surface waters. Because anaerobic conditions would be expected to occur within the water-saturated near-stream zone, riparian wetlands may play an important role with respect to both water routing and water quality. As with the less biodegradable chemicals, pollutants resulting from fertilizer applications (e.g., nitrate) would be expected to be transported predominantly horizontally to riparian zones, rather than vertically to deeper groundwater environments.

The stormflow response of the Reedy Creek watershed appears to occur primarily by a saturation overland flow mechanism in which the riparian wetlands play a dominant role, while the sideslopes and headwater watersheds are largely responsible for sustaining and recharging water levels in the wetland zone. However, on the basis of this study, we are unable to determine the extent to which steep, incised subwatersheds characterized by convergent topography may augment the saturation overland flow response of the riparian zone. Field observations suggest that some small convergent sideslopes tributary to Reedy Creek are drained by deeply incised channels, but that these channels frequently disappear into alluvial fan deposits at the base of the hillslopes. One important implication of this flowpath is that water flowing in upland channels may end up recharging groundwater within the wetland area, implying that some surface water actually can further interact with wetland groundwater through a type of bank storage phenomenon.

Further research is needed to clearly establish flowpaths within these convergent sideslope areas. A useful approach would be a detailed mapping of areas of complete soil saturation during rainstorm events using a series of simple crest piezometers or a more complex set of pressure transducers and data loggers. The relative homogeneity of the water-table aquifer, the presence of the shallow aquiclude, and the relatively accurate measurement of annual water balance make the Reedy Creek watershed an excellent site for conducting these types of experiments. A second logical field experiment would be a hillslope-scale tracer experiment in which the release of a material injected into the saturated zone could be tracked over time. Such an experiment, while costly, might allow the relative contributions of old and new water to be quantified further.

It is important to realize that the dominance by old water of the stormflow response of the watershed is also significant from a water quality perspective. That this water is riparian in its recent origin implies that contact of the water with microbially active zones is likely significant. Therefore, we would expect that the loading of many pollutants to the stream is probably reduced as a result of microbial and plant transformations within the wetland.

We also must conclude that our success in relating the computed new-water contributing areas to mapped topographic features (floodplains, channels, etc.) is an important step toward a quantitative understanding of flowpaths in coastal plain watersheds. The next logical research step should be application of the results toward independent parameterization and validation of a physical model of stormflow generation based on the saturation overland flow mechanism (e.g., TOPMODEL). Assuming success in this area, it would be reasonable to proceed in the direction of incorporating some simple pollutant transport and transformation modules within the framework of such a physically based simulation model. While parameterization would likely be even more difficult, such a model would be useful in predicting the response of coastal plain watersheds to a variety of human perturbations.

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Tables

Table 1.
Synoptic discharge measurements (mm/day) for six locations within the Reedy Creek watershed

Date	Measurement Location					
	RC6	TR3	RC3	TR2	RC2	TR1
06/08/89	1.47	0.66	0.40	0.54	0.43	–
06/28/89	0.25	0.13	0.22	0.22	0.18	–
07/21/89	0.41	0.39	0.59	0.44	0.46	–
08/21/89	0.28	0.26	0.44	0.28	0.41	–
03/28/90	–	–	0.94	0.70	0.94	0.43
04/26/90	–	–	0.83	1.93	0.94	1.22
05/16/90	–	–	0.64	0.48	0.62	0.37
06/05/90	–	–	1.50	0.92	1.54	0.49
07/25/90	–	–	0.22	0.12	0.15	0.06
09/11/90	–	–	0.20	0.82	0.22	0.00

Table 2.
Synoptic discharge measurements grouped by date and location and normalized by the mean value for each corresponding date (dimensionless units)

(a)

Date	Measurement Location					Mean by date
	RC6	TR3	RC3	TR2	RC2	
06/08/89	2.10	0.94	0.57	0.77	0.61	1.00
06/28/89	1.24	0.65	1.11	1.09	0.91	1.00
07/21/89	0.89	0.86	1.29	0.96	1.00	1.00
08/21/89	0.84	0.79	1.32	0.83	1.22	1.00
Mean	1.27	0.81	1.07	0.91	0.94	1.00

Table 2. Continued.

(b)

Date	Measurement Location				Mean by date
	RC3	TR2	RC2	TR1	
03/28/90	1.25	0.93	1.25	0.57	1.00
04/26/90	0.68	1.57	0.76	0.99	1.00
05/16/90	1.22	0.91	1.18	0.70	1.00
06/05/90	1.35	0.83	1.38	0.44	1.00
07/25/90	1.63	0.87	1.06	0.44	1.00
09/11/90	0.65	2.63	0.72	0.00	1.00
Mean	1.13	1.29	1.06	0.52	1.00

Table 2. Continued.

(c)

Date	Measurement Location			Mean by date
	RC3	TR2	RC2	
06/08/89	0.88	1.18	0.94	1.00
06/28/89	1.07	1.05	0.88	1.00
07/21/89	1.19	0.89	0.92	1.00
08/21/89	1.17	0.74	1.08	1.00
03/28/90	1.10	0.81	1.09	1.00
04/26/90	0.68	1.56	0.76	1.00
05/16/90	1.11	0.83	1.07	1.00
06/05/90	1.14	0.70	1.17	1.00
07/25/90	1.37	0.73	0.90	1.00
09/11/90	0.49	1.98	0.54	1.00
Mean	1.02	1.05	0.93	1.00

Table 3.
Estimates of lateral groundwater discharge (Q_{lat}) from an ungaged sideslope to Reedy Creek for 10 sampling dates

Date	Q_{lat} (computed) (mm/day)	Q_{lat} (site RC2) (mm/day)	Q_{lat} (computed) (cfs)	Q_{lat} (site RC2) (cfs)
06/08/89	0.00	0.43	0.00	0.20
06/28/89	−1.29	0.18	−0.60	0.09
07/21/89	−3.00	0.46	−1.40	0.21
08/21/89	0.86	0.41	0.40	0.19
03/28/90	3.43	0.94	1.60	0.44
04/26/90	−6.87	0.94	−3.20	0.44
05/16/90	1.50	0.62	0.70	0.29
06/05/90	9.23	1.54	4.30	0.72
07/25/90	−1.72	0.15	−0.80	0.07
09/11/90	−5.58	0.22	−2.60	0.10
Mean	−0.34	0.59	−0.16	0.27

Table 4.
Computation of specific discharge and saturated hydraulic conductivity (K_s) of the phreatic aquifer using Darcy's Law and baseflow discharge data for site RC2 for 18 dates

Date	Baseflow (cms)	Q_{lat} (cms)	Water table height (m)	Reach length (m)	dh/dl	q (cm/s)	K (cm/s)
07/25/90	0.076	0.00020	0.82	241	0.007	0.00010	0.015
08/08/90	0.048	0.00013	0.70	241	0.006	0.00007	0.012
09/11/90	0.117	0.00031	0.78	241	0.011	0.00016	0.014
10/09/90	0.070	0.00018	0.73	241	0.003	0.00011	0.030
10/30/90	0.113	0.00030	0.96	241	0.010	0.00013	0.013
11/27/90	0.162	0.00043	0.97	241	0.010	0.00018	0.018
12/17/90	0.395	0.00104	1.05	241	0.009	0.00041	0.044
01/22/91	0.322	0.00085	1.00	241	0.015	0.00035	0.023
02/05/91	0.217	0.00057	0.98	241	0.010	0.00024	0.024
02/12/91	0.253	0.00067	0.97	241	0.012	0.00029	0.024
03/05/91	0.381	0.00101	1.00	241	0.012	0.00042	0.035
03/26/91	0.137	0.00036	0.95	241	0.013	0.00016	0.012
04/12/91	0.223	0.00059	1.00	241	0.012	0.00024	0.020
05/05/91	0.139	0.00037	0.88	241	0.015	0.00017	0.012
05/23/91	0.173	0.00046	0.81	241	0.015	0.00023	0.016
06/13/91	0.067	0.00018	0.50	241	0.010	0.00015	0.015
07/03/91	0.062	0.00016	0.54	241	0.009	0.00013	0.014
07/25/91	0.022	0.00006	0.40	241	0.007	0.00006	0.009
						MEAN =	0.019

Table 5.
Saturated hydraulic conductivities (K_s) computed from rising head slug tests using the Hvorslev method

Hillslope Wells	$\log K_s$ (cm/sec)	Wetland Wells	$\log K_s$ (cm/sec)
1SH	−2.65	1WS	−1.88
1SM	−3.15	1WN	−2.58
1SL	−2.26	1C	−2.34
2SH	−3.44	2WS	−2.31
2SM	–	2WN	−2.05
2SL	−2.75	2C	−2.41
TW1	−2.18	TW2	−1.94
		TW3	−2.03
		TW4	−2.34
		TW5	−3.00
		TW6	−2.97
MEAN	−2.72	MEAN	−2.35

Table 6.
Chemical hydrograph separation results for eight stormflow events

Event Dates	Total Rainfall (in)	Maximum New Water (%)	Total New Water (%)	New Water Volume (10^3 ft^3)	NWCA[1] (ha)	NWCA/WA[2] (%)
08/08/90–08/22/90*	3.90	38.3	24.4	3020	86	1.9
10/16/90–10/29/90*	2.60	–	–	–	–	–
03/12/91–03/19/91*	1.76	31.6	12.1	670	42	0.9
03/26/91–04/06/91*	2.03	35.7	18.5	2390	131	2.9
07/25/91–07/31/91*	3.13	27.0	11.5	108	4	0.1
09/19/91–09/23/91**	1.90	20.0	7.0	53	3	0.1

*precipitation data from Ashland, VA station
**precipitation measured at Reedy Creek
[1]new water contributing area
[2]total watershed area

Figures

Figure 1.
Map of the Reedy Creek catchment near Ashland, Virginia.

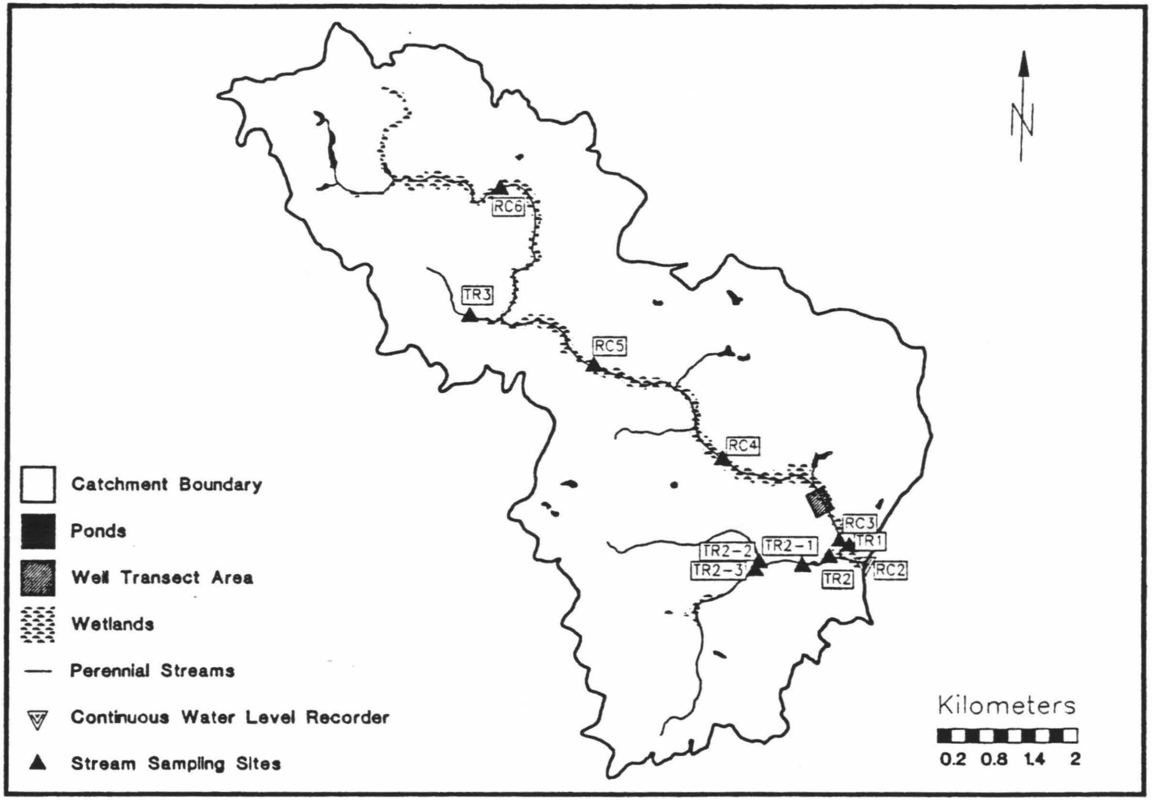


Figure 2.
Map of the well transect area just upstream from site RC3. Elevation contours (m) relative to temporary bench mark (TBM) established along transect T2.

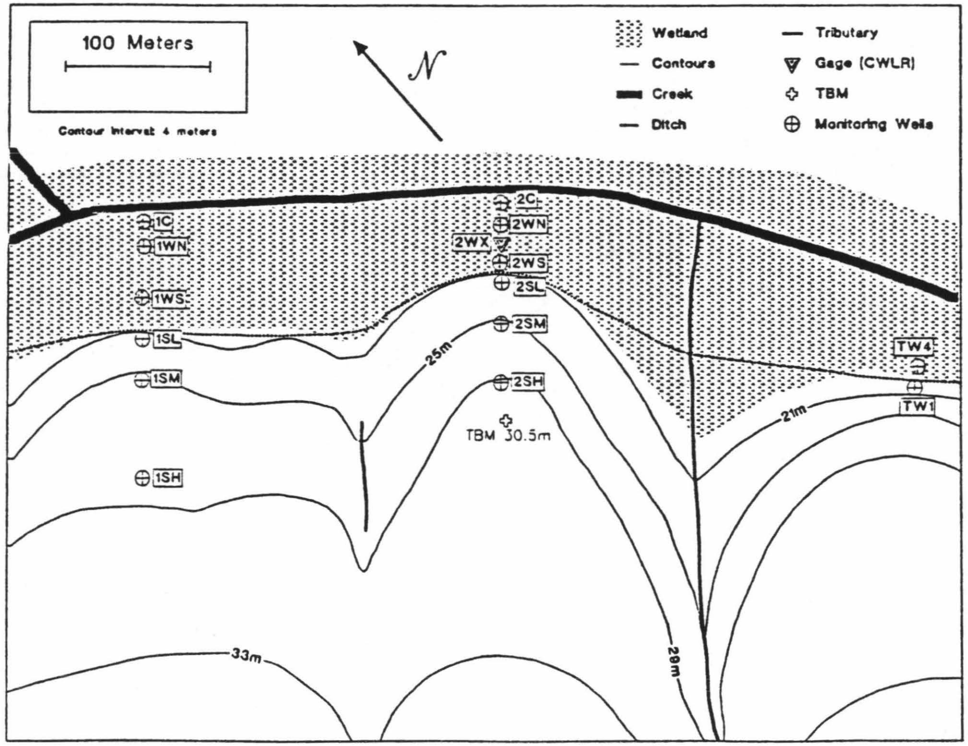


Figure 3.
Spatial configuration of the injection and monitoring wells used in the subsurface tracer experiment.

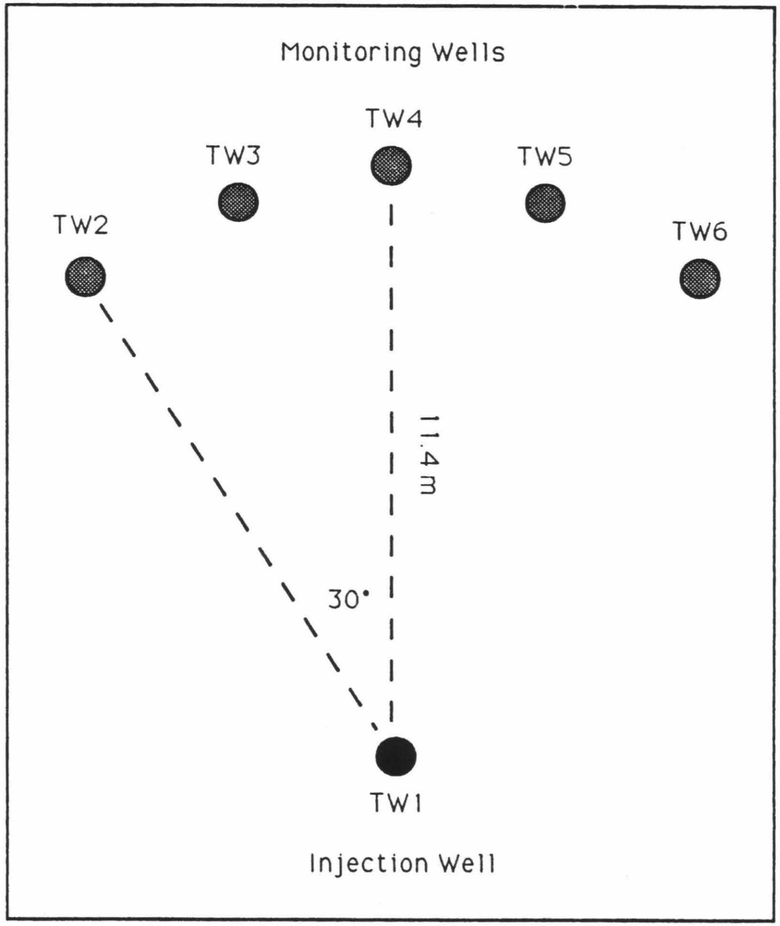


Figure 4.
Stage-discharge relationships for site RC2 developed from stream discharge measurements made during the project.

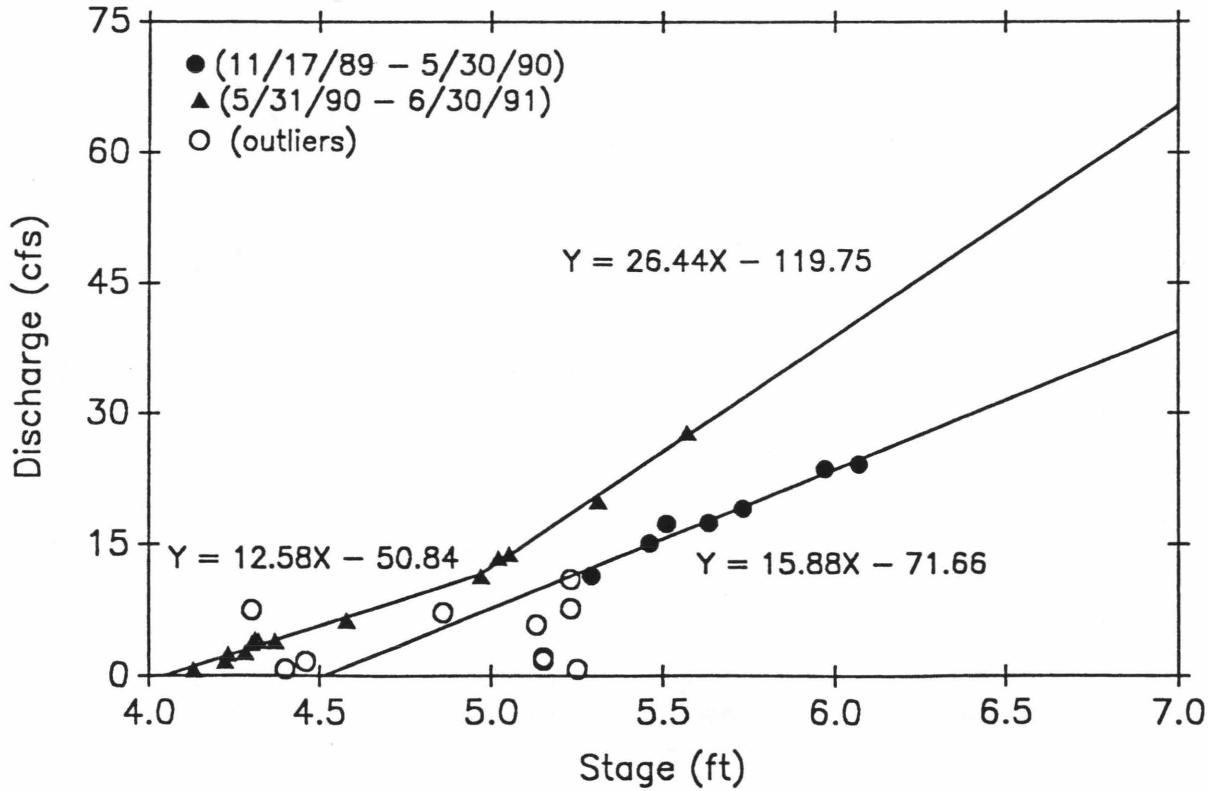


Figure 5.
Annual stream hydrograph for Reedy Creek for the 1990 water year (October 1, 1989 - September 30, 1990).

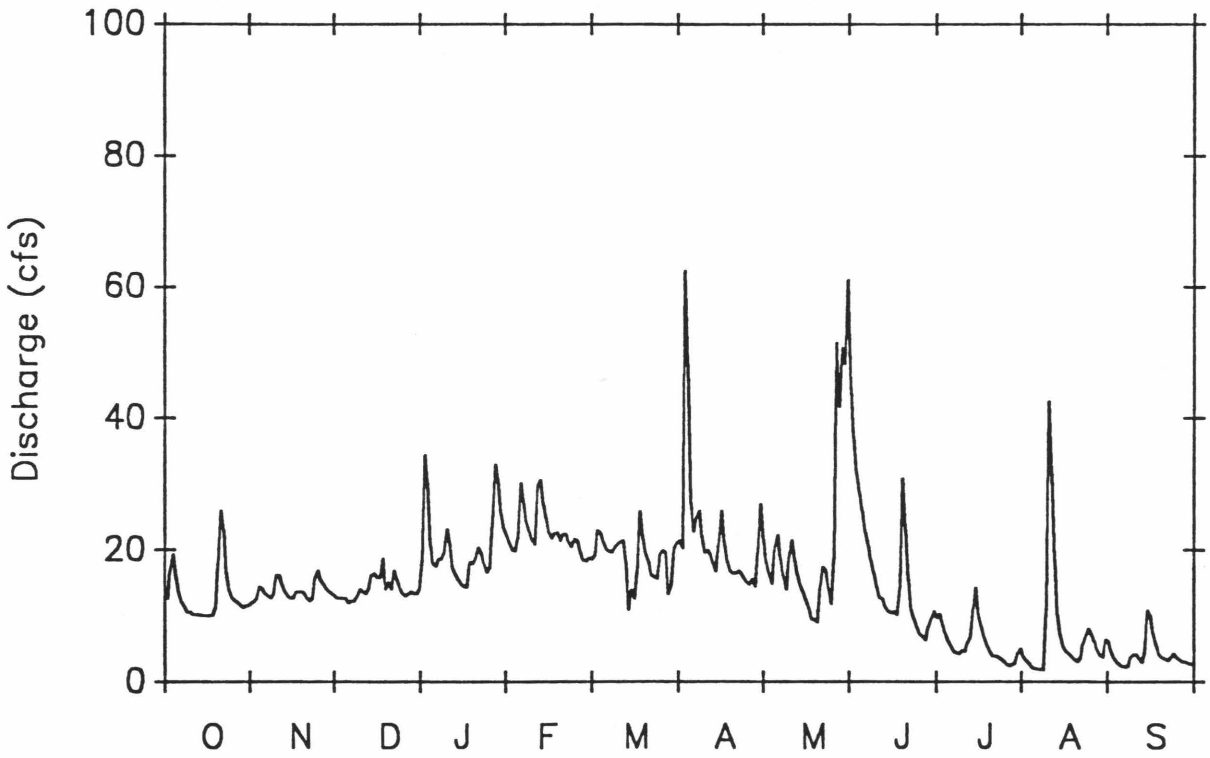


Figure 6.
Annual stream hydrograph for Reedy Creek for the 1991 water year (October 1, 1990 - September 30, 1991).

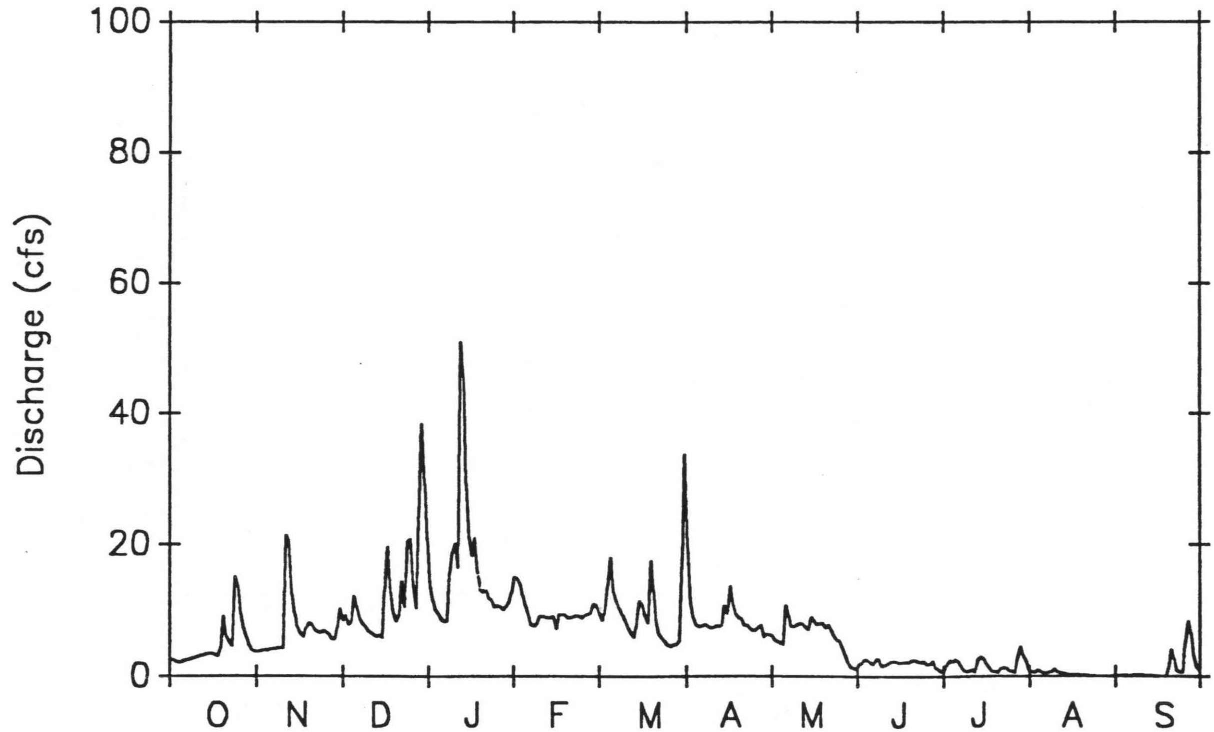
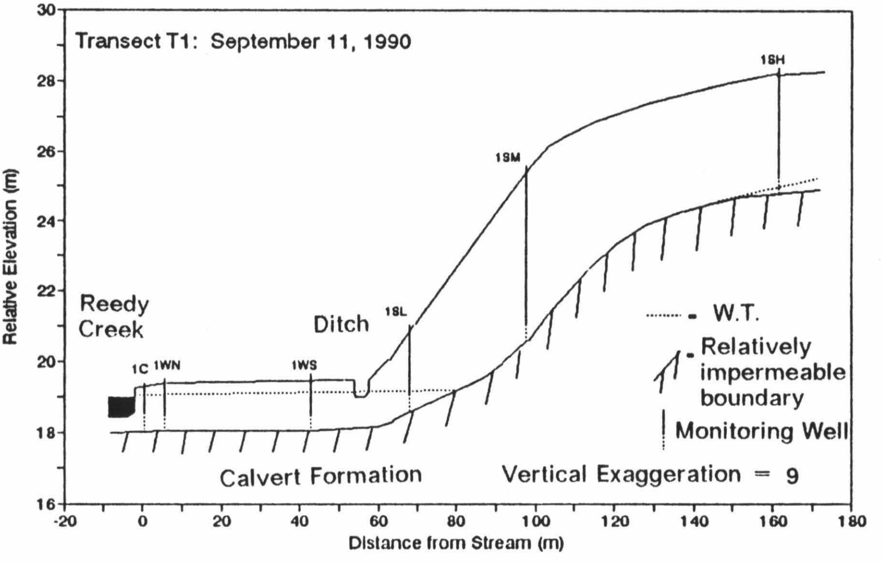


Figure 7.
Cross-sectional profiles showing the location of the water table along transects T1 and T2 in late summer, 1990.



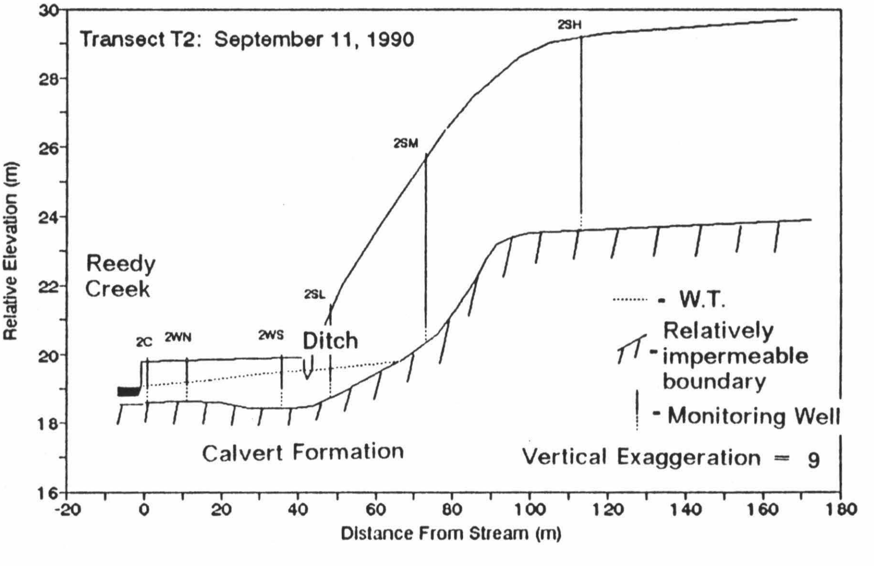
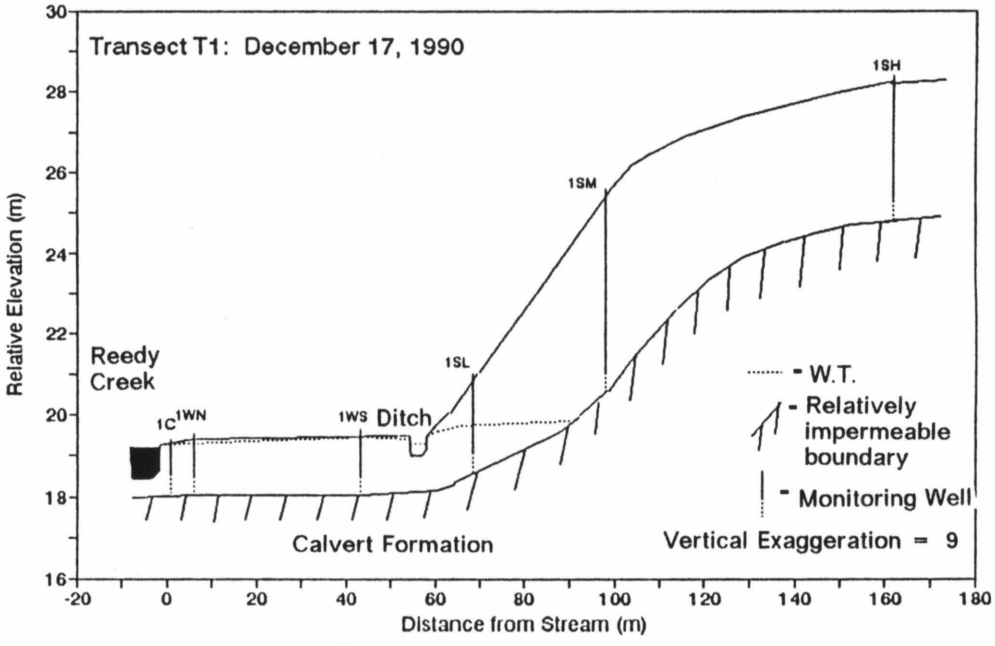


Figure 8.
Cross-sectional profiles showing the location of the water table along transects T1 and T2 in late autumn, 1990.



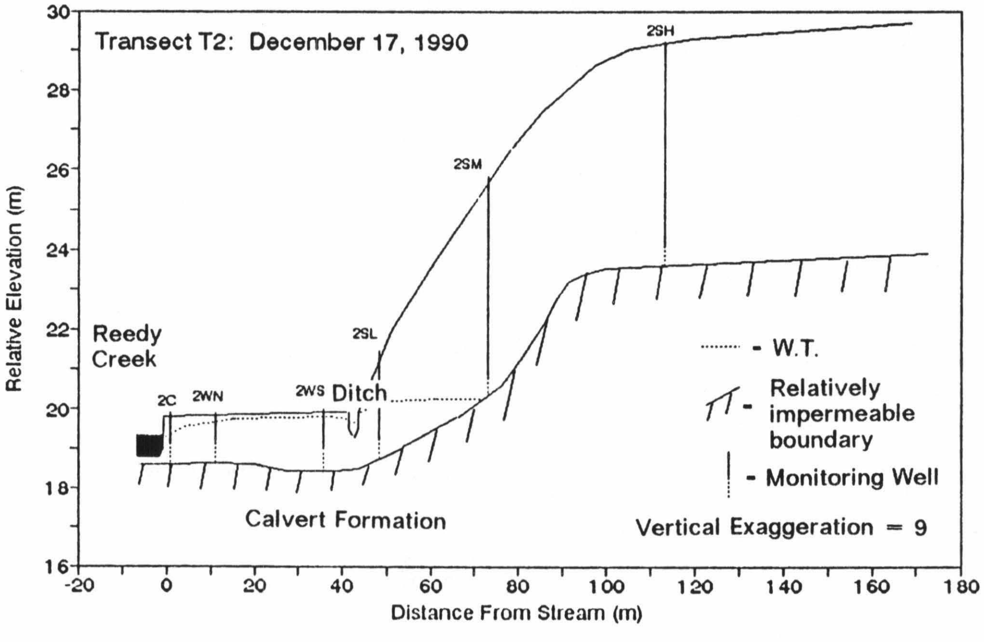


Figure 9.
Cross-sectional profiles showing the location of the water table along transects T1 and T2 in early spring, 1991.

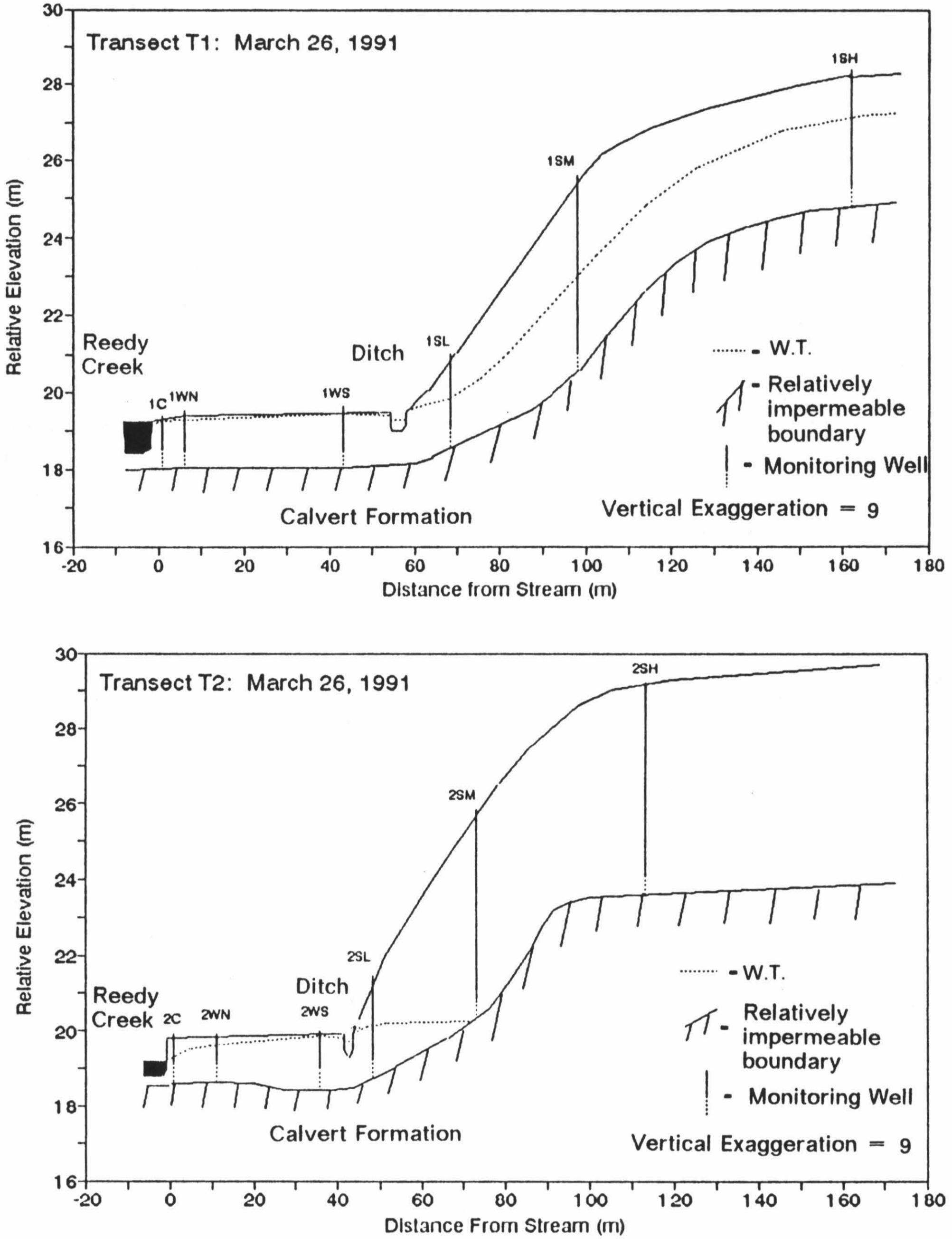
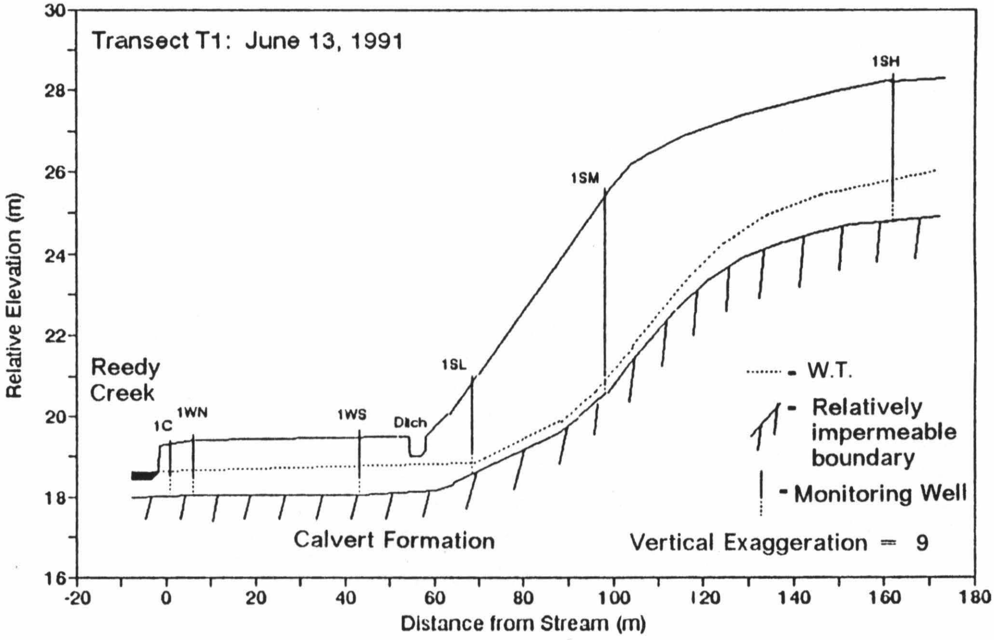


Figure 10.
Cross-sectional profiles showing the location of the water table along transects T1 and T2 in late spring, 1991.



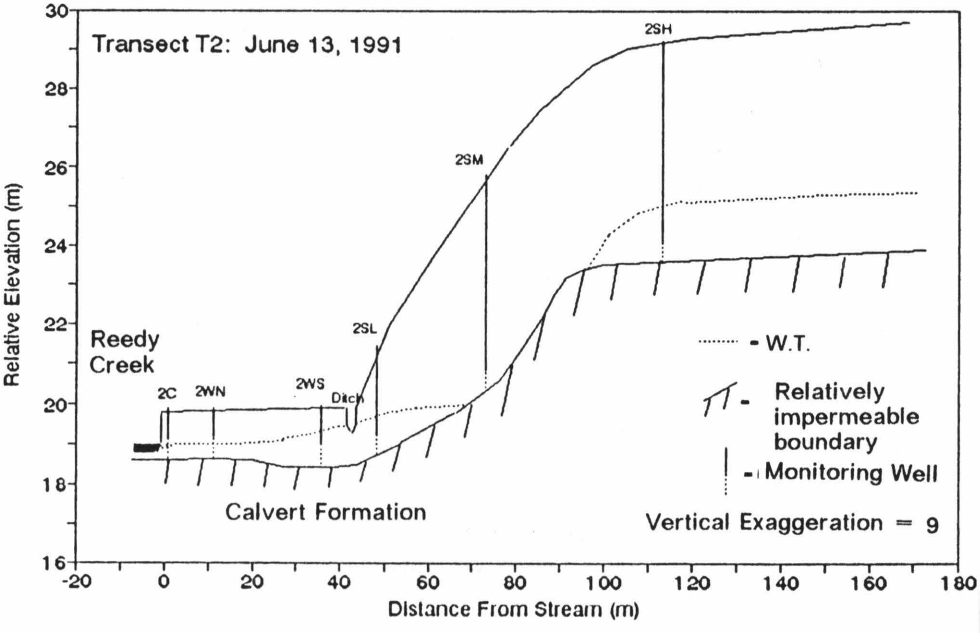
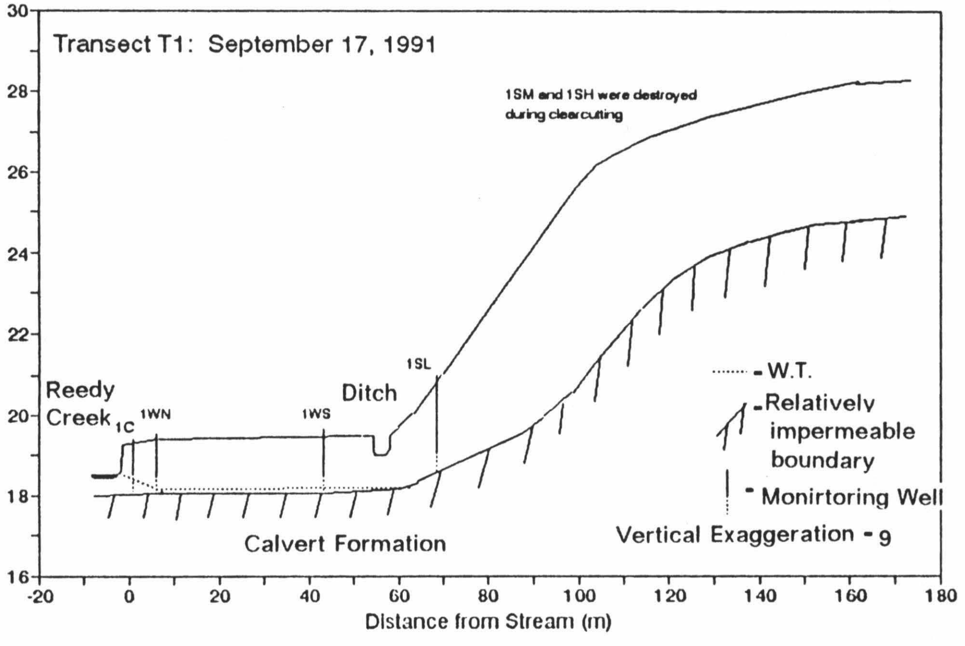


Figure 11.
Cross-sectional profiles showing the location of the water table along transects T1 and T2 in late summer, 1991.



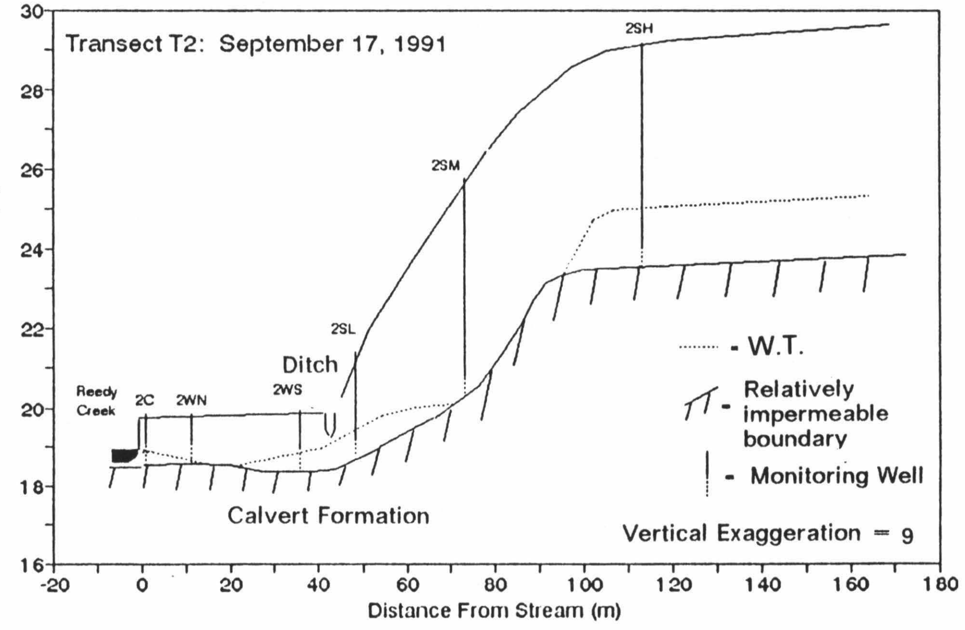


Figure 12.
Temporal patterns in chloride concentrations measured in shallow wells along the two transects and in creek water from July 1990 through July 1991.

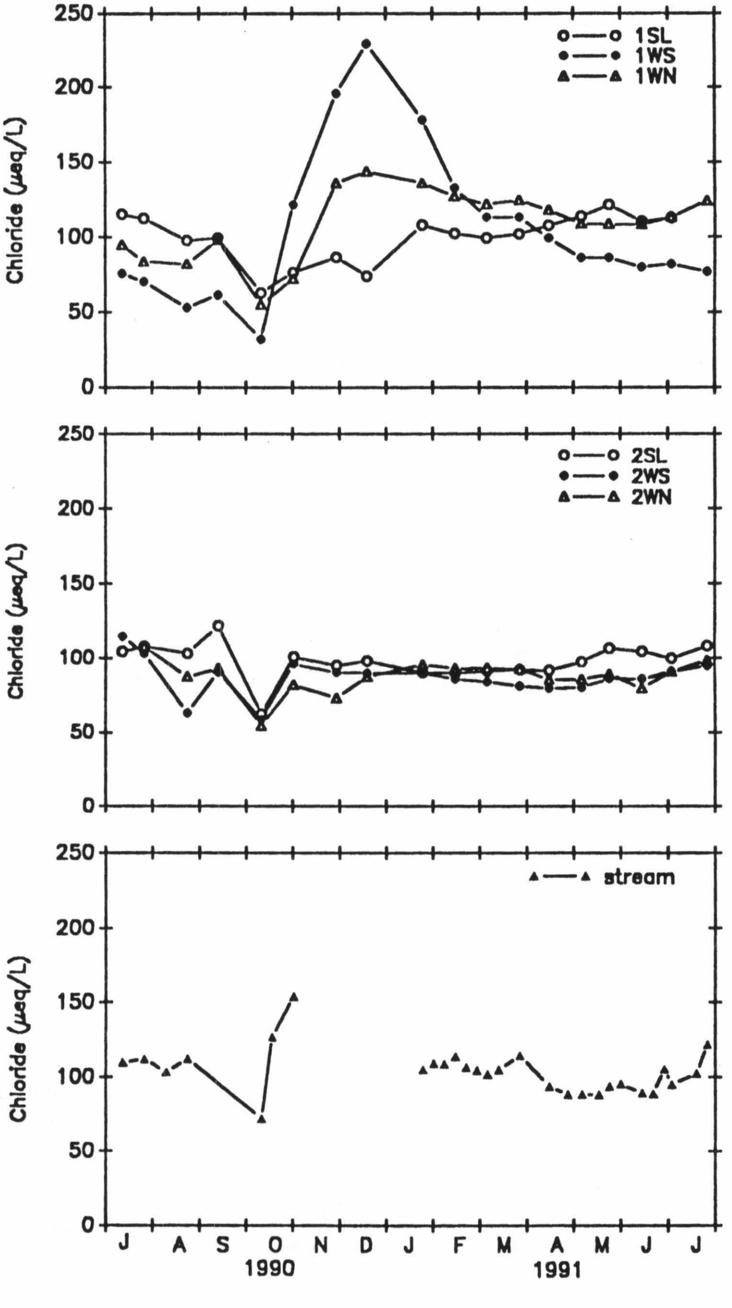


Figure 13.
Graph of computed saturated hydraulic conductivity (K_s) on creek baseflow (cms = m^3/sec).
(See section 2 for discussion of K_s computation using Darcy's Law.)

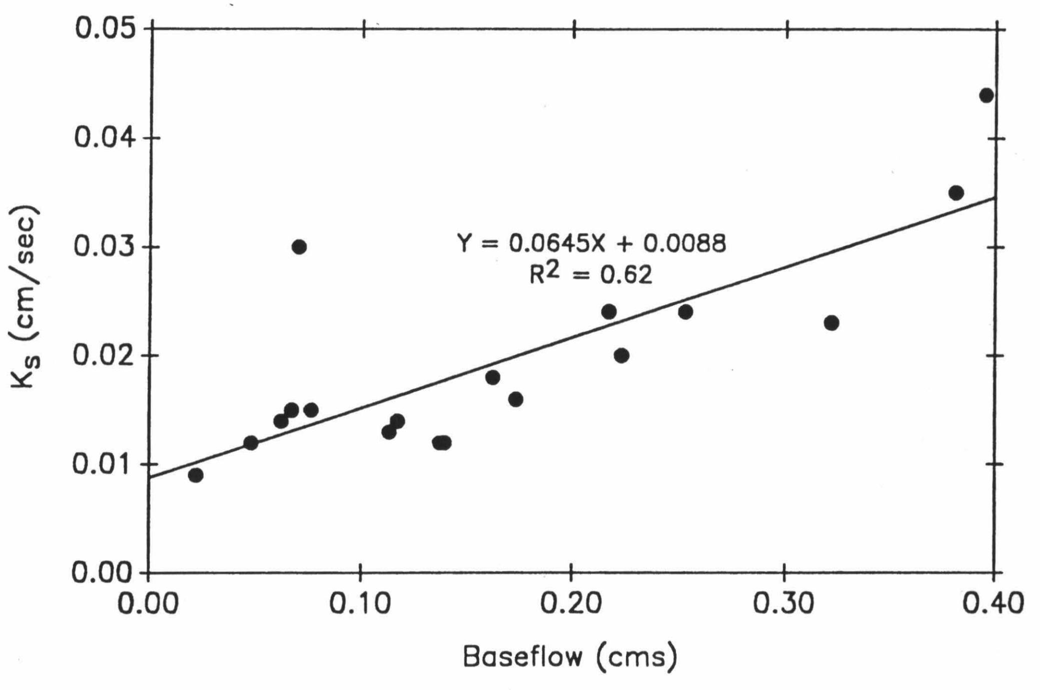


Figure 14.
Conductivity measurements in three monitoring wells (TW3, TW4, and TW5) during the course of the subsurface tracer experiment. Day zero represents the time of injection of the tracer.

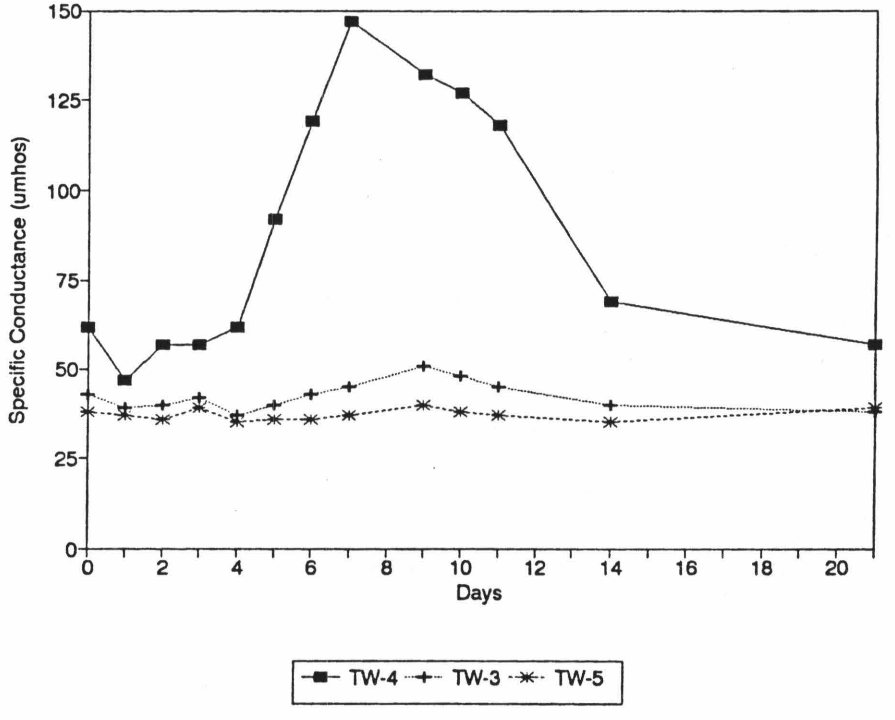


Figure 15.
Results from rising head slug tests performed on two wells during May 1991: (a) well 1WS; and (b) well 2WS.

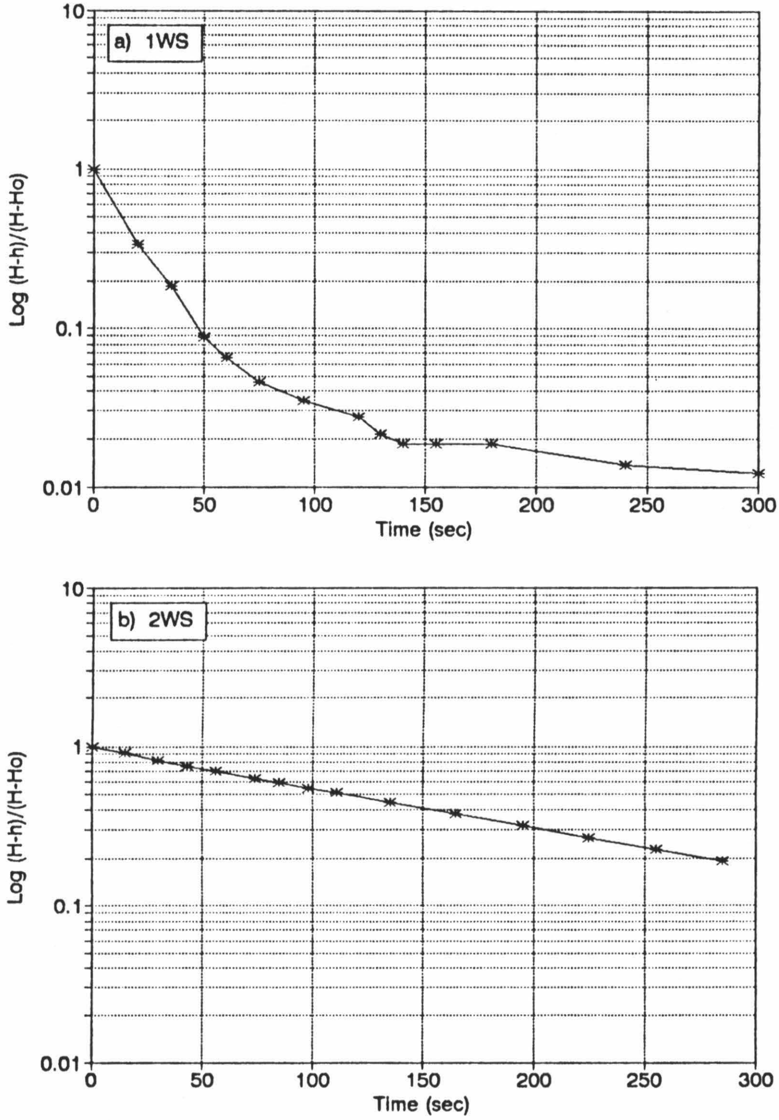


Figure 16.
Wetland water table (well 2WX) and streamflow (site RC2) responses to nine rainfall events during February - April 1991 (lower frame). Vertical bars (top frame) represent daily rainfall intensities measured at the Ashland station.

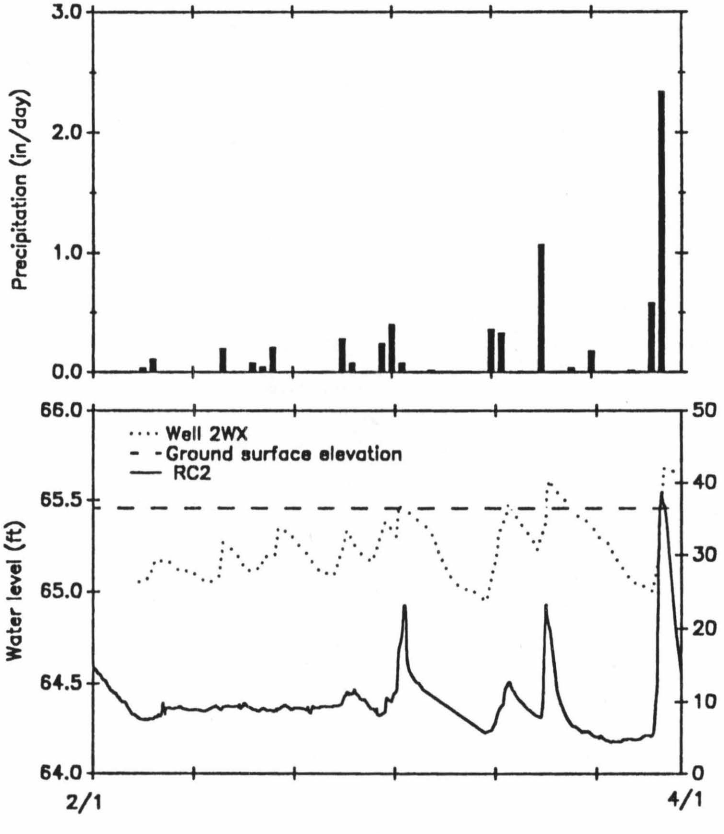


Figure 17.
Creek hydrograph and surface-water chloride concentrations for the August 8-22, 1990 storm event.

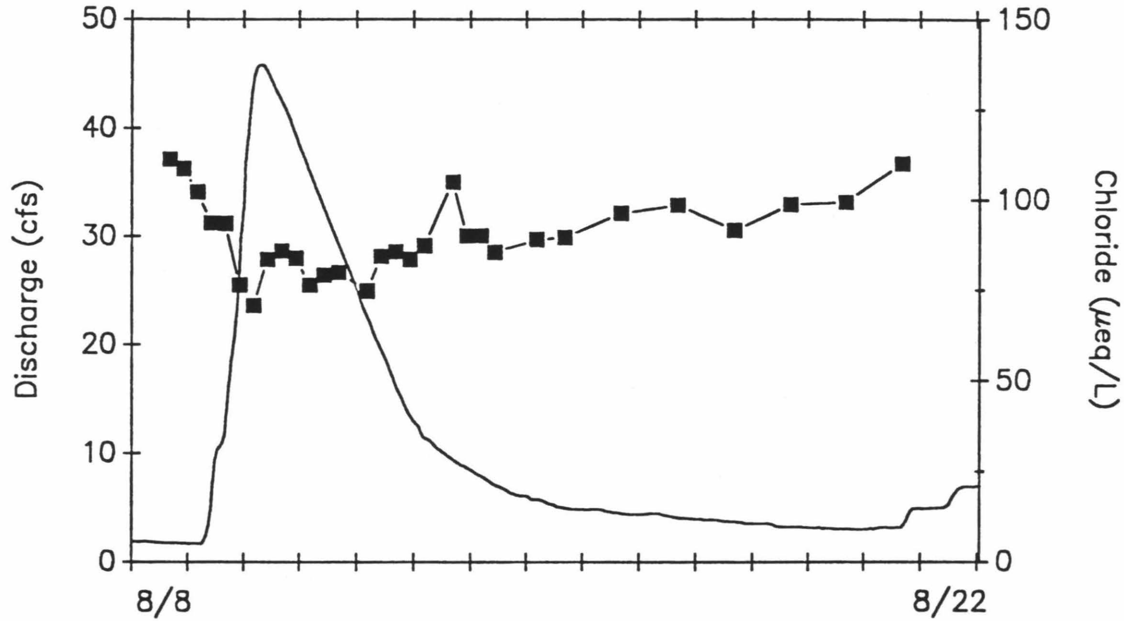


Figure 18.
Creek hydrograph and surface-water chloride concentrations for the two storm events during October 16-29, 1990.

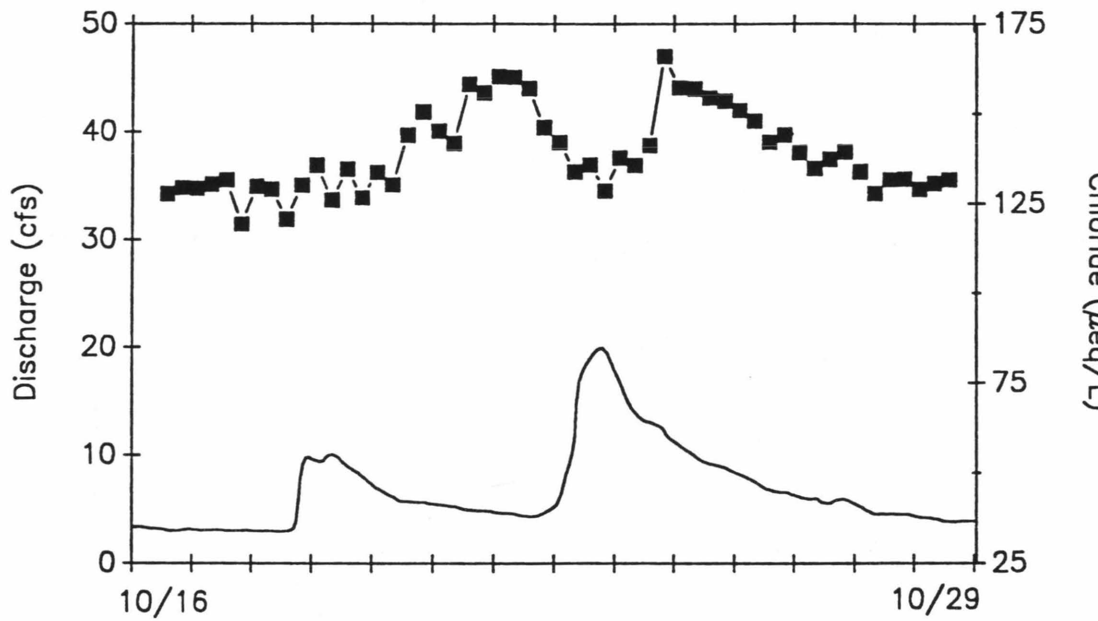


Figure 19.
Creek hydrograph and surface-water chloride concentrations for the three storm events during March 12 - April 6, 1991.

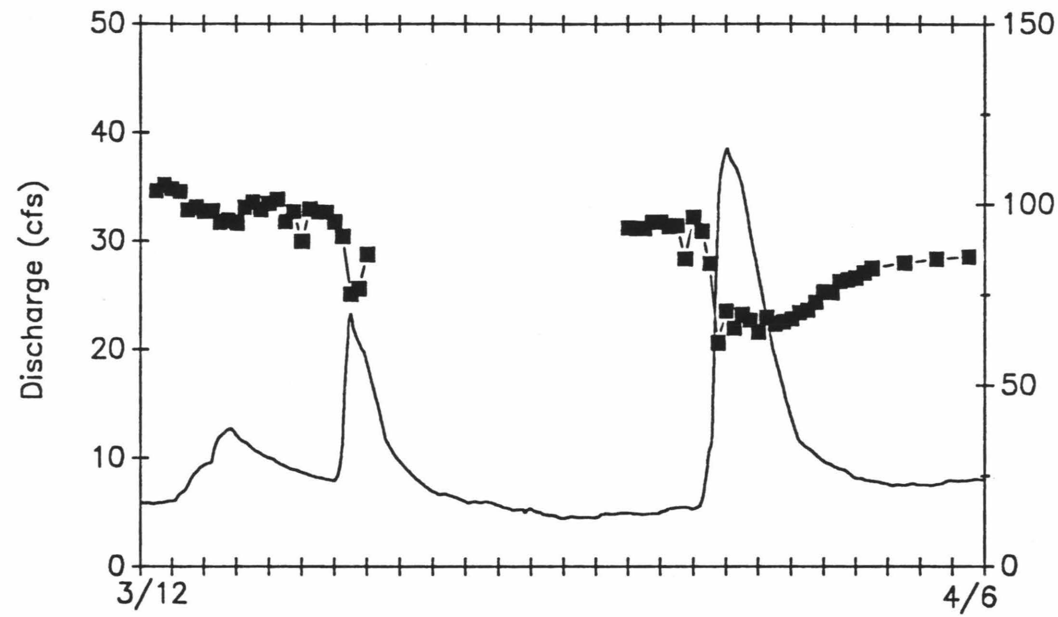


Figure 20.
Creek hydrograph and surface-water chloride concentrations for the July 27-31, 1991 storm event.

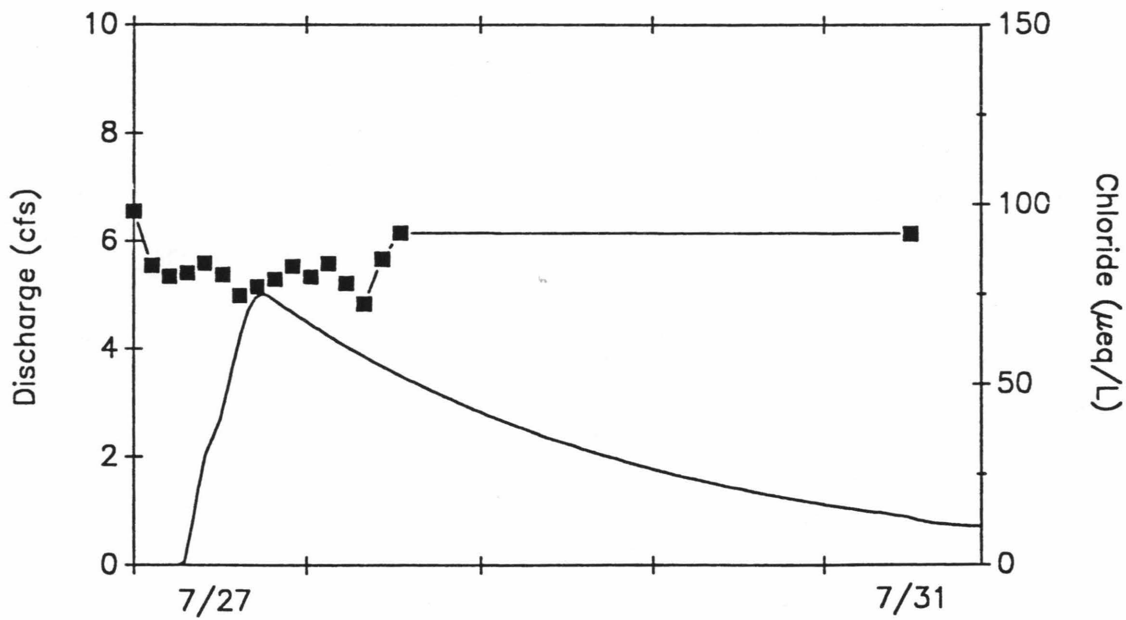


Figure 21.
Creek hydrograph and surface-water chloride concentrations for the September 19-23, 1991 storm event.

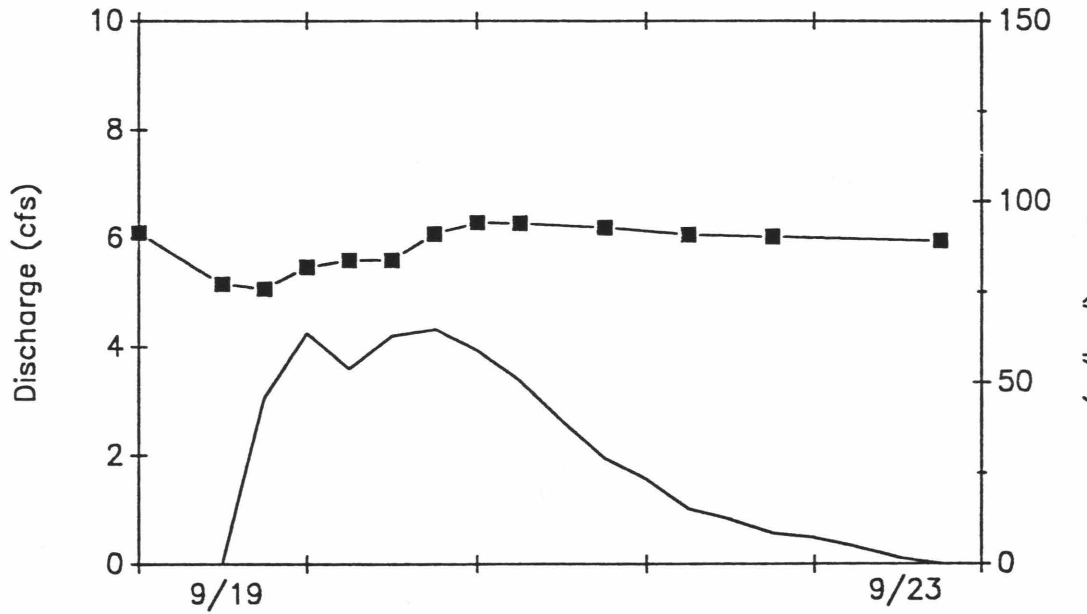


Figure 22.
Results of the chemical separation of the August 8-22, 1990 stormflow hydrograph using naturally occurring chloride. (New-water discharge is the difference between the total measured discharge and the old-water discharge.)

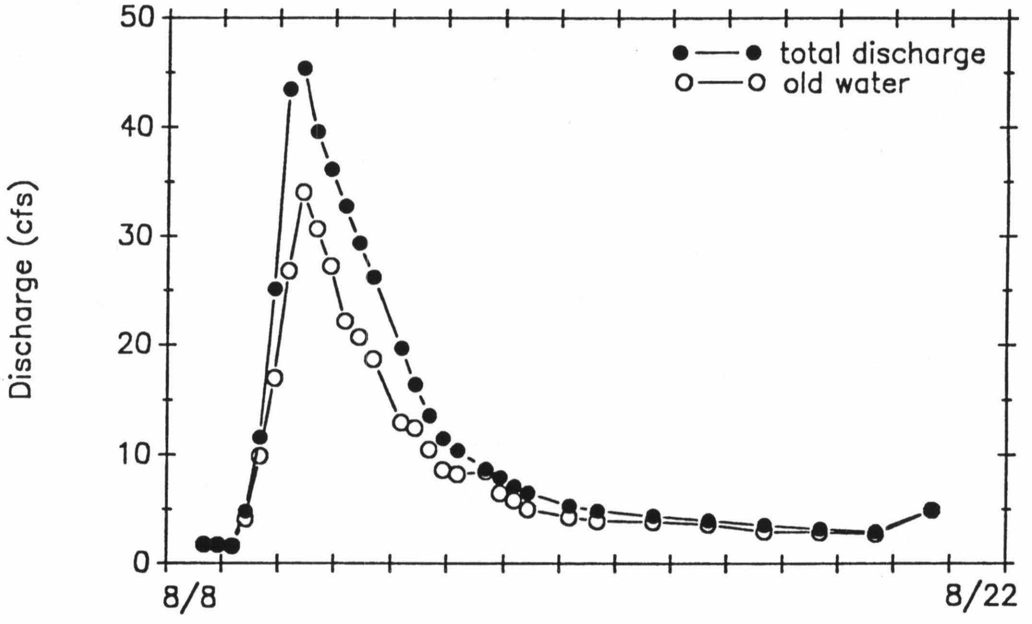


Figure 23.
Results of the chemical separations of two stormflow hydrographs during March 12-18, 1991, using naturally occurring chloride. (New-water discharge is the difference between the total measured discharge and the old-water discharge.)

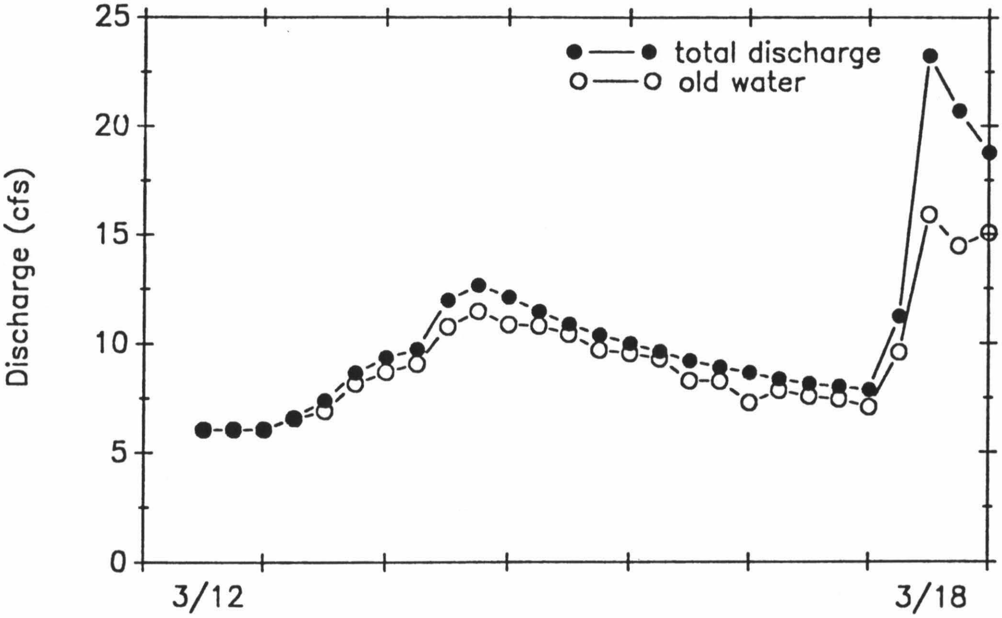


Figure 24.
Results of the chemical separation of the March 27 - April 6, 1991 storm-flow hydrograph using naturally occurring chloride. (New-water discharge is the difference between the total measured discharge and the old-water discharge.)

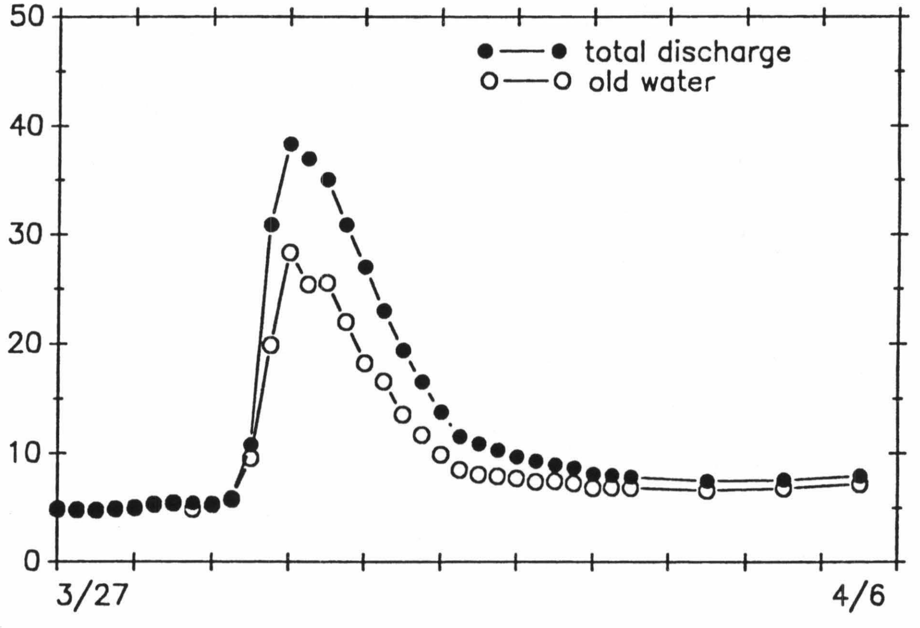


Figure 25.
Results of the chemical separation of the July 27-31, 1991 stormflow hydrograph using naturally occurring chloride.
(New-water discharge is the difference between the total measured discharge and the old-water discharge.)

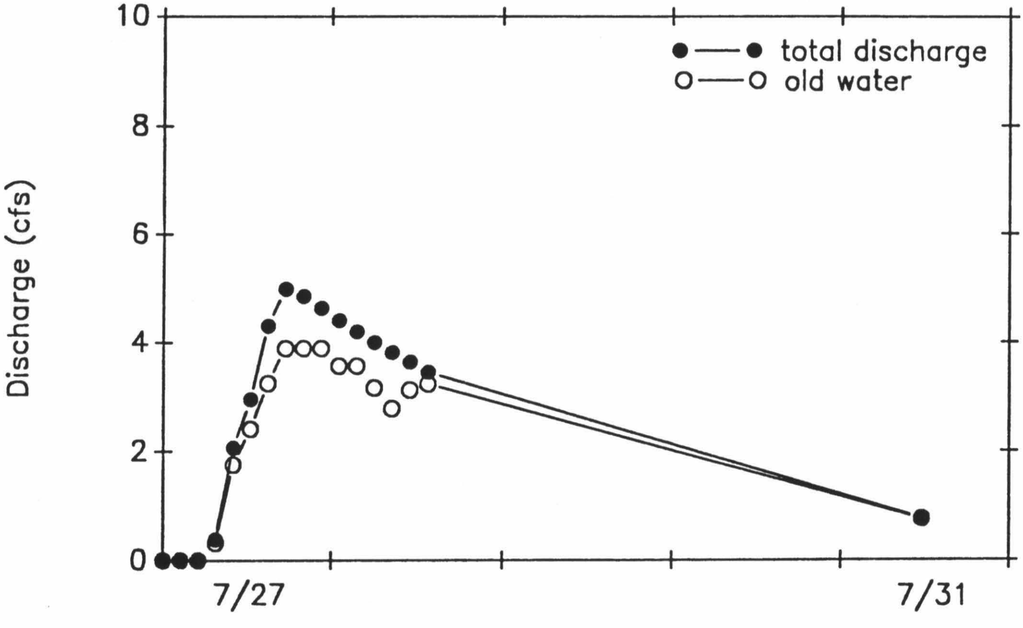
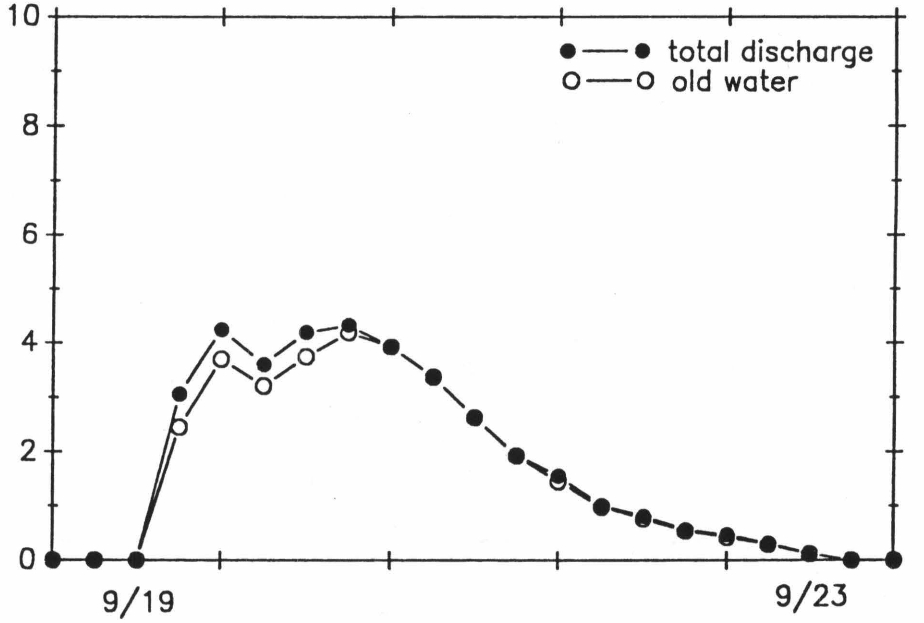


Figure 26.
Results of the chemical separation of the September 19-23, 1991 stormflow hydrograph using naturally occurring chloride. (New-water discharge is the difference between the total measured discharge and the old-water discharge.)



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