

# CONTRIBUTIONS TO THE HYDROLOGY OF SOUTH CAROLINA

Short papers on techniques, procedures, and transferable findings that  
can be used to better understand the hydrology of South Carolina

STATE OF SOUTH CAROLINA  
DEPARTMENT OF NATURAL  
RESOURCES



WATER RESOURCES DIVISION  
REPORT 14

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**State of South Carolina**  
The Honorable David M. Beasley, Governor

**South Carolina Department of Natural Resources**  
**Board Members**

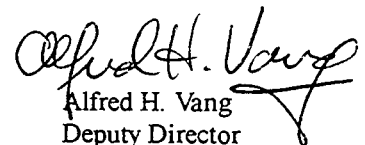
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**FOREWORD**

The following papers by staff members of the Water Resources Division of the South Carolina Department of Natural Resources are presented for the purpose of making available information of a type that would not ordinarily appear in our technical reports. The papers deal with techniques, procedures, and transferable findings that may help water-supply planners and developers to make the best use of the hydrologic data that become available. Also, it is hoped that consultants will be encouraged to obtain some kinds of information that they currently do not consider.

  
 Alfred H. Vang  
 Deputy Director  
 Water Resources Division  
 South Carolina Department of  
 Natural Resources

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# CORRECTION FOR TIDAL EFFECTS ON WATER-LEVEL MEASUREMENTS FLORIDAN AQUIFERS, SOUTHERN COAST OF SOUTH CAROLINA

by

Constance E. Gawne

## ABSTRACT

Water levels in coastal wells fluctuate in response to tides in the adjacent ocean and estuaries. Unless corrected for, these tidal effects obscure static water levels, hydrologic gradients and longer-term water-level changes. Several methods of tide correction are described, and examples are used to evaluate the relative accuracy of each method under different conditions.

Corrections may be made by calculations using measured tide levels; by averaging over a 24-hour, 50-minute diurnal tidal cycle; and by calculations using predicted tide levels. Under optimal tide and weather conditions, corrections made by any of these methods will reduce error in water-level measurements to 25 percent or less of the departures from static level observed in the well. As average tide levels depart from annual mean tide level, for astronomical reasons or because of weather disturbances, corrected levels calculated by using measured estuarine water levels appear to remain near static level. However, systematic errors appear in corrected water levels calculated by averaging or from predictions: Calculations from predictions do not remove weather-related effects, and averaging retains effects of both long-term tidal variations and weather.

## INTRODUCTION

Water levels measured in wells near ocean coastlines may be affected by marine tides. Along the coast of southern South Carolina, where large areas of land and many wells lie near the complex coastline (Fig. 1), maximum astronomical marine tide range exceeds 10 feet, and water levels in wells completed in the upper Floridan aquifer may vary by as much as 70 percent of the tide range. Water-level measurements must therefore be corrected for tidal effects before being used to map static water levels, or to calculate potentiometric gradients, rates of regional water-level decline or recovery, or drawdown and recovery during aquifer tests.

The present paper is not an exhaustive review of tidal effects; rather, it presents and evaluates the methods currently used by the author to correct for such effects in hydrologic studies of the Floridan aquifer system in southern South Carolina.

## CAUSE AND NATURE OF TIDAL EFFECTS

The cause of tidal water-level changes in coastal wells is loading and unloading of the land surface by the weight of flooding and ebbing tidal water (Hayes, 1979, p. 44-46; Parker and Stringfield, 1950, p. 448-450). These pressure changes are communicated to underlying aquifers where waves of potentiometric pressure are generated and move inland through the aquifers. Tidal changes occur in water bodies of different sizes and configurations: along the open coast, in estuaries of various widths and depths, and on tidal marshes and flats. The relative influence of these various tidal changes on water pressures in aquifers is poorly

understood. For a given well, the net water-level rise may result from some combination of loading of a distant nearshore area by a large mass of water, of a nearer estuary floor by a smaller mass of water, and of a nearby tidal marsh by a thin sheet of water for a fraction of the tidal cycle.

Inland transmission of tidal effects is seen mainly in confined aquifers; in unconfined aquifers, water-level changes dissipate within a few hundred feet of the shoreline. In general, within an aquifer the magnitude of the tidal efficiency (ratio of tidal change in the well to the marine tidal range) decreases and the tidal lag (time lag between marine highs and lows and those in the well) increases inland. Comparison of variation in tidal effects among different areas of the upper Floridan aquifer in southern South Carolina suggests that where the aquifer is shallow and poorly confined the tidal efficiency is greater near the shore but dissipates within a shorter distance inland, compared to areas where the aquifer is deeper and better confined. Figure 2 shows hourly measurements of tide elevation at the Town of Port Royal fishing dock and of water levels in four observation wells in southern Port Royal (Fig. 1, location A). The four wells, 27II-134, 27II-126, 27II-128, and 27II-133, are progressively farther inland and up gradient; efficiency decreases, tidal lag increases, and average water level rises inland. Tidal effects also differ vertically within the Floridan aquifer system. Tidal efficiencies in the middle Floridan aquifer, approximately 400 feet deeper than the upper Floridan aquifer (Gawne and Park, 1992), average 75 percent of those for adjacent wells in the upper Floridan aquifer. Lags average slightly longer than for adjacent upper Floridan wells.

The marine tidal cycle may be approximated by a cosine wave; however, tides along the coast of southern

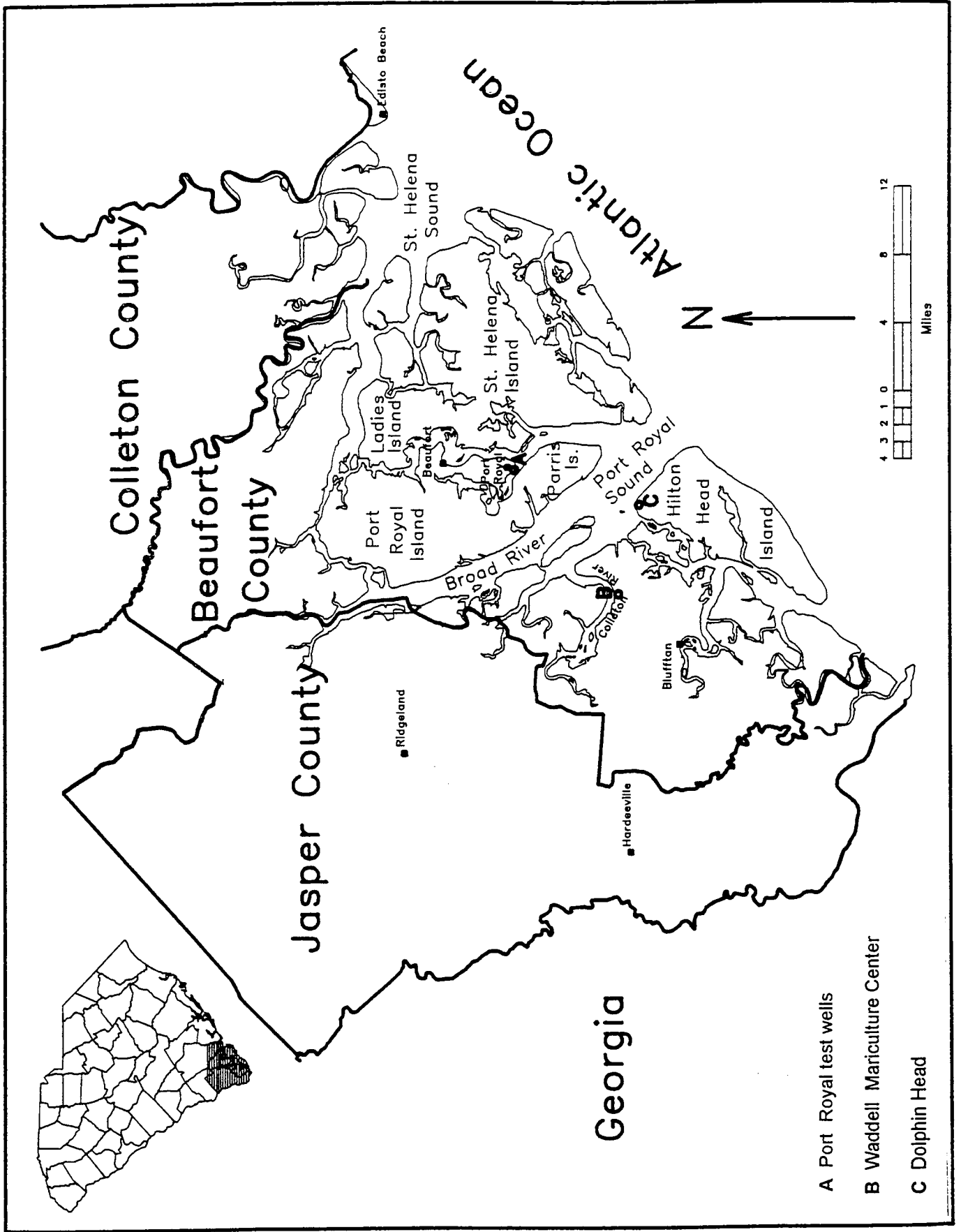


Figure 1. Location and major features of study area.

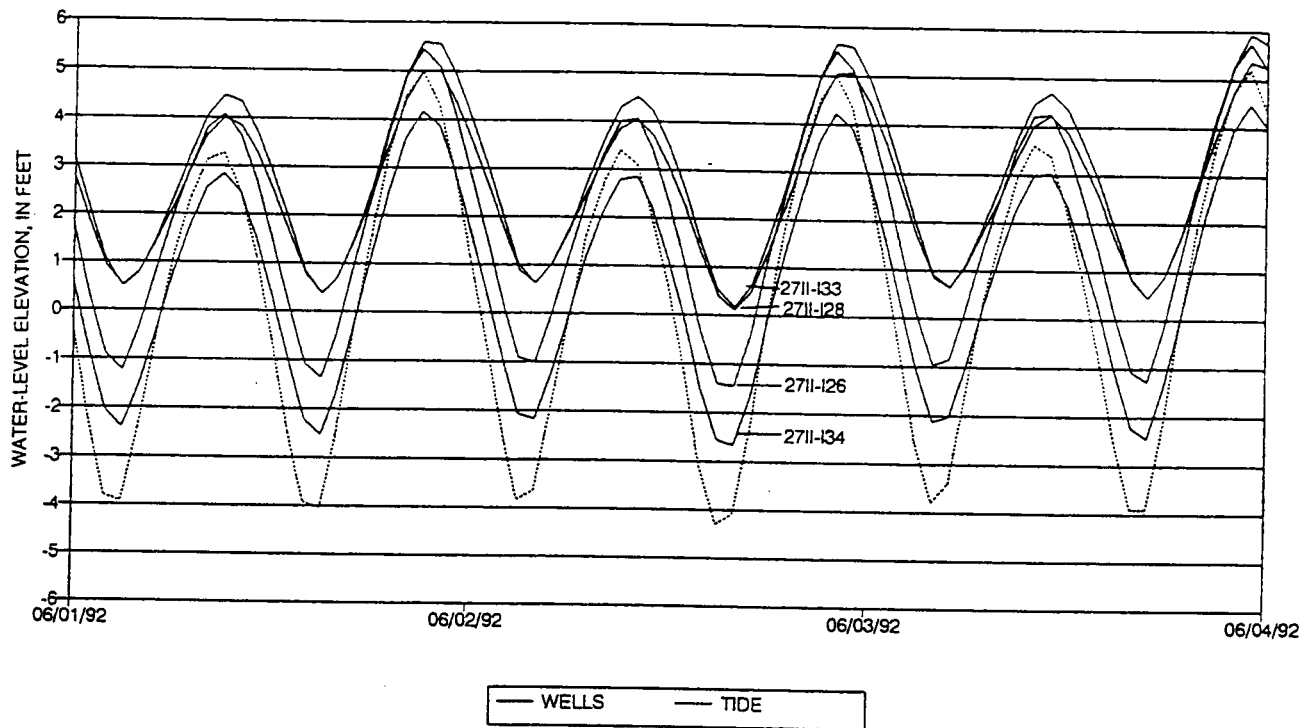


Figure 2. Water levels in Battery Creek and in four observation wells in Port Royal, measured at 1-hour intervals.

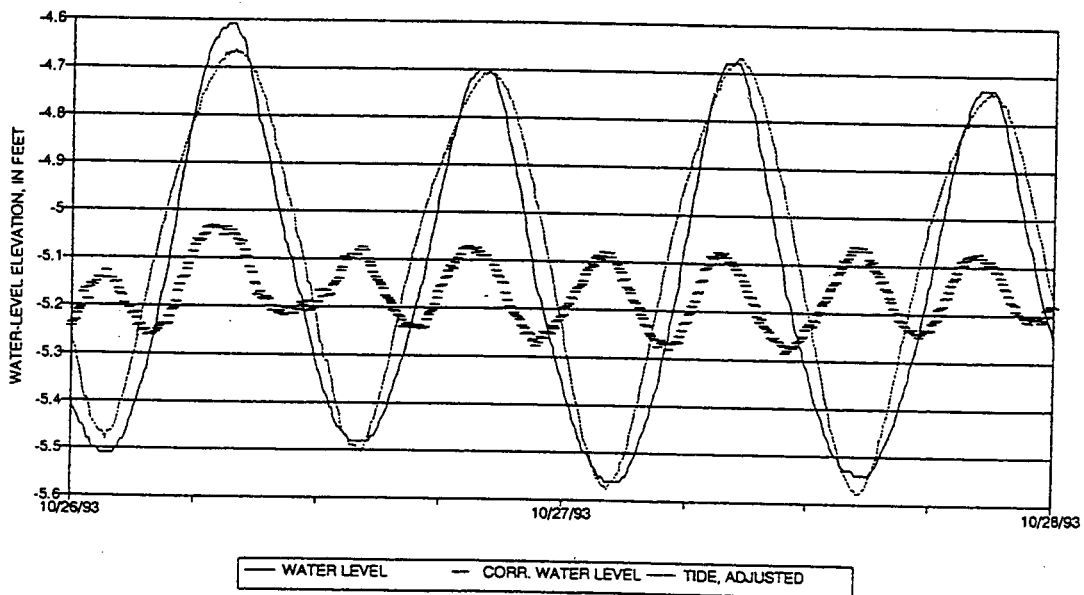


Figure 3. Uncorrected and corrected water levels measured in middle Floridan well 28JJ-p5, with measured tide levels adjusted for well lag and efficiency.



South Carolina are unequal (the daily tidal cycle includes a higher high, lower high, higher low, and lower low tide). This is illustrated in the tide predictions published in the Tide Tables, Table 1, for the Savannah River Entrance station, which is the reference station for the area of study. In estuaries the tides are even more asymmetrical, as shown by correction factors for local stations on Port Royal Sound, St. Helena Sound, and their tributaries (Tide Tables, Table 2).

Comparisons of simultaneous records of water levels in wells and in adjacent water bodies demonstrate that as waves of potentiometric pressure move through aquifers they become distorted. Figure 3 shows water levels recorded in a middle Floridan well near the Colleton River (Fig. 1, location B) and tides recorded simultaneously in the estuary. The tidal record has been shifted by the tidal lag of the well and multiplied by the well efficiency to match the well record as nearly as possible. Water levels in this well rise more rapidly, remain high a shorter time, drop more gradually and remain low longer than the tide in the adjacent estuary. Although the two curves are superficially similar, the vertical difference between them may exceed 0.2 ft (foot), approximately 25 percent of the tidal change in the well. Possible causes of this distortion include variations in aquifer transmissivity, interference by reflected and refracted waves, and interference by waves generated from different water bodies. In extreme cases, water-level fluctuation in wells may be poorly related to marine tides. For example, upper Floridan well 27KK-g4 on Hilton Head Island is 1 1/4 miles south of Skull Creek, a moderate-sized estuary; 2 miles southwest of Port Royal Sound, a major estuary; and 3 miles northwest of the Atlantic Ocean. Over a 4-day period, the tidal lags in this well ranged from 19 minutes to 192 minutes, and the tidal efficiency ranged from 0.03 to 0.14.

### **CALCULATION OF TIDAL EFFICIENCY AND LAG FOR A WELL**

The tidal effect on the water level for a particular well is described by the tidal lag and efficiency (defined above). These must be determined in relation to tides at a specific tide station. Ideally, these factors are calculated from simultaneous data obtained at the well and the tide station; however, simultaneous tidal data often are not available, and tidal constants must be calculated relative to tidal predictions. Typically, a water-level instrument is used to collect data at 3- to 12-minute intervals from the well over a period of several days. Lag is calculated by comparing the time of each high or low well-water level with that of the corresponding high or low tide (observed in simultaneous tide data or calculated for the nearest local tidal station by using the Tide Tables) and

averaging the individual differences to find the lag for the well. High lags and low lags should be averaged separately; in some areas the two are significantly different. In most cases the high or low water level in the well lags behind the marine tide, but in wells very near the coast it may occur earlier (a negative "lag"). The vertical magnitude of each tidal change in the well is divided by the corresponding difference between observed or predicted high and low tides; a number of these ratios are averaged to find the tidal efficiency of the well. Note that it is the changes in level that are compared, not actual elevations; the average water level in the well is usually different from average tide level.

Many wells with pumps have observation ports of small size, permitting water-level measurements by steel tape but not by recording instruments. Collection of sufficient data to calculate tidal lags and efficiencies for such wells is impractical. Tidal lags and efficiencies calculated for other wells nearby may be graphed against distance from the nearest tidal body; tidal constants for wells for which data are not available may then be estimated from the graphs. Hassan (1985, p. 17; Figs. 11, 12) utilized this method of extrapolating tidal constants between wells in his study of the ground-water hydrology of Ladies and St. Helena Islands; Hayes (1979) had previously used it but did not document it in his report.

### **CORRECTION OF MEASURED GROUND-WATER LEVELS**

Several methods may be used to calculate approximate mean water levels from tidally affected water-level measurements, depending on the data available and the precision needed. The accuracy of each method is addressed in the following discussion.

### **CORRECTION OF INSTRUMENTAL DATA BY AVERAGING**

Measurements made at regular intervals through a complete daily tidal sequence may be averaged to give the arithmetic mean water level in the well. Where tides are unequal, a complete tidal sequence includes two tidal cycles, a total time of approximately 24 hours and 50 minutes. If measurements are recorded hourly, data for 25 consecutive hours should be averaged. Either a single daily average or a running average may be calculated. Daily ground-water levels published by the U.S. Geological Survey for wells in South Carolina (for example BFT-1810, 27JJ-q3; Bennett and others, 1994, p. 465) are averages of 24 hourly measurements; such daily averaging removes most of the tidal effect, with the limitations discussed below.

**CORRECTION OF WATER-LEVEL  
MEASUREMENTS BY USING  
SIMULTANEOUS TIDE DATA**

This method requires data from a recording tide gage. Because of the large tide range and rapid water-level changes observed on the coast of southern South Carolina, frequent tidal measurements are needed; good results have been obtained with measurements at 6- and 12-minute intervals. The tide data are recalculated as departures above (+) and below (-) mean tide.

The tidal lag and efficiency for the well relative to the tide station are found as described earlier. Correction of a water-level measurement is then straightforward: subtract the lag time from the measurement time, and select the corresponding tidal water level; multiply the tidal water level by the tidal efficiency of the well to obtain the tidal correction; and subtract the tidal correction from the water-level elevation in the well. This method was used by Hayes (1979, p. 47).

For periodic instrumental measurements, using tide data taken with the same frequency, the procedure is the same. The calculations may be made conveniently by using a spreadsheet, with tide and water-level elevation data in parallel columns, to produce a column of corrected water elevations. The corrected water levels shown on Figure 3 were calculated by this method. A representative part of the spreadsheet used is shown below.

A similar method was used to correct water-level data taken at 1-hour intervals over a period of 16 months from four wells in Port Royal having different lags and efficiencies (Fig. 2). Tide corrections were made with 12-minute data from a tide gage in nearby Battery Creek. A FORTRAN routine was written to select the appropriate tide measurement for each well measurement, calculate the tide correction and corrected water-level elevation, and generate corrected water elevations for each ground-water data set. A spreadsheet was then used to plot hydrographs. Figure 4 shows tide data, uncorrected water elevations, and corrected water elevations for well 27II-128 for February 1992.

Sample Spreadsheet for Calculation of Ground-water Levels by Using Simultaneous Tidal Data

Row	Tide at WMC Pier		Water level in well 28JJ-p5			
	Date	Time	Tide level	Ground-water level	Explanations and formulas for column E	Corrected ground-water level
Column	A	B	C	D		E
4					Efficiency:	0.125
5					Lag: 60 minutes	
6	10/26/93	05:00	2.67			
7	10/26/93	05:06	2.80			
8	10/26/93	05:12	2.86			
9	10/26/93	05:18	2.99			
10	10/26/93	05:24	3.12			
11	10/26/93	05:30	3.19			
12	10/26/93	05:36	3.25			
13	10/26/93	05:42	3.31			
14	10/26/93	05:48	3.38			
15	10/26/93	05:54	3.44			
16	10/26/93	06:00	3.51	-4.77	+D16-C6*E4	-5.10
17	10/26/93	06:06	3.57	-4.75	+D17-C7*E4	-5.10
18	10/26/93	06:12	3.64	-4.74	+D18-C8*E4	-5.10
19	10/26/93	06:18	3.70	-4.72	+D19-C9*E4	-5.09
20	10/26/93	06:24	3.70	-4.70	+D20-C10*E4	-5.09
21	10/26/93	06:30	3.83	-4.69	+D21-C11*E4	-5.09

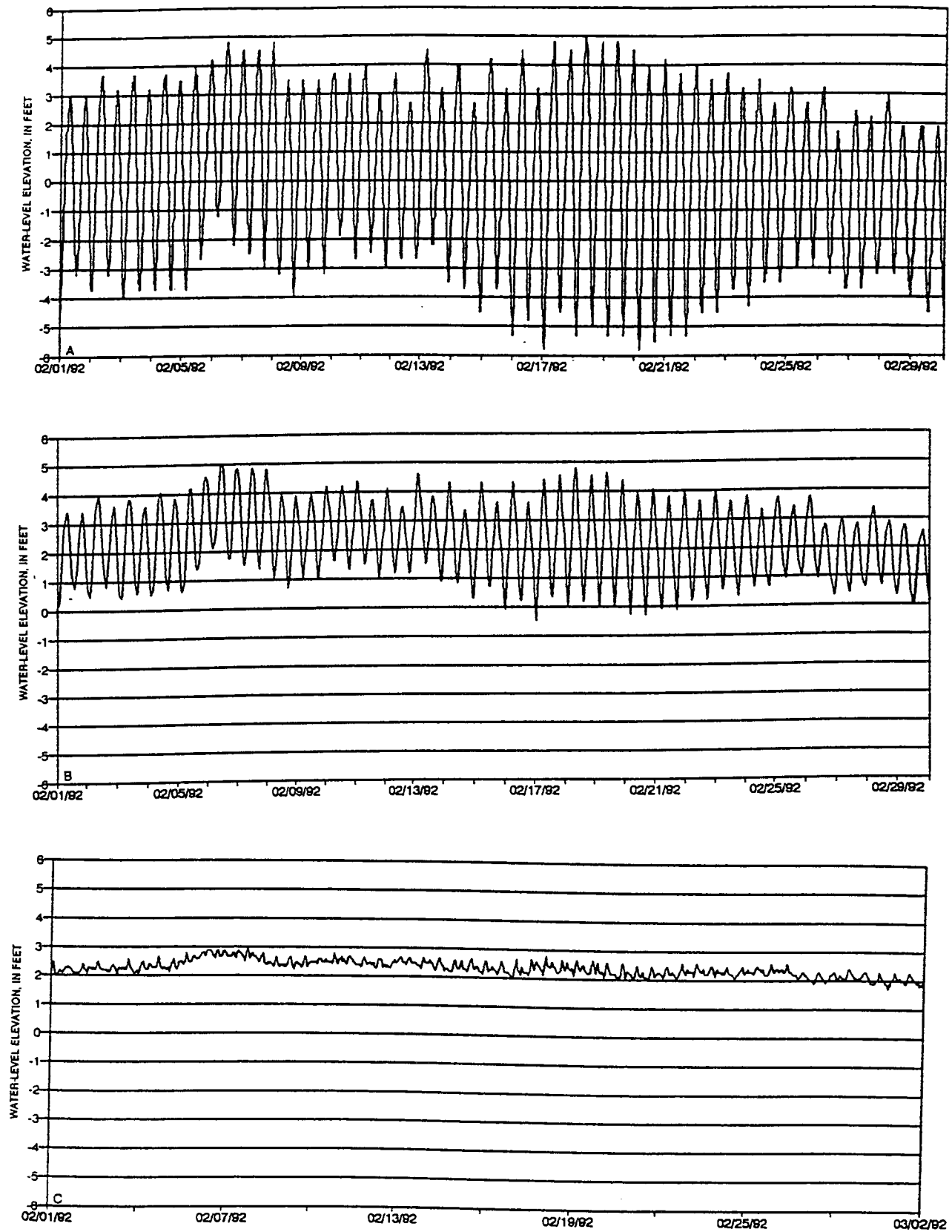


Figure 4. Tide in Batter Creek and water levels in well 27II-128, February 1992. A, Water levels in Battery Creek, February 1992. B, Uncorrected water levels, well 27II-128. C, Corrected water levels, well 27II-128.

## **CORRECTION OF WATER-LEVEL MEASUREMENTS BY USING TIDE PREDICTIONS**

For most water-level measurements, simultaneous tidal measurements are not available. Corrections can be made by using predictions from the Tide Tables. As discussed below, there are limitations to this method: weather conditions can cause large departures of actual tides from predicted tide levels, and interpolation of tidal levels between predicted high and low tides depends on the assumption that water-level changes can be described by a cosine function (Tide Tables, Table 3). The latter assumption is most nearly met along the open coast, but it becomes less and less true toward the landward extremes of the estuary system. Nevertheless, corrections based on predictions can remove much of the tidal effect from water-level elevations.

Hassan (1985, p. 17; Fig. 13) described a graphical method for calculating tide corrections from predictions. A spreadsheet has been developed by the author to accomplish the same purpose. Details of the spreadsheet, with instructions for installation and use, are given in the Appendix. The spreadsheet shown is used to correct a single measurement. Other versions of this spreadsheet have been prepared for correcting a number of measurements, with the calculations for each correction arranged on a single line. Uses include calculating corrections for measurements from a number of wells, calculating a series of corrections for instrumental measurements from a single well, and preparing a predicted tide curve for comparison with data. These spreadsheets were employed for correcting many of the water-level measurements used in preparing potentiometric maps and hydrographs for Beaufort, Colleton, Hampton, and Jasper Counties, South Carolina (Gawne, 1994). Figure 3 of that report shows locations of the water-level measurements to which corrections were applied.

### **CORRECTION OF AQUIFER-TEST DATA**

Tidal effects cause serious problems in the interpretation of aquifer tests because the changes in water level observed in late stages of pumping or recovery are small compared to the magnitude of tidal changes seen in many coastal wells. A test of 24 hours or more will show obvious undulations of water level in later stages of the test. In the worst case, a test 6 hours in duration begun at high or low tide might have apparent drawdown increased or decreased by several feet without apparent distortions in the data. An aquifer test made in any coastal well should be preceded by sufficient

observation of static water level to determine whether tidal correction is necessary.

In most cases, simultaneous tidal measurements are needed to correct tidally affected aquifer-test data. Pretest observations should be made for the purpose of calculating the tidal efficiency and lag of wells relative to the tide station (Fetter, 1988, p. 206-207; Kruseman and de Ridder, 1983, p. 180-181). The author's practice has been to make pretest observations for a period of at least 3 days and with measurements taken at intervals of 12 minutes or less.

The tidal measurements may be made at a temporary tidal station established at the nearest feasible point on the nearest major tidal water body. Alternatively, a well may be used as a tidal station. The well should be far enough from the site of the aquifer test to be unaffected by pumping and have at least as great a tidal efficiency as the well or wells being monitored for the test. The pretest should be used to choose the frequency of measurements needed for the aquifer test. The greater the tidal efficiency, the greater the frequency of measurements needed for tide correction. During the test, early water-level measurements may be taken at a greater frequency than the tide-gage measurements. During later stages of the test, the frequency of measurements should be maintained high enough to allow evaluation of the quality of the tide-correction calculations. Maintaining such a high frequency of measurements through an extensive pretest, pumping test, and recovery period requires either automated instruments or sufficient manpower to provide continuous measurements at each station. Recording data electronically facilitates tide correction as well as water-level measurement. The author's practice is to measure water levels electronically in wells and at the tide gage at intervals of 12 minutes or less, while making less frequent manual measurements for calibration and backup of electronic records.

The specific procedure for calculation depends on the format of the data and on the software used. The procedure used by the author is as follows:

- 1) The tide data are loaded into a spreadsheet and recalculated as necessary. They should be centered around mean sea level by subtracting the arithmetic mean of all the measurements from each measurement. The tide data should be oriented in the same sense as the aquifer-test data; for example, if the aquifer-test data are in the form of measurements from top of casing, as positive numbers, either the tide data should be inverted, or water levels should be recalculated as elevations (in step 2 below). Because the tide-correction process usually alters the initial water level of the test, the author has found it more convenient to calculate drawdown after tide correction.

2) The aquifer-test data are loaded in the spreadsheet in columns parallel to the tide data, and any necessary recalculation is done.

3) The tidal lag found from pretest data is used to relate each measurement of the aquifer test to the appropriate tidal measurement. For measurements made early in the test, when aquifer-test measurements are more frequent than tidal measurements, this may require interpolating between tidal measurements.

4) A formula is written to multiply the selected tidal measurement by the well's tidal efficiency to give the tidal correction. The tidal correction is subtracted from the well's water level to give the corrected water level. These may be combined into one formula. A similar calculation is performed for each measurement, (Table 1, Appendix).

5) The corrected data are plotted. If tidal effects remain, minor adjustments may be made to the tidal efficiency and lag to improve the correction.

6) The data are recalculated as drawdown and are plotted in the graphical format desired for interpretation (on semilog or log-log scales).

Figure 5 shows data from the middle Floridan piezometer of multipiezometer well 27JJ-x1, which was used as an observation well during an aquifer test at well 27JJ-q3, also middle Floridan (Fig. 1, Location C). For this test, an upper Floridan well, 27JJ-q1, was used as a tide gage. For this illustration, drawdown was calculated as negative numbers for convenience in plotting. In the middle Floridan piezometer, tidal changes lagged behind those in well 27JJ-q1 by 8 minutes, and the efficiency was 0.66 times that of the upper Floridan well. In this case, tidal effects during the course of the test were more than twice as great as water-level changes caused by pumping and recovery.

## ACCURACY OF CORRECTED WATER LEVELS

### SOURCES OF ERROR

Each method discussed above is subject to one or more sources of error. It is assumed in the following discussion that true static water level is that which would be observed in the absence of any variation in water level in adjacent marine water bodies.

1) **Averaging:** This method removes the effects of the cyclic changes occurring within the 24-hour and 50-minute period that is the average duration of a complete diurnal tidal cycle. However, predictions for the Savannah River Entrance station (Tide Tables, Table 1) show that the length of the tidal cycle actually varies through time. The average predicted water level also

can show significant departures from mean tide; the average water level predicted for a particular station may be higher or lower than annual mean tide for days or weeks at a time. Effects of these tidal variations will not be removed by averaging over a diurnal tidal cycle. The effects of weather-generated water-level changes that occur within the 24-hour and 50-minute cycle will be reduced, but not removed, by averaging; effects of those that last longer than the tidal cycle may not be reduced at all.

2) **Corrections using simultaneous tidal data:** Because the measured tide levels also include weather-generated changes, corrections based on them might be expected to remove all effects of varying water levels. However, some tide-related effects almost always remain. Comparison of tidal records and well records suggests the following reasons:

a) The tidal lag and efficiency are not constant. For example, lags of 14 and 32 minutes were calculated for well 28JJ-p6, in the upper Floridan aquifer near the Colleton River, from two sets of measurements several months apart, using tide data collected at the same position on the estuary. Efficiencies, although more constant than lags, show minor variation.

b) The distortion of tidal wave form within the aquifer leads to systematic error in calculated tide correction. In Figure 3, the adjusted tide has the same form as the correction factor that is subtracted from the observed water level. Ideally the traces of water level and correction factor would be identical. Where the well-water level is higher than the correction factor, the calculated corrected water level is higher than average; where it is lower, the corrected water level is below average. In this case, the residual error in corrected water levels is approximately  $\pm 0.1$  ft, reduced from approximately 0.5 ft in the original measurements.

3) **Corrections using predictions (Appendix):** These are subject to most of the sources of error seen in both of the above methods (weather-related alterations of tide levels, variations in lag and efficiency, distortions of tidal wave forms within aquifers) and also to error caused by failure of tidal cycles to conform to the ideal cosine-wave model. Effects of distortion of tides within estuaries should be minimized by applying correction factors for the nearest local tide station (Tide Tables, Table 2). Because of the unequal tides prevailing along the South Carolina coast, it is important that the part of the cosine wave used to calculate the tide correction be based on the high and low tides immediately preceding and following the time (adjusted by tidal lag) of the water-level measurement; extrapolating the cosine calculation into the preceding or following half-cycle will introduce additional error.

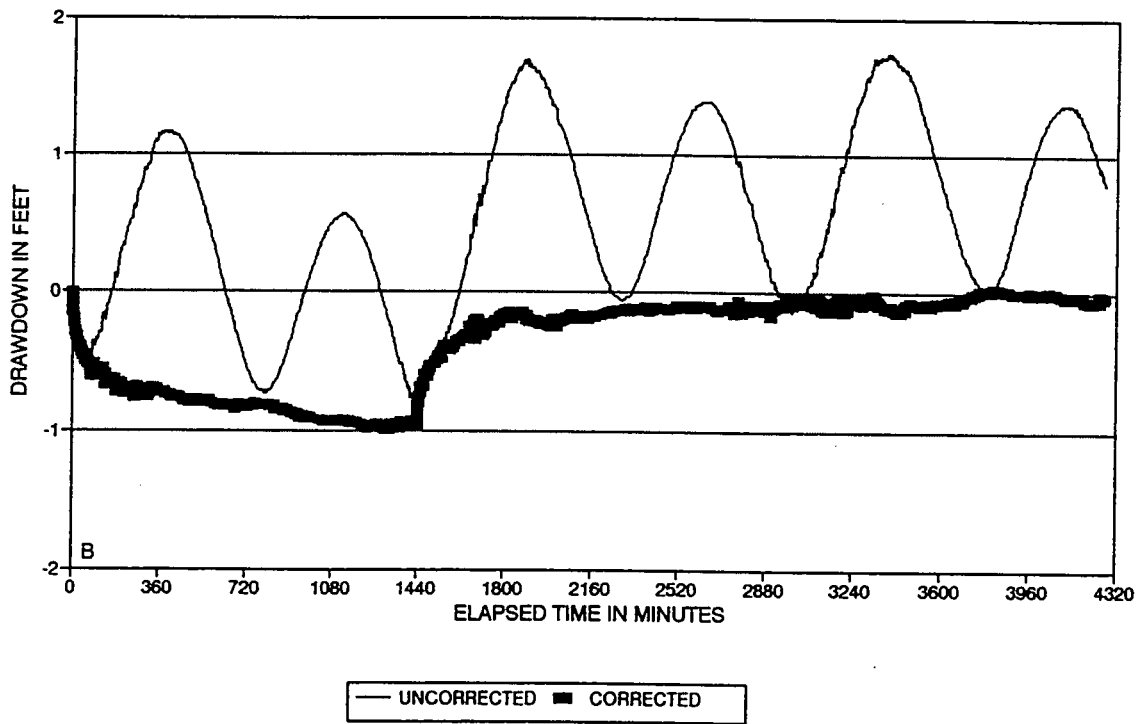
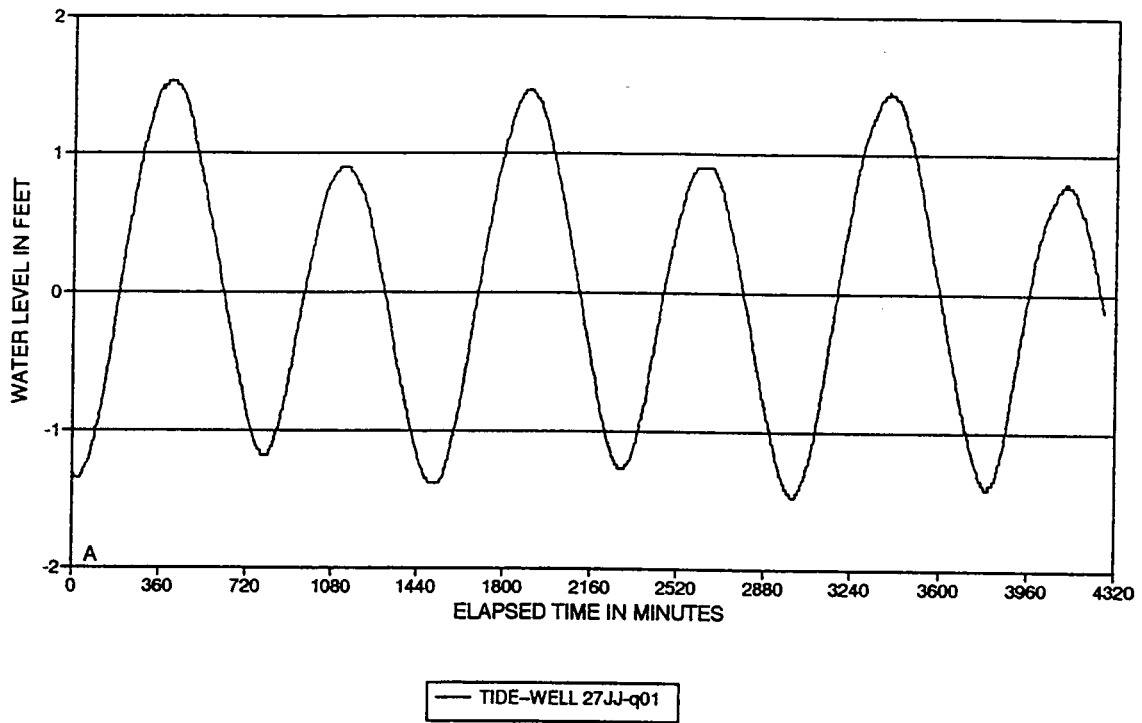


Figure 5. Tide correction of aquifer-test data, middle Floridan aquifer, Hilton Head Island. A, Water levels in upper Floridan well 27JJ-q1, used as tide gage. B, Uncorrected and corrected aquifer-test data from middle Floridan observation well 27JJ-x1.

## COMPARISON OF CORRECTION METHODS

All three methods were applied to several data sets to compare their effectiveness (Figs. 6-8). In each case a version of the spreadsheet shown in the Appendix was used to calculate predicted tides, based on predicted highs and lows at the Savannah River Entrance station as corrected for the nearest local station, for comparison with observed tides. Running averages of water levels shown were calculated over a period as near to 24 hours and 50 minutes as data frequency permitted, centered at the time for which the average is shown; the data used for averaging extended before and after the graphed interval.

1) Well 27JJ-128, upper Floridan aquifer, Port Royal (Fig. 1, location A), Feb. 20-23, 1992; lag 42 minutes, efficiency 0.48, relative to a tide station at the Port Royal fishing dock on Battery Creek (Fig. 6; see also Figs. 2 and 4). The weather during this interval was calm, and observed tides conformed closely to those predicted (Fig. 6A); however, both averaged below mean tide level, observed tides by 0.3 ft and predicted tides by 0.6 ft. The tide range decreased from nearly 10 ft to 7 ft over this period, and the range in water levels decreased from slightly over 4 ft to slightly over 3 ft (Fig. 6B). The corrections calculated by averaging form a smooth curve with very little height variation. The corrections calculated from tide data form a much more irregular line. Part of this can be ascribed to deformation of the tidal wave form within the aquifer; part is caused by the imprecision of the tide gage, a pressure probe reading in steps of approximately 0.3 ft. The tide correction, at a well efficiency near 0.5, cannot be more precise than  $\pm 0.15$  ft. Over any tidal cycle the range of calculated corrections is approximately 15 percent of the water-level range in the well. Figure 6B also shows corrections calculated from tide predictions at 1-hour intervals. These have approximately the same range and pattern of variation as those calculated from data, suggesting that in this case also, distortion of the tidal wave form in the aquifer is the major source of error.

For this record, all three methods appear to give results acceptable for most uses. The downward displacement of the running average compared to the other two corrections reflects the failure of this method to correct for the effect of the difference between average tide levels over this period and annual mean tide.

2) This example is also from well 27JJ-128, Feb. 4-7, 1992 (Fig. 7; see also Fig. 4). During this period offshore storm winds pushed water levels in Port Royal Sound and its tributaries to unusual heights; the recorded tide during this period averaged 0.9 ft above mean tide level, although it had been predicted to average 0.5 ft below mean tide. On Feb. 6 and 7 both high and

low tides also lagged significantly behind their predicted times.

The greater the departure of the tide from predictions, the greater was the separation between tide corrections made by using the three methods. Corrections made from tide data came closest to holding a constant average level through this weather-generated disturbance. The rise of over 0.5 ft shown on Fig. 7B and on Fig. 4C may represent a failure of the tide-correction calculation to remove all of the effects of extra water in the estuary. However, it may represent recharge from rainfall (0.81 inch was recorded at Beaufort on Feb. 5, and 1.5 inches were recorded at Hilton Head Plantation on Feb. 6 and 7), or barometric effects (see below), or both. The running average of water levels removed the cyclical tidal effects, but it did not remove the weather-generated effects. The corrections calculated from predictions were consistently high, averaging 0.6 ft above those based on data. The higher the tide above predicted level, the higher was the corrected water level above that calculated from observed tide level. At the height of the weather-generated rise, some corrections made from predictions were as much as 1.4 ft higher.

3) Well 28JJ-p6, upper Floridan aquifer, Waddell Mariculture Center, south side of the Colleton River (Fig. 1, Location B), Oct. 26-Nov. 3, 1993; lag 24 minutes, efficiency 0.15 (Fig. 8). Throughout this period, water levels were predicted to average nearly a foot above mean tide level; the tide range increased to a spring tide on Oct. 10 at 9:15 AM and then decreased. Estuarine water levels were further disturbed by strong winds associated with a major storm system, causing major departures from predicted levels over this period (Fig. 8A).

The record from well 28JJ-p6 (Fig. 8B) is complex. The potentiometric pressure in the aquifer was undergoing a gradual decline because of an ongoing drought. Superimposed on this were the effects of water-level changes in the estuary. The water rise corresponding to the spring tide of Oct. 30 was disproportionately high.

Three tide corrections are shown: a running average, a correction using tide data (measured with a pressure probe reading in steps of 0.064 ft), and individual corrections calculated at 3-hour intervals from tide predictions. At times when water levels in the sound were near those predicted, on Oct. 28 and Nov. 3, the corrections from data and from predictions are similar; that from averaging is higher because it does not remove the effects of the long-term tidal high. For times when estuary water levels were higher than predicted, on Oct. 26-27 and 29-30, the corrections from predictions are high; for times when water

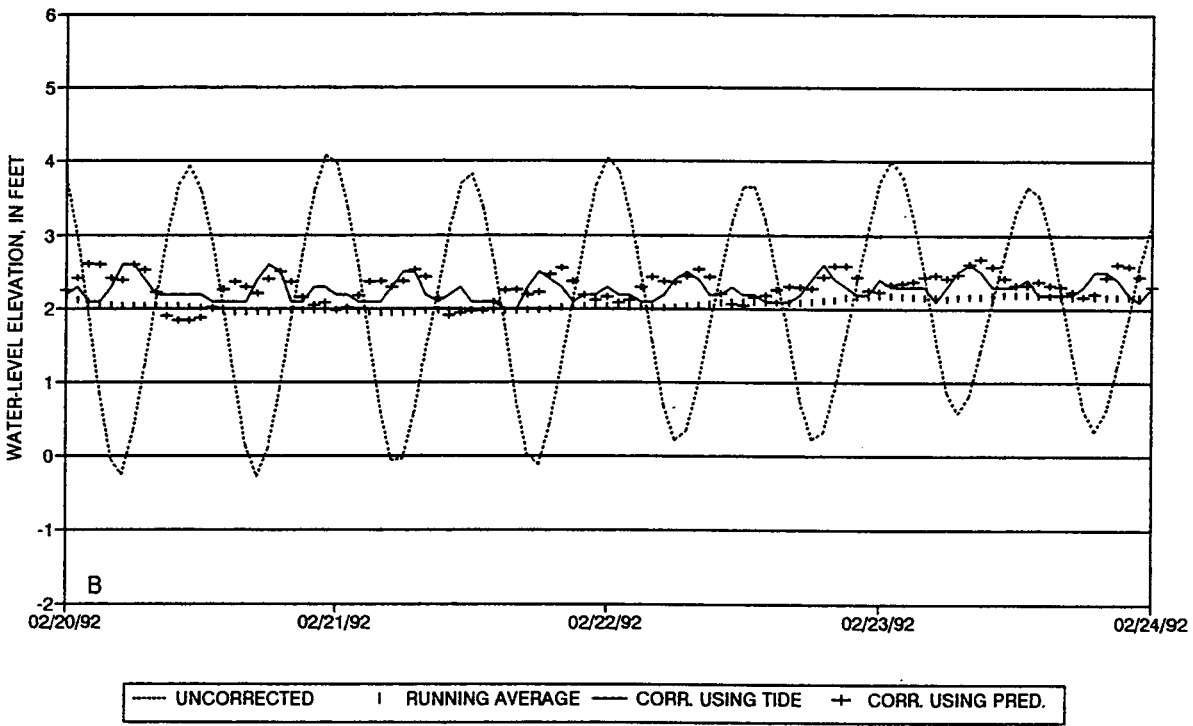
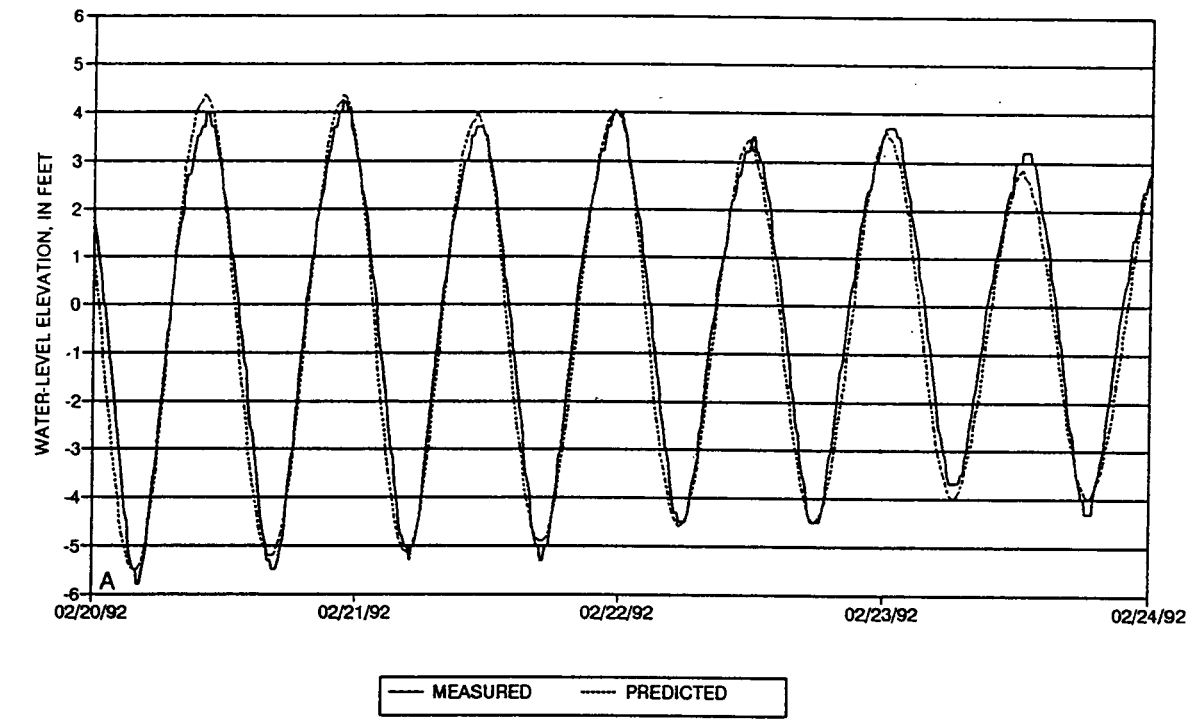


Figure 6. Comparison of tide-correction methods, Port Royal, calm weather. A, Water levels in Battery Creek, measured at 12-minute intervals. B, Uncorrected and corrected water levels, upper Floridan well 27II-128, at 1-hour intervals.



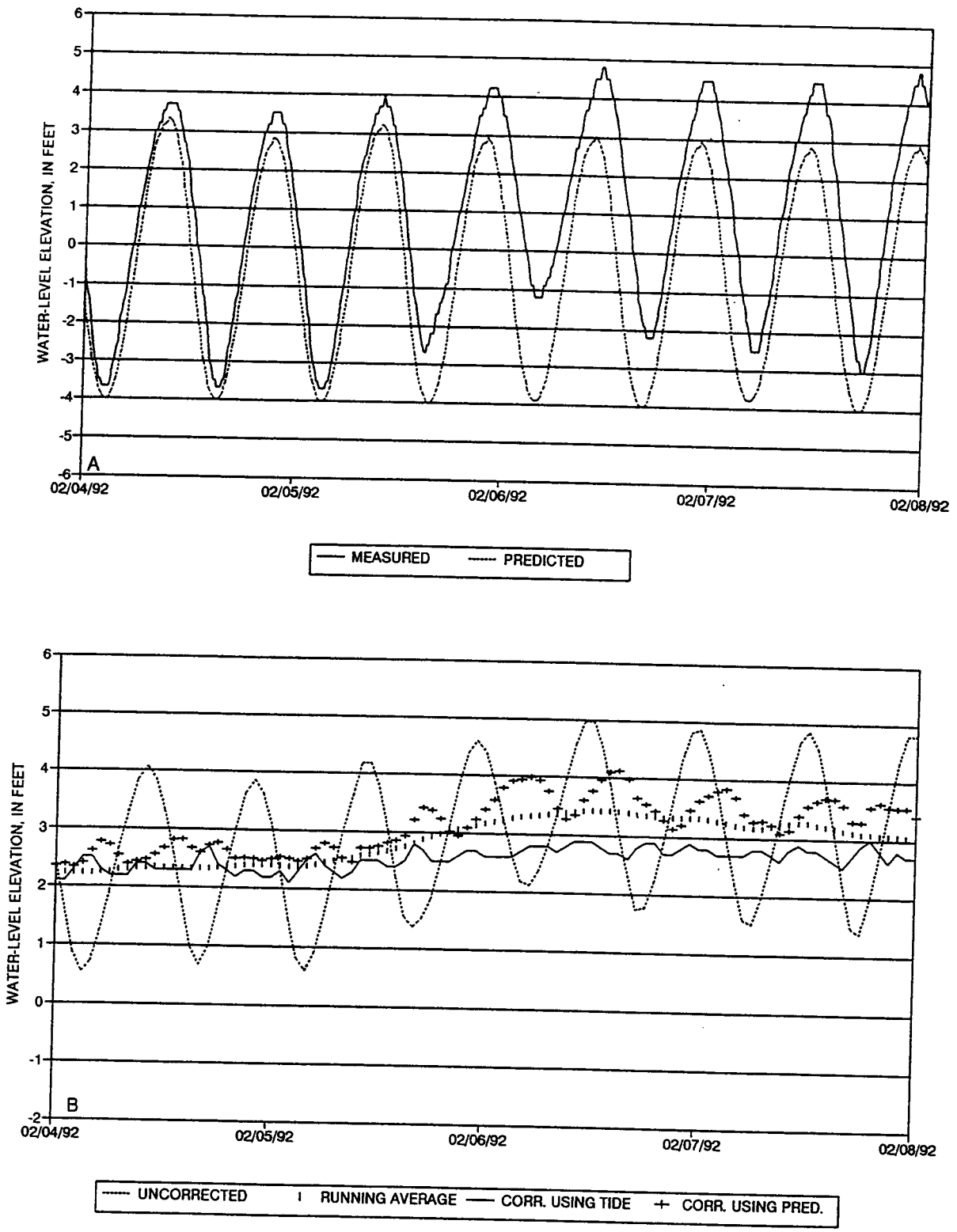
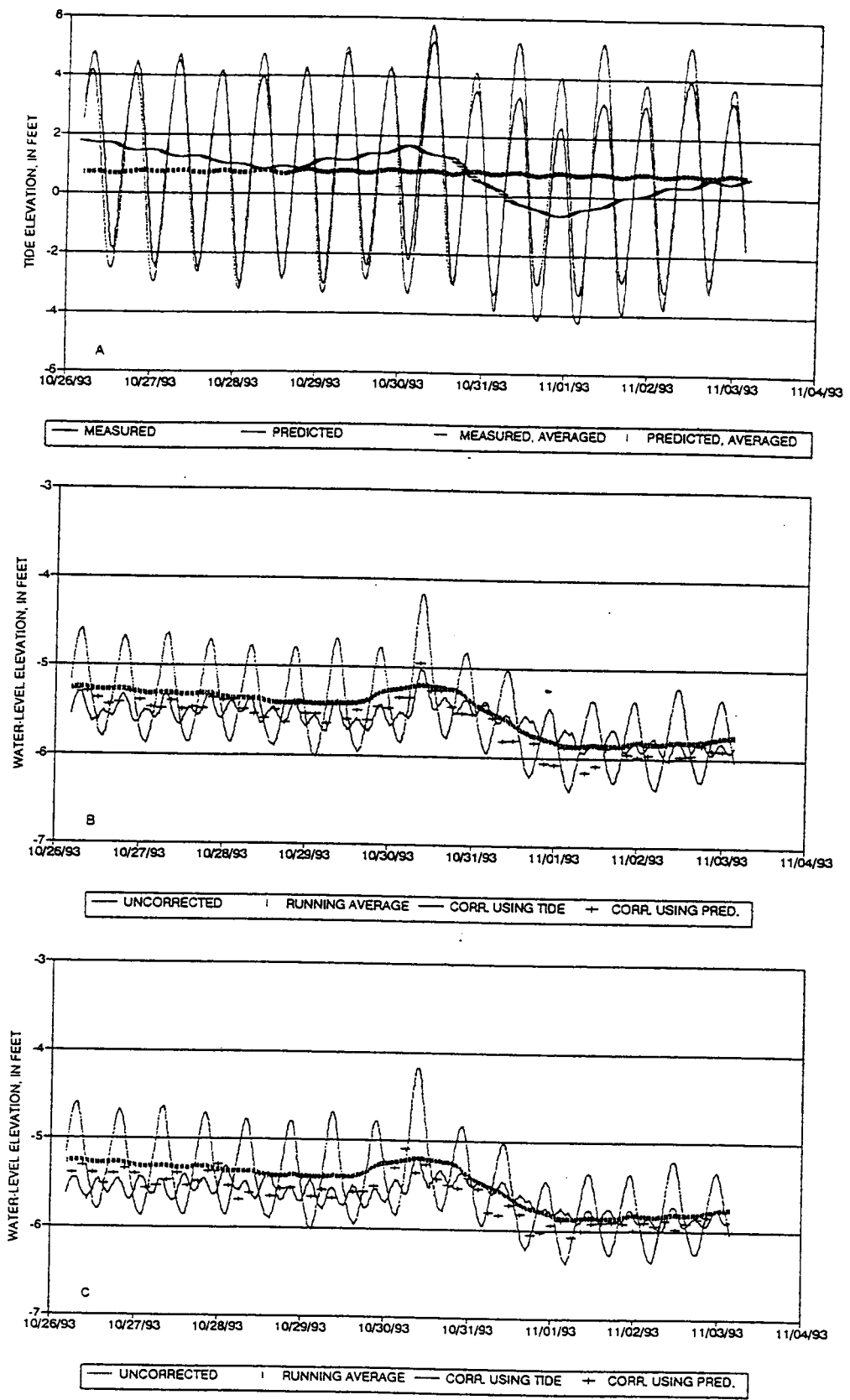


Figure 7. Comparison of tide-correction methods, Port Royal, weather disturbance. A, Water levels in Battery Creek, measured at 12-minute intervals, February 4-7, 1992. B, Uncorrected and corrected water levels, upper Floridan well 27II-128, at 1-hour intervals.



**Figure 8. Comparison of tide-correction methods, south of the Colleton River, weather disturbance.**  
**A, Water levels in the Colleton River, measured at 6-minute intervals. B, Uncorrected and corrected water levels, upper Floridan well 28JJ-p6, at 6-minute intervals; efficiency 0.15. C, Uncorrected and corrected water levels, upper Floridan well 28JJ-p6, at 6-minute intervals; efficiency varied.**

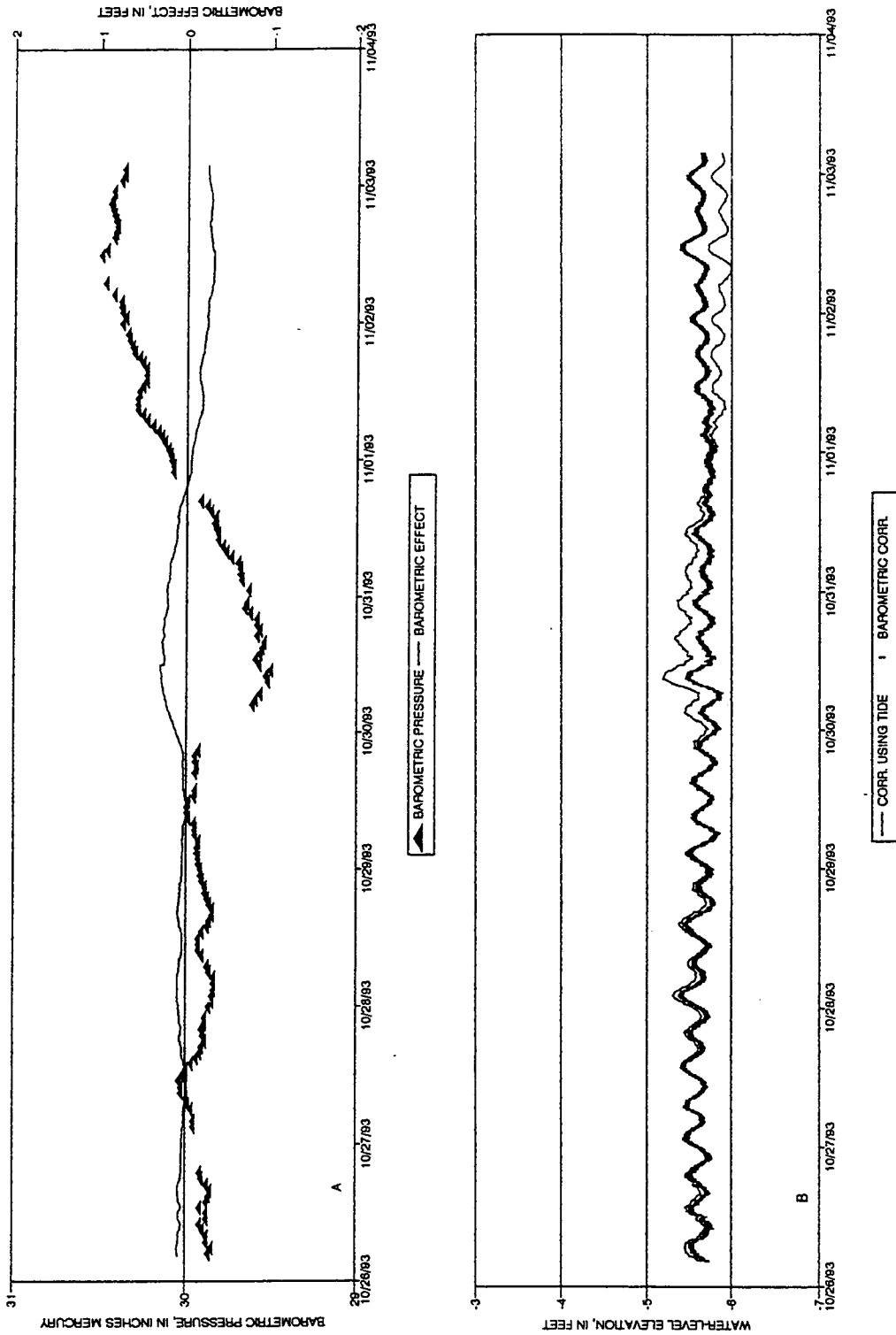


Figure 9. Barometric correction of tidally corrected water levels. A, Atmospheric pressure measured hourly at the Beaufort Marine Corps Air Station, Oct. 26 - Nov. 3, 1993, and calculated barometric effect, B, Corrected water levels, upper Floridan well 28JJ-p6, with and without barometric correction.

levels were lower, on Oct 31-Nov. 1, they are low. All three methods fail to correct out the unusual high associated with the spring tide of Oct. 30

The undulations remaining in the correction calculated from data (Fig. 8B) result from distortion of the tidal wave form in the aquifer, as in Fig. 3. Over the whole record, the high tides are undercorrected and the low tides are overcorrected, so that each forms a high relative to lows at midtides. For the period Oct. 26-30, the undercorrection of highs is more extreme; for Oct. 31-Nov. 2, the overcorrection of lows is excessive. This suggests that the tidal efficiency for this well varies over the period shown. A second correction was prepared (Fig. 8C), in which the efficiency was reduced in steps from 0.18 through 0.13 through time. This resulted in a better balance between corrections of highs and lows and reduced the size of fluctuations in the corrected water levels.

The anomalous high water levels of Oct. 30-31 remain in all three sets of corrections. None of the methods discussed correct for barometric effects (Bear, 1972, p. 211-213), which were described for upper Floridan wells in this area by Siple (1960, p. 31, Fig. 32) and by Counts and Donsky (1963). As mentioned above, a major storm moved through the area on Oct. 30. A nearly continuous series of hourly atmospheric pressure readings from the Beaufort weather station is shown in Figure 9A. Water in an open well in a confined aquifer rises or falls in inverse relationship to barometric pressure. Counts and Donsky (1963) measured fluctuations in well 30JJ-g2, in the upper Floridan aquifer approximately 9 miles west of 28JJ-p6, averaging 0.35 ft with a barometric change of 0.56 inch of mercury, or 0.63 ft of water; this corresponds to a barometric efficiency of 0.56 (ratio of change in well's water level to atmospheric pressure change expressed in terms of water height) (Bear, 1972, p. 212). This relationship was used to calculate the expected water-level change in well 28JJ-p6 caused by changes in atmospheric pressure (extrapolated between measured pressures) over the period Oct. 26-Nov. 3 (Fig. 9A). This barometric effect was subtracted from the tide-corrected water levels to obtain levels corrected for both effects (Fig. 9B). The barometric effect appears adequate to explain the major irregularities remaining in the well-water level record after tide correction.

Few water-level measurements from the Floridan aquifer system have been corrected for barometric effects. Hayes (1979, p. 44), although recognizing that the barometric effect could be as great as a foot, did not attempt to correct for it. For most water-level measurements, adequate local barometric data have not been available. This example indicates that atmospheric pressure should be recorded

during aquifer tests at Floridan wells, and a barometric correction applied if significant pressure changes occur (Fetter, 1988, p. 206-207; Kruseman and de Ridder, 1983, p. 180-181).

## CONCLUSIONS

Under the best conditions observed, when measured tides were close to those predicted, and tides were predicted to average near annual mean tide level, the three methods described gave consistent results, and each gave corrected levels that varied 23 percent or less of the observed water-level variation in the well. That is, the application of any of these methods reduced the error in a water-level measurement by 77 percent or more. For a well with an efficiency of 0.35 and for a tidal range of 10 ft this represents a reduction of range from 3.5 ft to 0.8 feet. More typically, for an efficiency of 0.20 and a tidal range of 7 ft, the measured range of 1.4 ft in the well would be reduced by tide correction to 0.3 ft. The maximum departure from static level should therefore be 0.4 ft in extreme cases, and more commonly it should be less than 0.2 ft.

As average predicted tidal level departed from annual mean tide level, and as weather-related water-level changes were superimposed on tidal levels, corrections calculated from measured tides remained most consistent and appeared to remove most or all of the effects of these additional disturbances. Systematic errors appeared in those corrections calculated by averaging or from predictions; these methods gave corrected water levels higher or lower than static level, depending on whether estuarine water levels averaged higher or lower than normal. Averaging failed to remove the effects of changes from either cause; it did, however, produce a constant level of error rather than extremes. Calculations from predictions failed to remove the effects of weather-related disturbances. In the worst case studied (Fig. 7, Feb. 6), the poorest corrections departed from static levels by approximately 40 percent of the measured range, and the average departed by approximately 25 percent, still a significant improvement in accuracy over uncorrected measurements.

These examples demonstrate that tidal correction of water-level measurements in tidally affected wells will improve the accuracy of reported water-level elevations, and therefore of potentiometric maps, calculated gradients, and calculated ground-water flow rates. Corrections should be made to measured tide data if possible. If predicted tides are to be used, measurement of well-water levels should be avoided during periods when estuarine water levels are affected by strong winds. In interpreting hydrographs plotted from instrumental data corrected by averaging,

both departures of average predicted tide level from mean tide level and weather-related water level changes should be taken into account. Barometric effects, although less

significant than tidal effects, may be great enough to require correction.

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## APPENDIX

### SPREADSHEET FOR CORRECTION OF WELL-WATER LEVELS, USING PREDICTED TIDES

Table 1 is a sample spreadsheet, and Table 2 gives the corresponding cell-by-cell calculations and instructions. The spreadsheet was written for Quattro or Quattro Pro (Borland International, Inc.), but it can be modified easily for other spreadsheets. The spreadsheet consists of four parts:

1) Lines 3-6 record information about the water-level measurement and calculate the time of day of the measurement in minutes, corrected for the well lag and for daylight-saving time if necessary. Note that in this spreadsheet the water level is entered as an elevation above mean sea level, not depth below land surface.

2) Lines 7-11 record predictions of high and low tide for one of the major tide stations for which detailed predictions are published (Tide Tables, Table 1) and recalculate times of day in minutes. The predictions chosen are for the high and low tide (in order of time) immediately preceding and following the water-level measurement; they bound a tidal half-cycle with an average length of 6 hours and 13 minutes.

3) Lines 12-17 calculate tidal levels and times for the nearest local tidal station. For these stations, listed in Table 2 of the Tide Tables, correction factors are provided for use in calculating levels and times from those at the reference station.

4) Lines 18-20 calculate the corrected water level from the measured level and local tide predictions. The duration is the length of the tidal half-cycle in minutes. The elapsed time is the corrected measurement time relative to the tidal half-cycle in minutes. It should always be examined. If it is negative or exceeds the duration, the corrected measurement time is not within the tidal half-cycle selected. This will not prevent the spreadsheet from calculating a tide correction; however, where tides are unequal, such a correction will be less accurate than one based on the half-cycle within which the measurement was taken. The user should return to line 7 and select the next earlier or later half-cycle from the tide table. This in turn requires changing the correction factors in lines 12-17.

This version of the spreadsheet was written for the purpose of making a single tide correction and printing a record of the data and calculation. The entire spreadsheet can be copied as a block to make additional corrections. Alternatively, the spreadsheet can be rewritten in more compact form, with constant factors on the first line and each correction on a single line; the exact format can be varied by the user to fit a particular application (for example making corrections to a number of measurements of the same well within a short time, or making corrections to measurements of a number of wells on the same date). To use the spreadsheet to interpolate a tide level between predicted levels, set lag to 0 and efficiency to 1 and use the calculated tide correction rather than the corrected level.

Table 1

SPREADSHEET TIDECORR (WRITTEN FOR QUATTRO OR QUATTRO PRO)

	A	B	C	D	E	F	G	H
1	TIDE CORRECTION FOR WATER-LEVEL MEASUREMENT IN WELL							
2	CONSTANTS	DST	60	PI	3.1416			
3	DATA	WELL	BFT 750	DATE	7/30/90	M. ELEV.	-20.77	
4			hour	MIN		MIN		
5	TIME		14	42	LAG	38	EFF	0.19
6	COMPARISON TIME 784 MIN							
7	REFERENCE STATION: ENTER TIDES PRECEDING AND FOLLOWING MEASUREMENT							
8	SAVANNAH R. ENT.	hour	7	50	hour	MIN		
9	TIMES				14	26		
10	LEVELS		0.6		6.7		FT ABOVE LWL	
11	TIMES (RECALC. IN MIN)			470		866	MIN	
12	LOCAL STATION:	BRADDOCK PT			TIME, LEVEL CORRECTIONS			
13	(MAKE SURE HIGH AND LOW TIDES ARE IN SAME ORDER AS FOR REF. STATION)							
14	TIME		0	-2	0	5	MTL	3.6
15	LEVEL RATIO		0.95		0.98			
16	LOCAL TIDE TIMES			468		871	MIN	
17	LOCAL TIDE HEIGHTS		0.57		6.566		FT ABOVE LWL	
18						MTR		3.568
19	DURATION	403	ELAPSED T	316	ANGLE	2.46	MTC	-0.032
20	AMP	-2.998	TIDE H	2.33	TIDE CORR	0.44	CORR ELEV	-21.21

Table 2

Explanation of cell contents for spreadsheet TIDECORR

CELL	CONTENTS	EXPLANATION AND INSTRUCTIONS FOR USE
A1:	'TIDE CORRECTION FOR WATER-LEVEL MEASUREMENT IN WELL	
A2:	'CONSTANTS	
B2:	' DST	When setting up spreadsheet, enter labels and formulas in cells listed (some cells empty).
C2:	' 60	When using, change only cells marked *ENTER... (contents given are examples only).
D2:	' PI	Explanations and definitions enclosed by [ ] .
E2:	' 3.1416	
A3:	'DATA	
B3:	'WELL	
C3:	'BFT 750	[Label: Daylight Savings Time correction] *ENTER 60 if measurement time on DST, 0 if not
D3:	'DATE	*ENTER Well I.D., any format desired
E3:	'7/30/90	*ENTER date of measurement, any format
F3:	'M. ELEV.	[Measured water level converted to elevation]
G3:	'-20.77	*ENTER water-level measurement in ft above MSL
C4:	' HOUR	
D4:	' MIN	
F4:	' MIN	
B5:	'TIME	[Time of measurement, in 24-hour format:]
C5:	' 14	*ENTER hour portion of time of measurement
D5:	' 42	*ENTER minute portion of time of measurement
E5:	' LAG	
F5:	' 38	*ENTER tidal lag for well, in minutes
G5:	' EFF	
H5:	' 0.19	*ENTER tidal efficiency for well
B6:	'COMPARISON TIME	
D6:	'+C5*60+D5-F5-C2	Formula: measurement time -lag (-60 min DST if needed)
E6:	'MIN	
A7:	'REFERENCE STATION: ENTER TIDES PRECEDING AND FOLLOWING MEASUREMENT	



A8: 'SAVANNAH R. ENT. \*ENTER name of reference station in Tide Table 1  
C8: ' HOUR [Information below is obtained from Tide Tables,  
D8: ' MIN Table 1, for reference station; for date and time of  
E8: ' HOUR water-level measurement, locate high or low tide  
F8: ' MIN preceding measurement, and that following it]  
B9: 'TIMES  
C9: 7 \*ENTER hour portion of time of preceding tide  
D9: 50 \*ENTER minute portion of time of preceding tide  
E9: 14 \*ENTER hour portion of time of following tide  
F9: 26 \*ENTER minute portion of time of following tide  
B10: 'LEVELS  
C10: 0.6 \*ENTER height of preceding tide, in ft  
E10: 6.7 \*ENTER height of following tide, in ft  
G10: 'FT ABOVE LWL [Tide heights are measured from mean lower low water]  
A11: 'TIMES (RECALC. IN MIN)  
D11: +C9\*60+D9 Formula: time of preceding tide, recalculated in minutes  
F11: +E9\*60+F9 Formula: time of following tide, recalculated in minutes  
G11: 'MIN  
A12: 'LOCAL STATION: [corrections below obtained from Tide Tables Table 2]  
C12: 'BRADDOCK PT \*ENTER name of nearest local tide station  
E12: 'TIME, LEVEL CORRECTIONS  
A13: ' (MAKE SURE HIGH AND LOW TIDES ARE IN SAME ORDER AS FOR REF. STATION)  
A14: ' TIME  
C14: 0 \*ENTER hour portion of time diff. for preceding tide  
D14: -2 \*ENTER minute portion of time diff. for preceding tide  
E14: 0 \*ENTER hour portion of time diff. for following tide  
F14: 5 \*ENTER minute portion of time diff. for following tide  
G14: ' MTL  
H14: 3.6 \*ENTER mean tide level for local station  
A15: ' LEVEL RATIO  
C15: 0.95 \*ENTER correction ratio for height of preceding tide  
E15: 0.98 \*ENTER correction ratio for height of following tide  
A16: 'LOCAL TIDE TIMES  
D16: (C14\*60+D14)+D11 Formula: time of local preceding tide, in minutes  
F16: (E14\*60+F14)+F11 Formula: time of local following tide, in minutes  
G16: 'MIN  
A17: 'LOCAL TIDE HEIGHTS:  
C17: +C10\*C15 Formula: height of local preceding tide  
E17: +E10\*E15 Formula: height of local following tide

G17: 'FT ABOVE LWL  
G18: ' MTR  
H18: (E17-C17)/2+C17  
A19: 'DURATION  
B19: +F16-D16  
C19: 'ELAPSED T  
D19: +D6-D16  
E19: 'ANGLE  
F19: +D19\*E2/B19  
G19: ' MTC  
  
H19: +H18-H14  
A20: 'AMP  
B20: +C17-H18  
C20: 'TIDE H  
  
D20: +B20\*@COS(F19)  
E20: 'TIDE CORR  
F20: +H5\*(D20+H19)  
G20: 'CORR ELEV  
H20: (F2) +G3-F20

[Tide heights are measured from mean lower low water]  
[Middle of range of this tidal half-cycle]  
Formula: height of middle of tidal half-cycle  
[Length of tidal half-cycle, in minutes]  
Formula: time between preceding and following tides  
[Time of water-level measurement relative to tide cycle]  
Formula: Minutes from preceding tide to comparison time  
[elapsed time/duration expressed as angle in radians]  
Formula: elapsed time x pi/duration  
[Difference between middle of this tidal half-cycle  
and mean tide level, in feet]  
Formula: MTR-MTL  
[Amplitude of tidal half-cycle]  
Formula: Height of following tide - MTR  
[Calculated height of tide at time corresponding to  
that of water-level measurement]  
Formula: Amplitude X COS(ANGLE)  
[Tide correction to be subtracted from water elevation]  
Formula: EFF X (TIDE H + MTC)  
[Corrected water-level elevation relative to msl]  
Formula: Measured water-level elevation - TIDE CORR  
[(F2): numeric format set to 2 decimal places; not part  
of formula]

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# RADIOCARBON ISOTOPES IN THE INTERPRETATION OF GROUND-WATER FLOW REGIMES AT HILTON HEAD ISLAND, SOUTH CAROLINA

by

Joffre E. Castro

## ABSTRACT

Hilton Head Island, a barrier island at the south end of South Carolina, obtains virtually all of its water supply from the upper Floridan aquifer. Changes in the ground-water gradient of this aquifer during the past 50 years, owing to intensive pumping in Georgia, have reversed the flow direction underneath the island. Moreover, lower hydraulic heads and increased rates of vertical infiltration have produced a serious threat of contamination of the aquifer with saltwater. To better understand and quantify the ground water regime in Hilton Head Island, a series of hydrologic and geologic studies has been made.

This study uses radiocarbon and chemical data to define the ground water flow regime of the island following earlier work by Back and others (1970). The radiocarbon analysis demonstrates that the island's ground is a mixture of water of various chemical compositions and ages. The ground water at the north end of the island is characterized by freshwater, in the middle by a mixture of freshwater and old native ground water, and at the south end by old native ground water and probably leakage from overlying aquifers. Large fractions of modern seawater were identified north of the island and underneath Port Royal Sound. Finally, stable and isotopic carbon data can provide valuable information about the ground water system when used in combination with chemical data. Geochemical and isotopic models, thus, can be used to correct  $^{14}\text{C}$  activities and compute meaningful ground water ages.

## INTRODUCTION

Hilton Head Island is in Beaufort County, S.C., (Fig. 10). For most of the past century, its inhabitants have relied solely on ground water for water supply. Virtually all ground water is obtained from the upper part of the Floridan aquifer, a Tertiary limestone formation. This limestone is the upper unit of the prolific Floridan aquifer system that underlies Florida and parts of Alabama, Georgia, and South Carolina.

Underlying Hilton Head Island, which is a barrier island, is a series of unconsolidated to poorly consolidated sediments ranging in age from Late Cretaceous (70+ million years) to Holocene (7,000+ years)(Colquhoun and others, 1983). The upper Floridan aquifer, the most important aquifer in this system, is a permeable limestone confined at the top by the clayey Hawthorn Formation of Miocene age and at the bottom by poorly confining limy units overlying the lower Floridan aquifer (Table 1). The upper Floridan aquifer lies about 100 ft (feet) below the land surface and has an approximate thickness of 80 ft (Hughes and others 1989).

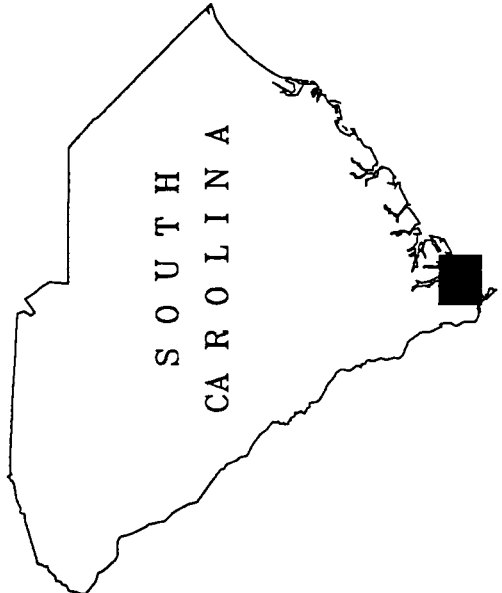
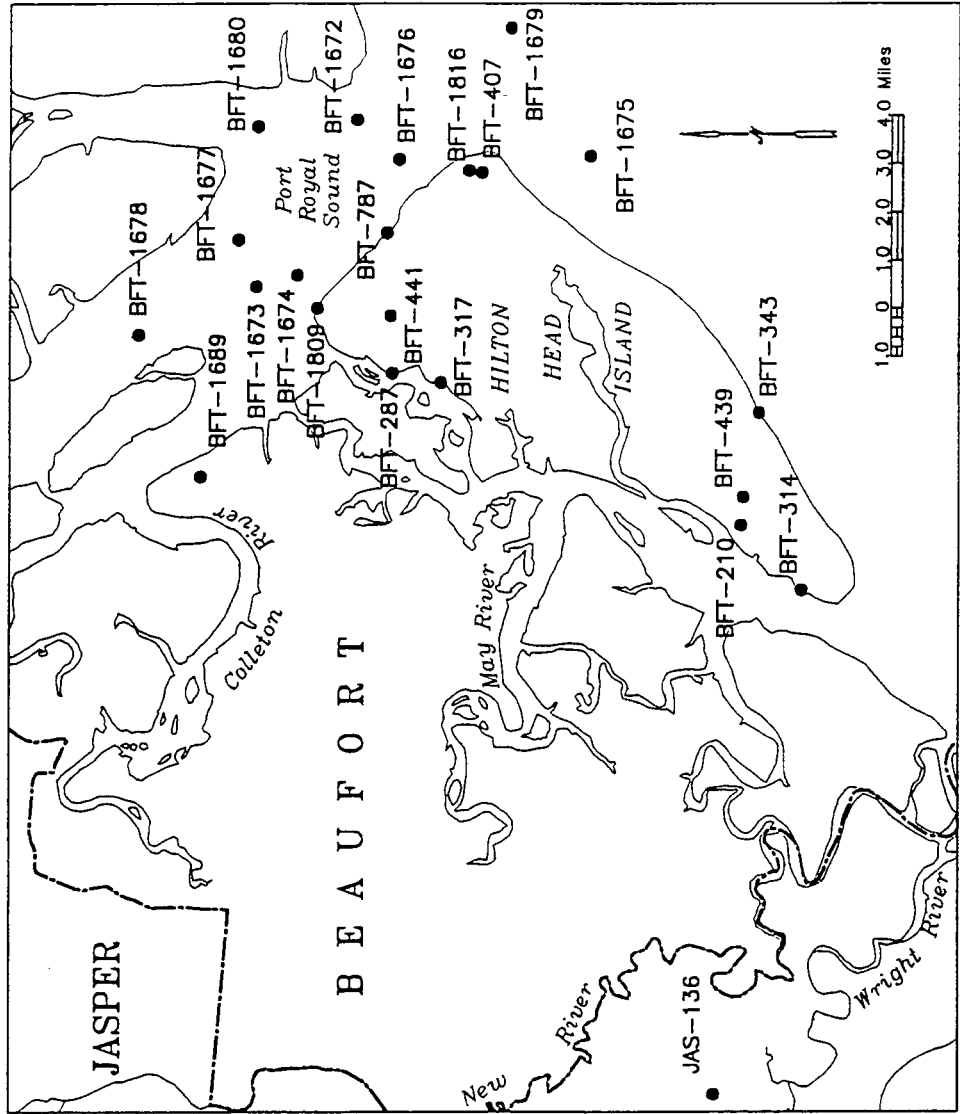
## HYDROLOGY

Prior to intensive economic development of Savannah and surrounding Chatham County, Ga., and the Beaufort area of South Carolina, which began in the late 1800's, ground water flowed into Hilton Head

Island from the west (Fig. 11)(Bush and Johnston, 1988). Recharge came mostly from rainwater that entered the ground-water system 150-200 miles northwest of Hilton Head Island, in east-central Georgia. Ground water discharged into Port Royal Sound, north of Hilton Head Island.

With development, the potentiometric surface of the region (Fig. 12) has undergone significant changes (Burt, 1989). Ground-water flow in Hilton Head Island, for example, has reversed direction. It is now flowing in a southwesterly direction, toward Savannah, where a pronounced cone of depression has developed because of intensive pumping. This feature controls not only the local ground-water regime in Savannah but also the regional flow in southern South Carolina and eastern Georgia. The flow reversal at Hilton Head Island has increased the risk of ground-water contamination of the island. Lower hydraulic heads and the change in direction of the ground-water flow have caused the saltwater-freshwater interface to move closer inland, thus jeopardizing existing public supply wells. Lower hydraulic heads, moreover, have resulted in larger rates of vertical infiltration (local recharge), which when coupled with the common practice of waste water spraying for golf course irrigation, presents a threat to the quality of drinking water on the island (Stone and others, 1986).

These and other problems have prompted a keen interest from government and the scientific community



EXPLANATION

● BFT-343 Well location and number.

Figure 10. Locations of selected wells on Hilton Head Island and its vicinity, South Carolina (modified from Burt, 1989).

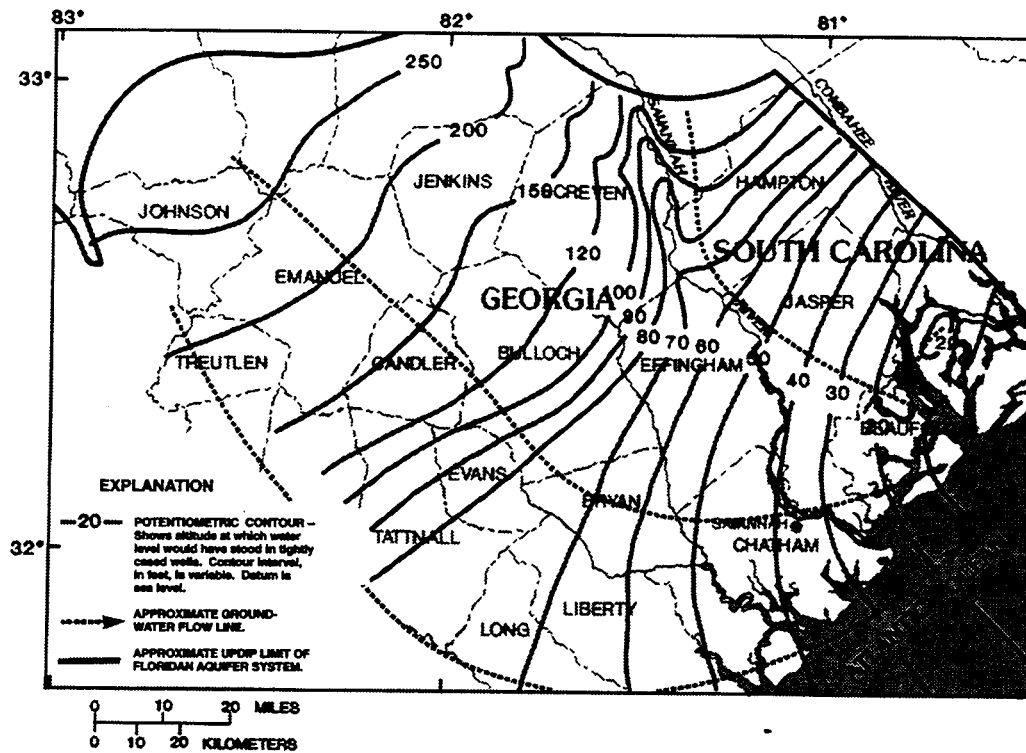


Figure 11. Potentiometric surface (estimated) and ground-water flow directions in the Upper Floridan aquifer prior to ground-water withdrawals, 1880 (from Bush and Johnston, 1988).

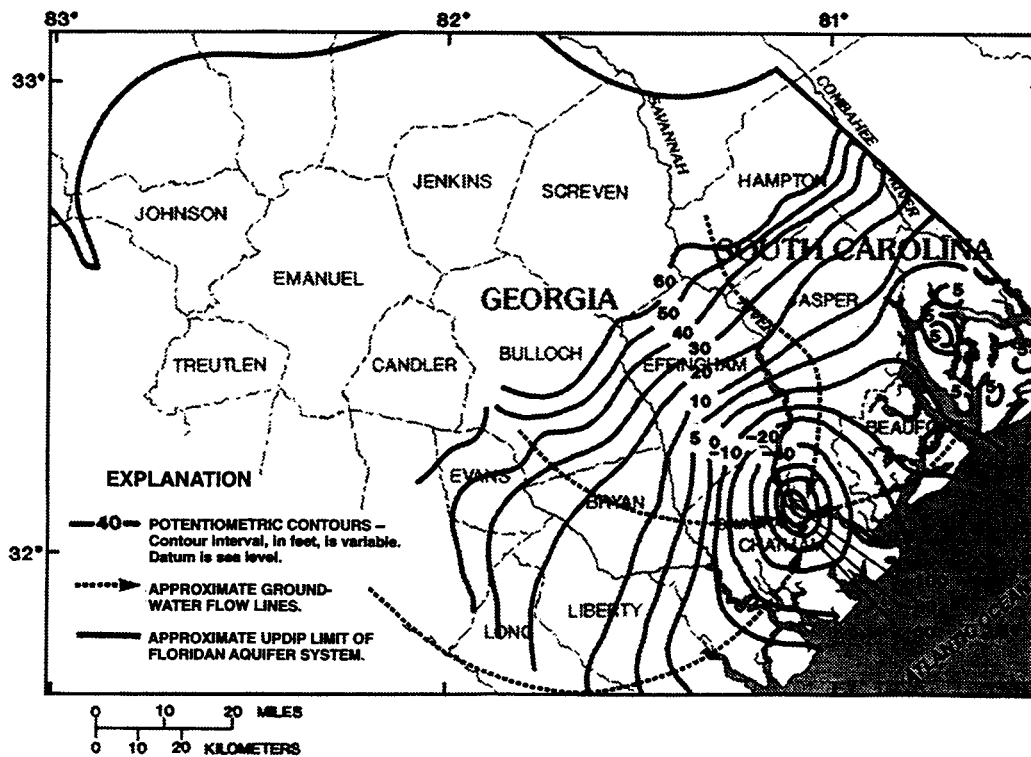


Figure 12. Potentiometric surface and ground-water flow directions in the Upper Floridan aquifer, 1984 (from Burt, 1993).

because of implications to the economic development of the island and the quality of life of the islanders. Several studies on the island and in the region have been aimed to promote a better understanding of the hydrogeology and geochemistry of the ground-water system.

### RADIOCARBON DATA

Back and others (1970) used radioisotopes of carbon and hydrogen to date ground water and to derive a reasonable interpretation of the ground-water flow system of Hilton Head Island. Burt (1989) used chemical and isotopic information to characterize the evolution of the ground water. Landmeyer and Stone (1995) used radiocarbon data to identify areas of local recharge to the upper Floridan aquifer.

A closer inspection of the ages computed from <sup>14</sup>C data in the earlier studies shows some inconsistencies. Inland ground-water ages (upgradient) appear to be older than island ground-water ages (downgradient). Furthermore, travel times computed from the <sup>14</sup>C ages did not always agree with those computed from Darcy's Law. In the following paragraphs, an alternate explanation is given to clarify this apparent contradiction.

The data for this study were obtained from a U.S. Geological Survey geochemical investigation in southern South Carolina and southeastern Georgia (Burt, 1989). The data are summarized in Table 2, in which WELL refers to the well's county identification number; Producing interval lists the top and bottom depths, in feet, of the screened portion of the aquifer or of the unscreened open interval; Cl<sup>-</sup> is milligrams of chloride per liter of water (original data were provided in millimoles per kilogram); δ<sup>13</sup>C (carbon-13) is given in per mil (PDB standard); and <sup>14</sup>C (carbon-14) is given in units of percent modern carbon (pmc).

Figure 13 shows the areal distribution of <sup>14</sup>C and the locations of wells. According to these data, the north end of Hilton Head Island presently is an area of recharge—higher <sup>14</sup>C activity compared to lower <sup>14</sup>C activity at the south end of the island. This is in accordance with the hydrologic data shown in the upper Floridan potentiometric map (Fig 12).

### METHODOLOGY

The <sup>14</sup>C activities were used in computing a ground-water age, which represents the time of travel between two points. If one of the points is near a recharge area of the aquifer, then the ground-water age would represent the approximate minimum time since the water was last exposed to the atmosphere, which is the time that the water has been in the ground-water system. As ground water flows away from the recharge area (source of

Table 1. Geologic section of the Floridan aquifer system (modified from Burt, 1989, Fig. 5)

Lithology	Feet below sea level	Age	Formation	Aquifer	
	0	Holocene	Undif. sand	Surficial	
		Pleistocene	Pamlico		
			Talbot		
			Miocene	Hawthorn	Hawthorn
	200	Eocene	Ocala Limestone	Upper Unit	Floridan aquifer system
				Lower Unit	
	400	Eocene	Santee Limestone	Lower Unit	Lower Floridan aquifer
	600	Eocene	Santee Limestone	Lower Unit	Lower Floridan aquifer
	800	Eocene	Santee Limestone	Lower Unit	Lower Floridan aquifer

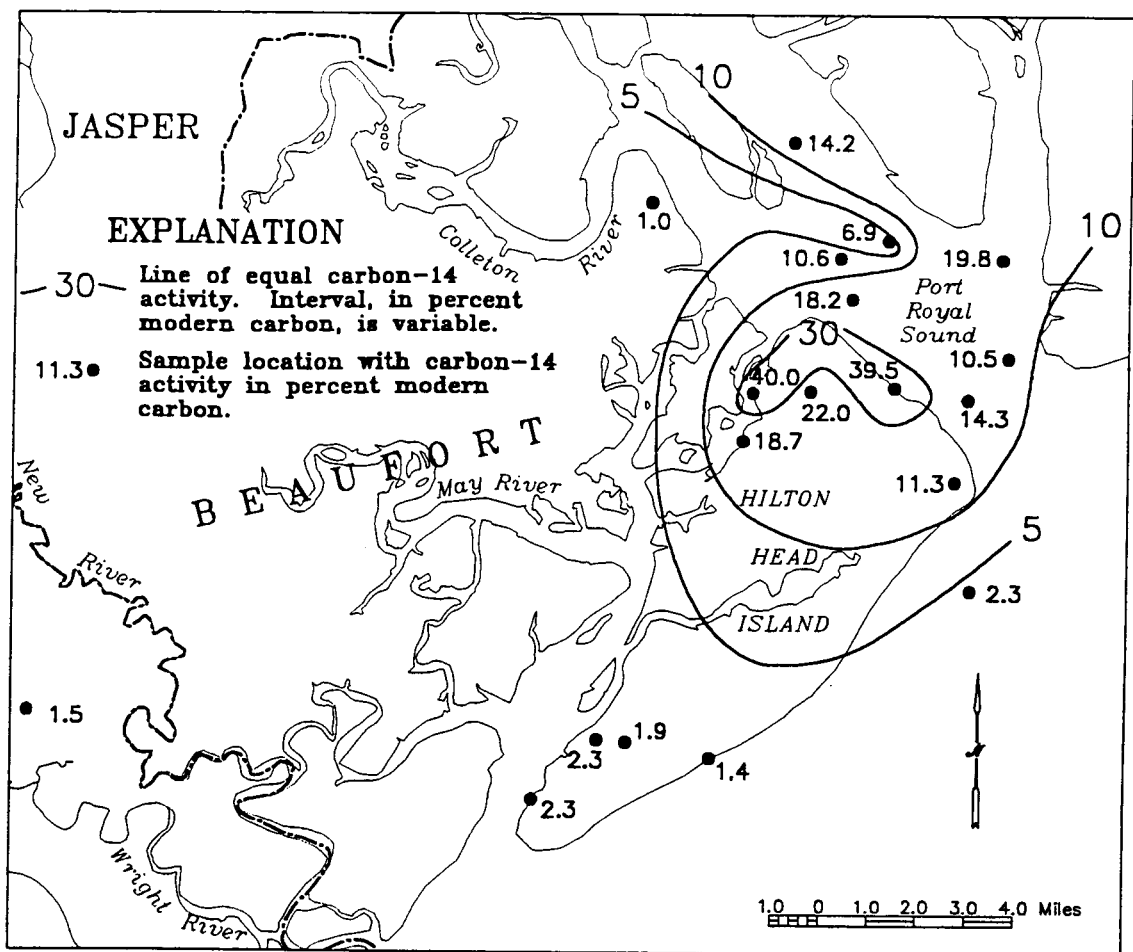


Figure 13. Distribution of carbon-14 activities in ground-water in the upper Floridan aquifer (from Burt, 1989).

Table 2. Isotopic ground water analyses for Hilton Head Island (from Burt, 1989)

WELL	Producing interval (depth in feet)		Cl (mg/L)	$\delta^{13}\text{C}$ (mil)	$^{14}\text{C}$ (pmc)
	TOP	BOTTOM			
JAS-136	200	245	4	-3.0	<1.5
BFT-441	187	216	42	-9.5	22.0
BFT-787	172	185	63	-12.2	39.5
BFT-210	130	160	17	-1.8	2.3
BFT-439	182	214	29	-1.6	1.9
BFT-1672	95	107	259	-8.9	10.5
BFT-1672	174	211	9,730	-7.8	7.5
BFT-1674	170	174	11,900	-9.0	18.2
BFT-1675	186	212	160	-3.4	2.3
BFT-1676	97	110	60	-10.9	7.3
BFT-1676	178	182	735	-8.8	14.3
BFT-314	125	230	7	-9.0	2.3
BFT-287	90	195	30	-12.9	40.0
BFT-317	86	196	33	-8.7	18.7
BFT-343	124	200	34	-5.1	1.4
BFT-407	140	214	112	-13.7	11.3



modern CO<sub>2</sub>) the <sup>14</sup>C activity decreases by radioactive decay, thereby creating a geochronometer. Several geochemical processes may dilute the amount of <sup>14</sup>C in solution affecting the geochronometer.

Most important among the geochemical processes are fractionation and dilution. Carbon isotopes are fractionated by physical and chemical reactions, such as photosynthesis and carbonate precipitation. During photosynthesis, the carbon in plants becomes depleted in <sup>13</sup>C and <sup>14</sup>C with respect to <sup>12</sup>C. During carbonate precipitation, carbon in the carbonates becomes enriched in <sup>13</sup>C and <sup>14</sup>C with respect to <sup>12</sup>C. Thus, the isotopic composition of the carbon varies according to the source: the carbon from marine plants has a δ<sup>13</sup>C<sup>1</sup> content of -27‰, and the carbon from calcite has a δ<sup>13</sup>C of 0‰. Furthermore, fractionation of <sup>14</sup>C is assumed to be twice that of <sup>13</sup>C (Faure, 1986). Dilution occurs when carbon of different isotopic composition is added to or removed from the ground water, for example calcite dissolution adds carbon devoid of <sup>14</sup>C (old carbon). If the <sup>14</sup>C activities are not corrected for fractionation or dilution the <sup>14</sup>C ages are overestimated—they are too old. The <sup>14</sup>C-age was computed by

$$t = -\frac{5,730}{\ln 2} \cdot \ln \left( \frac{A_c}{A_o} \cdot \frac{1}{f} \right) \quad (1)$$

$$f = \frac{\delta_s^{13} - \delta_c^{13}}{\delta_g^{13} - \delta_c^{13}} \quad (2)$$

The correction (f) for dilution was taken from Mook (1980), where:

A is <sup>14</sup>C activity, pmc (percent modern carbon).

δ<sup>13</sup>C is carbon-13 content, ‰ (per mil).

ln is the natural logarithm.

f is the correction factor computed from δ<sup>13</sup>C data.

s, c, g are subscripts referring to the δ<sup>13</sup>C for sample, carbonate, and biogenic sources.

t is <sup>14</sup>C age in years since 1950.

5,730 is the <sup>14</sup>C half-life.

Alternate methods for correcting the <sup>14</sup>C activities are: the DIC (dissolved inorganic carbon) concentration (Mook, 1980); and the "dilution factor Q" (Freeze and Cherry, 1979). Although all <sup>14</sup>C activity correction methods are expected to provide similar <sup>14</sup>C ages, it was observed that they did not. For this study, the δ<sup>13</sup>C-correction method was selected, as it provided more reasonable results, although the <sup>14</sup>C activity appeared to be overcorrected when the δ<sup>13</sup>C content of the water was isotopically heavier (closer to zero values), which normally resulted in ages younger than those computed from the other methods. (Table 3).

<sup>1</sup> δ<sup>13</sup>C = [(<sup>13</sup>C/<sup>12</sup>C) - 1] × 10<sup>3</sup> ‰

In the upper Floridan aquifer at Hilton Head Island, corrections for dilution are larger than for fractionation, and the latter was ignored except when a geochemical model (NETPATH) was used. Corrections for fractionation can be ignored because (a) fractionation factors for the carbonate-carbon dioxide system are small (Faure, 1986), and (b) the most important chemical process is calcite dissolution, which is not affected by fractionation.

The small discrepancy between the <sup>14</sup>C age and that of Back and others (1970) may be due to the different half-life used (Equation 1). Whereas Back used the "old" value (5,568 years), this writer used the "new" value (5,730 years) (Faure, 1986). Differences between the <sup>14</sup>C ages from the different methods are larger and more difficult to explain. These methods, including the δ<sup>13</sup>C-correction method, computed <sup>14</sup>C ages that are inconsistent. For example, the ground water at the south end of the island is consistently older than that at the north end, but it is inconsistently younger than that from inland areas (west and upgradient). The ground water age of well BFT-210 is younger than that of well JAS-136, which is at least 10 miles inland and upgradient.

Although the ground water at the north end of the island has the youngest <sup>14</sup>C age, it is older than would be expected. From Darcy's Law, assuming a transmissivity of 40,000 ft<sup>2</sup>/day, a porosity of 0.2, and a natural gradient of 1.4 ft/mile, a ground water travel time of 22 ft/yr is calculated. This means that the ground water could travel the length of the island (13 miles) in approximately 3,000 years. Therefore, <sup>14</sup>C ages between 3,000 and 22,000 years (oldest age of ground water at the south end) are suspect.

To understand this apparent inconsistency of the <sup>14</sup>C ages, the geochemical system was studied with a mass-balance model (NETPATH) developed by the U.S. Geological Survey (Plummer and others, 1991). The purpose of the modeling was to identify types of water of different chemical and isotopic composition that may have mixed with the native ground water and thus altered the <sup>14</sup>C activity of the resulting water. These other sources may have been freshwater from local recharge, which would be characterized by low Cl<sup>-</sup> concentration and high <sup>14</sup>C activity; diluted seawater, characterized by high Cl<sup>-</sup> concentration and either high <sup>14</sup>C (new) or low <sup>14</sup>C activity (old); or leakage from overlying aquifers, characterized by low Cl<sup>-</sup> concentration and moderate <sup>14</sup>C activity. Although the latter type of water was identified as a potential source of recharge for the upper Floridan aquifer, it was not used in the analysis because chemical and isotopic data were not available.

In the geochemical modeling of the ground water system, the chemical data were used first to obtain a calibrated mass-balance model, which was then veri-

fied by using the isotopic information. Once a successful model was obtained—one that satisfied both the chemical and isotopic constraints—the corrected  $\delta^{13}\text{C}$  content and  $^{14}\text{C}$  activity for the ground water were estimated and a  $^{14}\text{C}$  age computed. The modeling of the ground water system helped to understand the chemical and isotopic evolution of the ground water. Potential sources for the ground water and flow regimes were tested, mixing ratios estimated, and  $^{14}\text{C}$  activities computed, thus allowing for a more accurate determination of the  $^{14}\text{C}$  age and for a better interpretation of the hydrologic system.

## DISCUSSION

Four wells were selected for this analysis, they are: BFT-287 and BFT-317 at the north end of the island; BFT-314 at the south end; and BFT-1675 to the east and 1½ miles offshore. Geochemical modeling of the ground water evolution was simplified by considering the three most important chemical reactions (Burt, 1989); organic-matter oxidation, calcite incongruent dissolution, and sodium-calcium exchange. The organic matter oxidation was not simulated, instead a  $\text{CO}_2$  phase was assumed to be present (Castro, 1995), which was assumed to be biogenic (Chapelle and others, 1988). In modeling the evolution of the ground water, several scenarios were hypothesized that reflected different hydrogeologic conditions that may have affected the chemical composition of the ground water. Some scenarios considered the evolution of the ground water from the time it entered the aquifer as rainwater, reacted with mineral phases in the unsaturated zone under a constant supply of  $\text{CO}_2$  (open system), recharged the water table, and later reacted with minerals in the aquifer (closed system) and mixed with old ground water. Other scenarios modeled relative chemical and isotopic changes of the ground water, for example, changes that might occur between two points (wells) along a flow line. The latter scenario was combined with the input of young freshwater (rainwater) to simulate local recharge.

**BFT-314.** Ground water from this well was simulated as a mixture of rainwater that entered the aquifer at the recharge area (east-central Georgia) and old seawater (seawater trapped in the pore spaces of the aquifer during the last rising of the sea). Ground water ages and mixing results computed by the models are presented in Table 4.

On the basis of this chemical and isotopic evolution (Table 4) the model computed a  $^{14}\text{C}$  age of 21,000 years, which is in good agreement with the  $^{14}\text{C}$  age computed from equation 1 and Back's estimates. Although the ground water at BFT-314 is older than that at JAS-136, the  $^{14}\text{C}$  activity of the water at these wells seems

Table 3. Carbon-14 ages for water in the upper Floridan aquifer

WELL	$^{14}\text{C}$ AGE <sup>1</sup>	NETPATH	Back and Others (1970)
JAS-136	17,200	15,000	
BFT-441	4,500		
BFT-787	1,700		
BFT-210	9,400		
BFT-439	10,000		
BFT-1672	10,000		
BFT-1672	11,800		
BFT-1674	5,600		
BFT-1675	14,700	3,100	
BFT-1676	14,800		
BFT-1676	7,500		
BFT-314	22,700	21,000	22,500
BFT-287	2,100	<100	2,100
BFT-317	5,100	2,100	5,000
BFT-343	22,200		22,400
BFT-407	13,000		12,500

<sup>1</sup>,  $^{14}\text{C}$  age corrected by f

NETPATH,  $^{14}\text{C}$  ages from geochemical model

oddly inconsistent. The  $^{14}\text{C}$  activity of the ground water at well BFT-314 (2.3) and at some of the surrounding wells is greater than at well JAS-136 (1.5), suggesting that another solution may be possible. The increase in the  $^{14}\text{C}$  activity may be explained by leakage from overlying units. This ground water could be freshwater that is locally recharging the shallow aquifers. By comparison with the upper Floridan ground water, this water would have greater  $\text{Cl}^-$  concentration and  $^{14}\text{C}$  activity.

Most likely, this water is precipitation that was slightly modified by sea spray (increased  $\text{Cl}^-$ ) and that has been recharging the water table in recent times. No data were available on the chemical and isotopic composition of the shallow ground water; therefore, this scenario was not tested.

In general, ground water at the south end of the island appears to have mixed not with brackish water (old seawater), but with freshwater from overlying aquifers (leakage).

**BFT-1675.** The ground-water composition at this well was simulated by postulating various scenarios. First, it was considered that there may not have been chemical changes in the ground water owing to the flow reversal. Thus, the water in BFT-1675 was considered to be made up of rainwater that entered the aquifer in the outcrop area (east-central Georgia) and old seawater. The model estimated a  $^{14}\text{C}$  age of 13,000 years, which is similar to the age of the ground water from wells north

of BFT-314 (Table 3). These wells are just upgradient from BFT-1675. The  $^{14}\text{C}$  age of 13,000 years for BFT-1675 is, however, younger than would be expected. According to the assumptions used in this scenario, the flow was from the west and the ground water age should have been older than 22,000 years. Thus, this scenario was deemed improbable.

A second scenario considered that BFT-1675 evolved from a mixture of ground water from well BFT-314 and seawater. No feasible results were obtained. The geochemical modeling suggested the need for a source of freshwater with modern carbon age. This would imply that local recharge to the aquifer from freshwater is feasible not only on the north end (Landmeyer and Stone, 1995) but also in the middle of the island. This fresh-water recharge could also help explain the younger ages—less than 22,000 years—observed at wells BFT-210, BFT-439, and BFT-343 and the higher tritium activity reported by Landmeyer and Stone (1995) at some the wells.

A third scenario considered ground water at BFT-1675 to be a mixture of ground water similar to that of well BFT-287, seawater, and freshwater (rainwater recharge). For this scenario the model estimated a  $^{14}\text{C}$  of 3,100 years. The implications of this assumption are that freshwater is recharging the upper Floridan aquifer and that the ground water has been flowing seaward for at least 3,100 years.

**BFT-287.** This well, of all wells studied, has the highest  $^{14}\text{C}$  activity (40 pmc). The ground water composition at this well was simulated as a mixture of rainwater and seawater. The  $^{14}\text{C}$  age indicates that this ground water is of modern age, less than 100 years old. This agrees well with the hydrologic data that show a recharge mound at the north end of the island.

**BFT-317.** This well is a mile south and just west of well BFT-287. Several scenarios were used to model the chemical composition of the ground water at this well. First, it was considered to be ground water that is a mixture of rainwater and seawater. The computed  $^{14}\text{C}$  age was 2,000 years, which appears to be too old, considering that the recharge areas may be only 1 mile to the north. A second scenario tested assumed that the ground water at well BFT-317 is a mixture of BFT-287, seawater, and rainwater. The calculated  $^{14}\text{C}$  age was 2,300 years, which is similar to the age computed in the previous scenario. Other scenarios were also tried but were unsuccessful. None of these models could achieve chemical mass balance. For example, the composition of the ground water at BFT-210 was used in combination with that of with seawater and rainwater. This scenario evaluated the chemical contribution of the ground water from the south end of the island in the makeup of the

water around the middle of the island. The failure of the model to achieve a solution might suggest that this contribution is minimal. Freshwater, by contrast, appears to be the dominant source of ground water. This freshwater may come from the overlying sediments (Hawthorn Formation) and from recharge areas at the north end of the island. In either case, the  $^{14}\text{C}$  activity must be less than that of freshwater (50 pmc) and greater than that of the ground water at the south end (2 pmc). Failure to include a water type of such chemical and isotopic composition in the simulations may explain the inability of the models to compute a more reasonable age (less than 2,000 years). Unfortunately, the necessary chemical and isotopic information for the Hawthorn Formation at relevant sites was not available at the time of this investigation.

Using the ground-water ages, the chloride concentrations, and following Back's original work, a fence diagram outlining the flow of the ground-water regime of Hilton Head Island (Fig. 14) was sketched.

## SUMMARY AND CONCLUSIONS

Geochemical modeling of the ground water system of Hilton Head Island suggests that there are at least four water types that mix and make up the present chemical composition of the water. Old native ground water is recharged about 150 miles west of Savannah (east-central Georgia) and is present throughout most of the south end of the island. This water is characterized by an age greater than 22,000 years and by less than 60 mg/L (milligrams per liter) chloride concentration. New freshwater, which comes from direct precipitation on the island, is characterized by an age less than 2,000 years and less than 60 mg/L chloride concentration. This type of water appears to predominate at the north end of the island (recharge) and has a strong influence on the composition of the ground water in the middle of the island and offshore. The other two types of water are both seawater: one is old seawater and the other is modern seawater. Because the percentage of seawater present in the ground water is significantly smaller than that of the other water types, the models were insensitive

Table 4. Geochemical-model results

WELL	PERCENT MIXING			$^{14}\text{C}$ AGE (YEARS)
	RAIN	SEA	BFT-287	
BFT-314	99.96	0.04		21,000
BFT-1675	63.50	0.70	35.80	3,100
BFT-287	99.85	0.15		<100
BFT-317	99.84	0.16		2,000

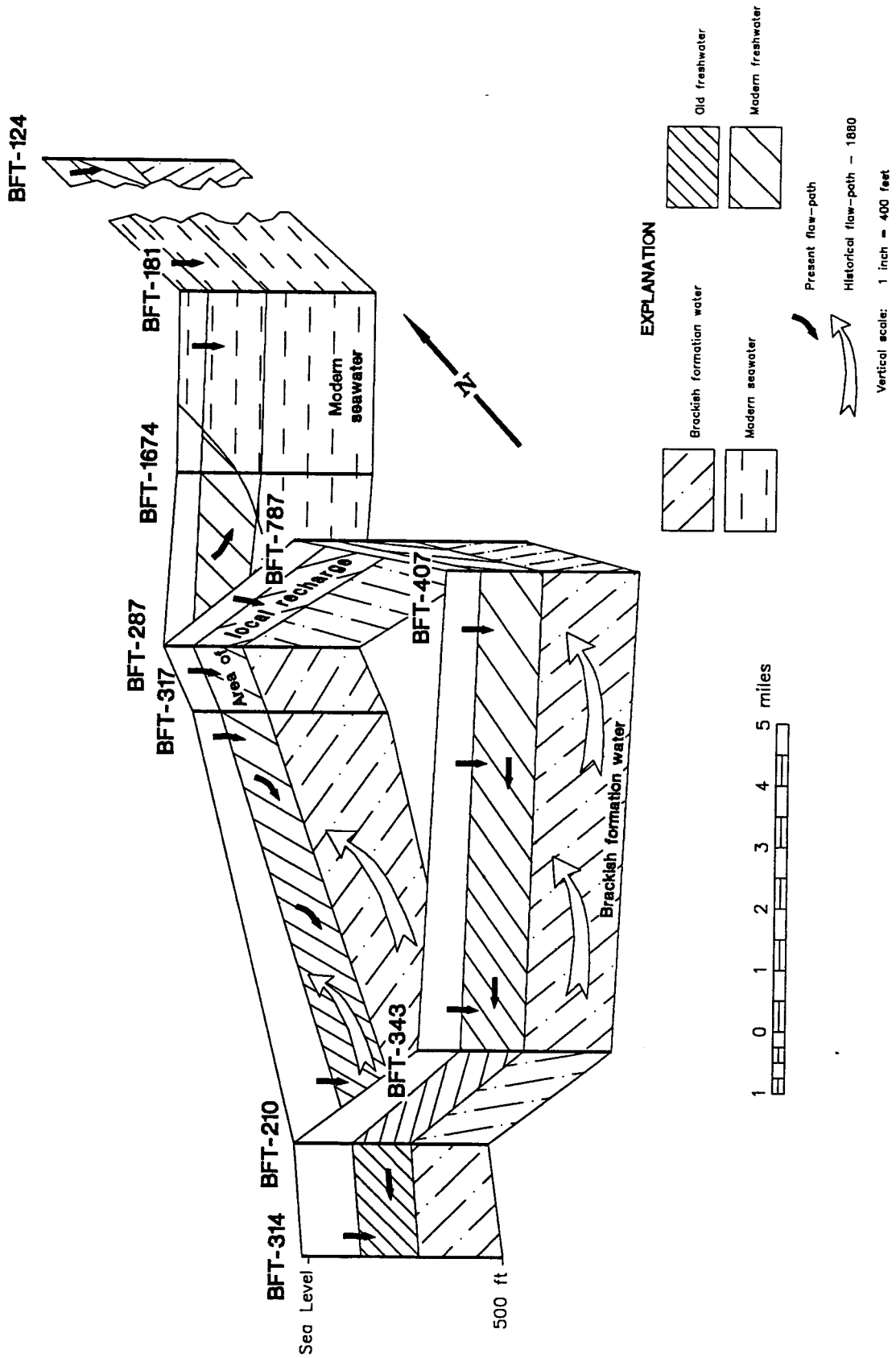


Figure 14. Hydrologic conditions as interpreted from chemical and radiocarbon isotopes (modified from Back and others, 1970, Fig. 8).

to the  $^{14}\text{C}$  activity of the seawater. Thus, it was difficult to distinguish between old and modern seawater. Brackish formation water (diluted old seawater), which is characterized by an age greater than 22,000 years and a chloride concentration greater than 1,000 mg/L, is present throughout most of the island below the upper Floridan aquifer (depth >200 ft). The modern seawater, which is characterized by an age of less than 2,000 years and a chloride concentration greater than 1,000 mg/L, presently is encroaching the aquifer beneath Port Royal Sound.

According to Figure 14, new freshwater enters the north end of the island, mixes with old native ground water and flows southwesterly toward Savannah. Additionally, brackish formation water migrates upward, because of decreasing head in the upper Floridan aquifer, and mixes with old native ground-water and/or

freshwater. Thus, ground water at Hilton Head is a composite of four different water types. Moreover, because of lower aquifer heads the ground water may be progressively contaminated with modern seawater (seawater intrusion), brackish water (upconing), and leakage from overlying aquifers.

Correct interpretation of the relation of  $^{14}\text{C}$  activities to computed ground water age can be achieved by developing simple geochemical models. Both isotopic and chemical data are necessary for correcting measured  $^{14}\text{C}$  activities to compute the ground water ages and to describe the hydrologic system. This approach, used here to study the ground water at Hilton Head Island, can be applied to other regions of the State where the understanding of hydrologic systems and ground water evolution and the evaluation of ground-water contamination is required.

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# CHEMICAL EVOLUTION OF GROUND WATER IN CRETACEOUS AQUIFERS OF THE COASTAL PLAIN IN SOUTH CAROLINA

by

Joffre E. Castro

## ABSTRACT

Three water samples from the Black Creek Formation and the chemical composition of rainwater and seawater were used to characterize the makeup of the ground water in northeastern South Carolina. Specific chemical reactions were identified from calculations of equilibrium speciation and mineral saturation, using WATEQ4F, and by mass-balance calculations using NETPATH. Both WATEQ4F and NETPATH are computer programs developed by the U.S. Geological Survey.

It was found that ground water near the recharge area in Marlboro County is noticeably affected by oxidation of pyrite and organic matter that renders the water soft and aggressive, with low pH, almost no buffering capacity, and little dissolved solids. Farther downgradient, around Marion and Dillon Counties, ground water is principally affected by calcite dissolution and calcium-sodium exchange. The water has a neutral pH, strong buffering capacity, and often a high concentration of hydrogen sulfide. Close to the coast in Horry County, the ground water is significantly affected by calcite dissolution. Because this water has been in contact with the aquifer matrix for a longer period of time, it has dissolved many of the mineral phases present in the aquifer. The ground water, thus, has a high concentration of constituents in solution. Specifically, the water is characterized by high pH, bicarbonate, chloride, and sodium. Generally the ground water was found to be of only marginal drinking-water quality.

## INTRODUCTION

Cretaceous aquifers are a principal source of water supplies in the Coastal Plain area of South Carolina. Currently, only the larger municipalities and public water utilities obtain their supply from the generally more expensive surface water installations. Ground water will continue to be an important resource because it is accessible, inexpensive to obtain, and generally requires minimum treatment. Consequently, it is important that we improve our understanding of the hydrogeology and geochemistry of these aquifers in order to develop the best management practices.

Several reports have been published that discuss the geologic and hydrologic framework of the Coastal Plain. Few reports, however, address the quality of ground water, and even fewer explain the chemical evolution that recharge water (rainwater) undergoes as it flows through the intricate paths of an aquifer. The purpose of this report is to discuss briefly the chemical reactions likely to be controlling the quality of ground water and to explain the seemingly capricious changes that characterize different regions of the Coastal Plain.

The investigation focuses on the aquifers of the Black Creek Formation in northeastern South Carolina. Figure 15 shows the study area and a ground-water flow path from Clio in Marlboro County to Conway in Horry County. This study is predicated on the

assumptions: (a) aquifers of the Black Creek Formation are primarily recharged by precipitation in Marlboro County (Zack, 1977) (b) the ground water sampled and analyzed in this study follows the aforementioned flow path on its way to the coastal area where the ground water eventually discharges into adjacent aquifers.

## METHOD OF ANALYSIS

In the forthcoming geochemical analysis, the chloride (Cl<sup>-</sup>) ion was considered a nonreactive element (tracer) and used to determine volumetric mixing ratios. This assumption is justified because aquifers of the Coastal Plain do not contain chlorine in their chemical makeup. The sources of chloride are thus connate seawater, encroaching modern seawater, or rainwater from coastal areas with elevated concentrations of chloride from sea spray.

During transgressions of the sea, seawater was trapped as pore water (connate water) in the sediments of the Coastal Plain. Since the last regression, however, recharging freshwater has been moving downgradient, flushing the aquifers and decreasing the salt content of the system. Freshwater, in this context, is recharge water from precipitation that has entered the ground water system in areas where the aquifers crop out.

Ground water in this region can be thought of, at least partially, as a mixture of rainwater and seawater.

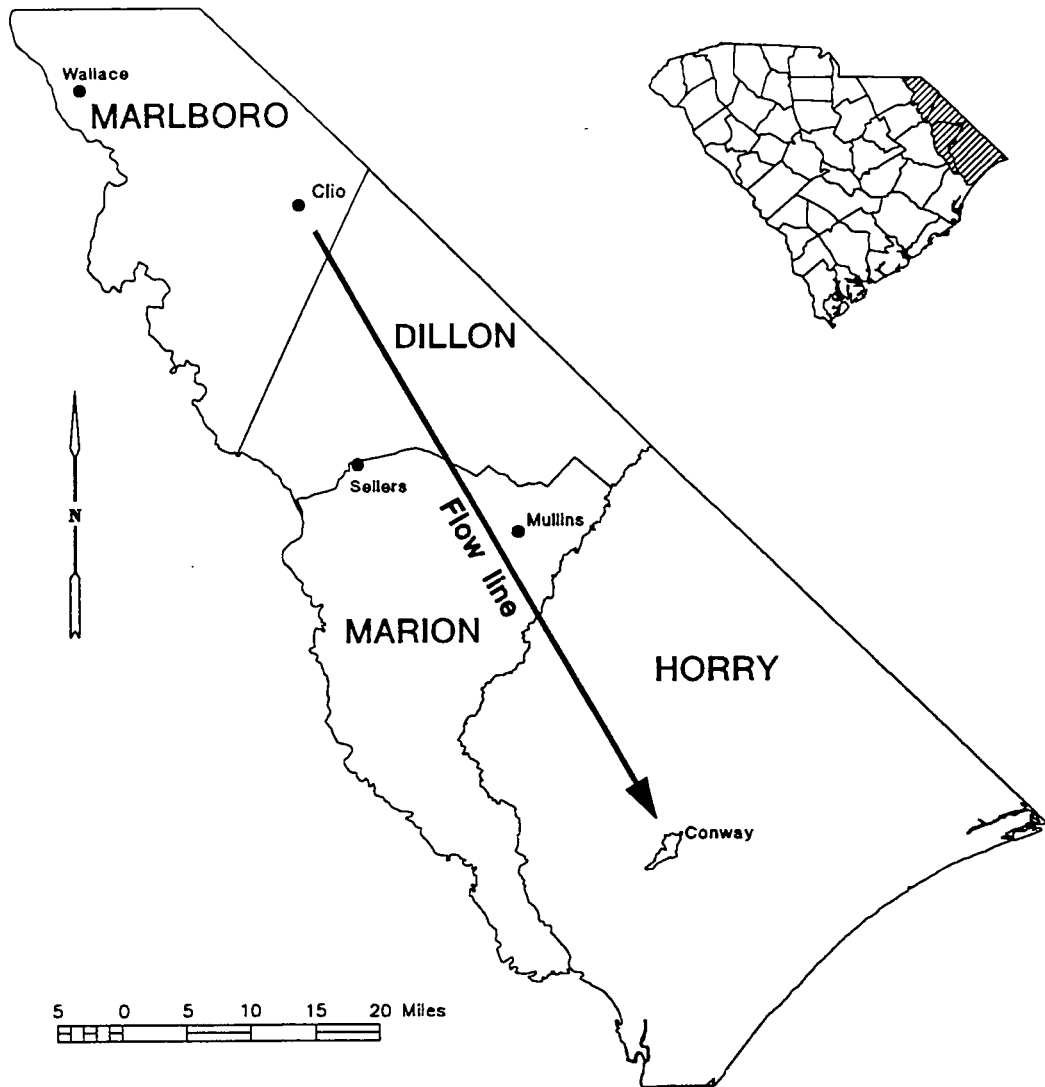


Figure 15. The study area and the ground-water flow path for the Black Creek aquifers.

This becomes apparent when chloride concentration is plotted along a flow path—a line marking the probable trajectory of a water particle moving from a recharge point to a discharge point.

Figure 16 shows that chloride ions increase in concentration from 2.5 mg/L (milligrams per liter) at Wallace in Marlboro County to 116 mg/L at Conway in Horry County. Other wells in Horry County have even higher chloride concentrations, but generally they are less than 600 mg/L. By contrast, rainwater has a chloride concentration of less than 1 mg/L and seawater of about 19,000 mg/L (Hem, 1985).

Figure 17 can be used to determine mixing proportions of rainwater and seawater in water samples. Water samples collected between Wallace and Mullins, with a chloride concentration of 5 mg/L, are a mixture

of 99.98 percent rainwater and 0.02 percent seawater. The water sample from Conway, with a chloride concentration of 116 mg/L, is a mixture 99.44 percent rainwater and 0.56 percent seawater. Since most public-health standards place an upper limit of 250 mg/L on chloride concentration, a mixture containing as little as 1.3 percent (247 mg/L of chloride) of seawater is near the point of being undesirable for human consumption.

The chemical composition of rainwater, and consequently that of the ground water recharge, is variable. Its quality depends not only on geographic location but also on the time of sampling. Table 5 shows an average composition of rainwater computed from 27 samples collected in Virginia and North Carolina between August 1962 and July 1963 (Hem, 1986).

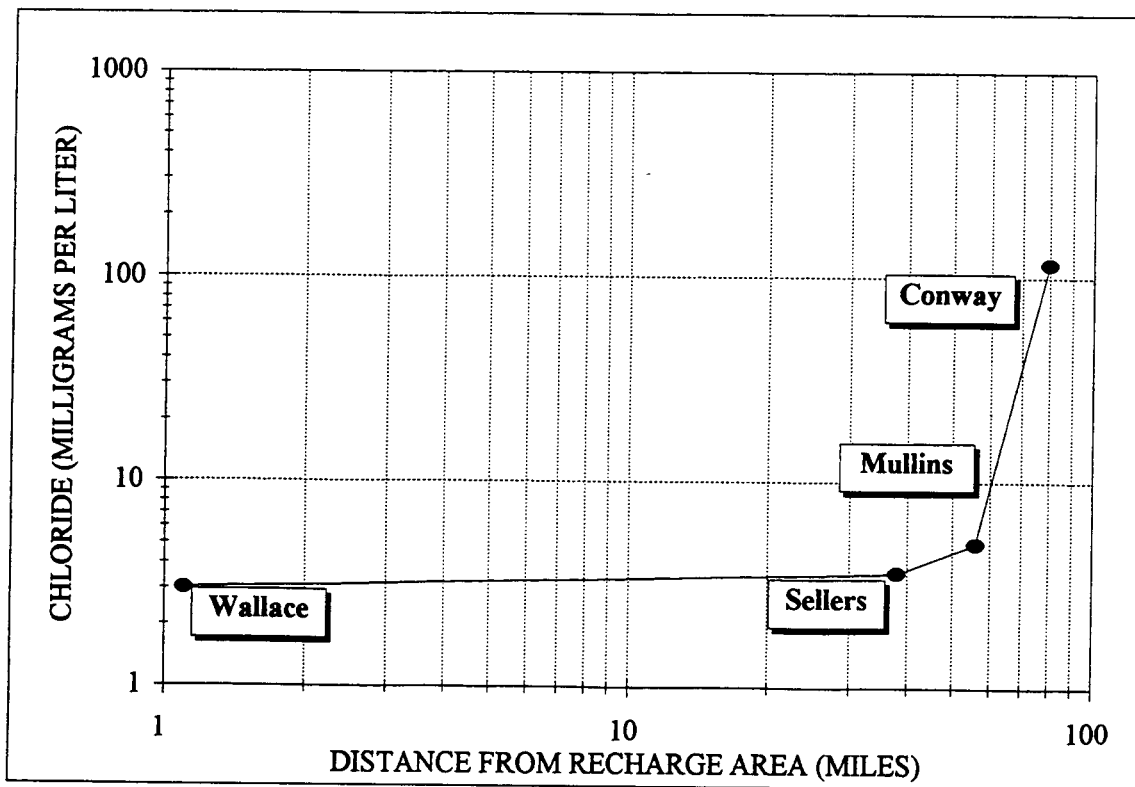


Figure 16. Correlation between chloride concentration in the Black Creek aquifers and distance along the flow path.

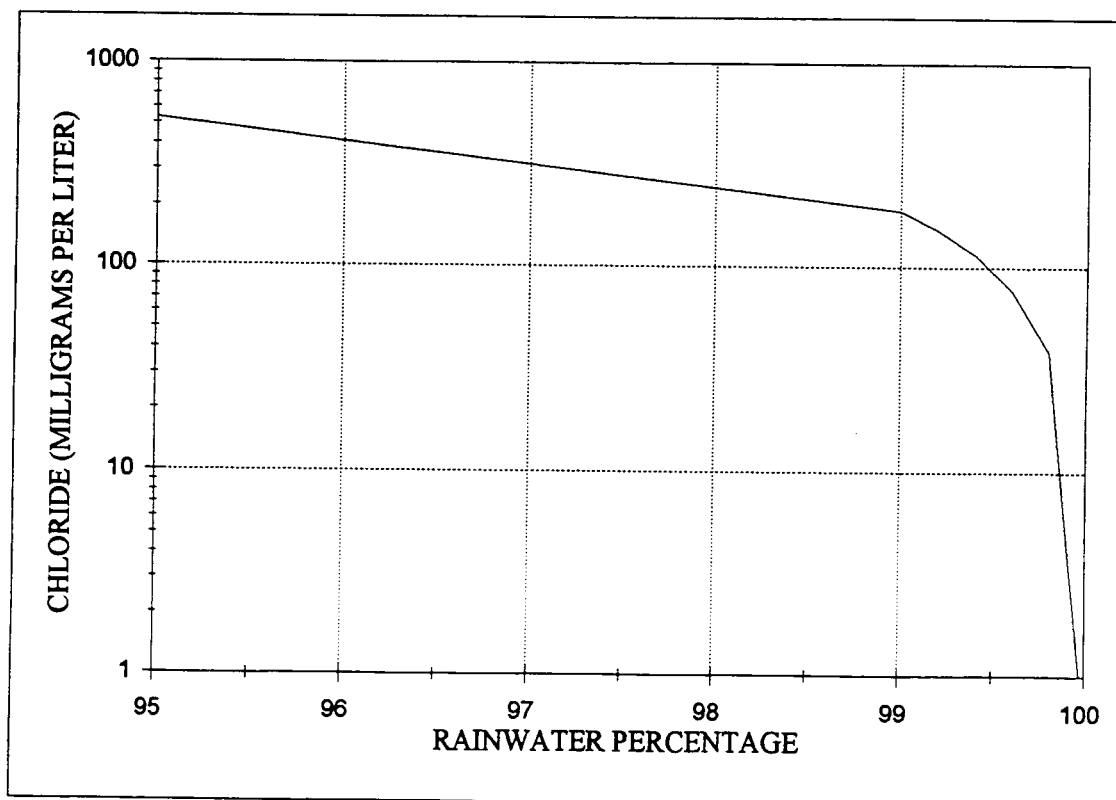


Figure 17. Chloride concentration for a given mixture of rainwater and seawater.



**Table 5. Rainwater and seawater chemical compositions**

Constituent	Rainwater	Seawater
	Concentration (mg/L)	
Calcium	0.65	410
Magnesium	.14	1,350
Sodium	.56	10,500
Potassium	.11	390
Sulfate	2.18	2,700
Chloride	.57	19,000
Nitrate	.62	.6

After Hem, 1986

The chemical composition of rainwater, as listed in Table 5, is characterized by few dissolved solids, a low pH (5.6), and low alkalinity (0.4 mg/L). Sulfate ( $\text{SO}_4^{2-}$ ) is the major anion and calcium ( $\text{Ca}^{2+}$ ) is the major cation. The high sulfate concentration of this sample—almost six times greater than that encountered in pristine environments—may be due to air pollution. Seawater, in contrast to rainwater, has a relatively stable composition.

To understand the chemical evolution of rainwater as it travels through the pore space of an aquifer, it is important to formulate a model that simplifies and best represents the natural system being considered. In this analysis it has been assumed that (a) water samples selected for analysis have traveled along the same flow path, (b) water samples can be described as a mixture of a previous sample with seawater, and (c) all mixtures have been affected by chemical reactions with soil horizons and aquifer sediments.

The flow path selected (Fig. 15) starts in northern Marlboro County where the Black Creek aquifers crop out and rainwater recharges them. It continues in a southeasterly direction, crossing Clio in eastern Marlboro County and Sellers in Marion County, and ends at Conway in Horry County. Three water samples were selected for the analysis. Locations and compositions are listed in Table 6.

When rainwater reaches the ground surface, only a small fraction of the total precipitation infiltrates the upper soil layers to eventually recharge the aquifers. These soil horizons, which are characterized by the presence of certain organic and inorganic components, have a strong influence on the chemical composition of ground water. Some of these geochemical processes are readily inferred by comparing the chemical composition of the Clio water sample with that of the rainwater. It is apparent that the Clio water has not reached chemical equilibrium with minerals present

**Table 6. Composition of selected water samples from the Black Creek aquifers**

Constituent	Clio	Sellers	Conway
	Depth (ft)	150	360
Temperature	22.0	22.0	24.0
pH	6.6	7.4	8.5
Bicarbonate	13.2	112.	550.
Calcium	2.6	5.9	4.2
Magnesium	1.8	1.3	.5
Sodium	3.3	38.	265.
Potassium	3.3	4.9	3.7
Chloride	2.5	3.6	116.
Fluoride	.1	.36	3.9
Silica	12.	33.	14
Iron	.5	.14	.0
Sulfate	6.	3.4	3.4

Concentration in mg/L; temperature in degrees Celsius; pH in standard units.

in the aquifer (low concentrations), but has reacted with organic matter and carbonate minerals enough to increase its alkalinity and pH. Moreover, because this water has traveled only a short distance from the recharge area, its chemical composition still resembles rainwater. The other two samples, which were collected from wells 20 miles (Sellers) and 60 miles (Conway) down the gradient from Clio, have had a longer time to react with the aquifer and therefore display a much higher concentration of dissolved solids. Ground water in eastern Horry County has been in the aquifer for approximately 50,000 years and, thus, has moved closer to attaining chemical equilibrium with the aquifer.

The evolution of ground water in the aquifers of the Black Creek Formation has been divided into three stages. In the first stage, the Clio water, which is representative of ground water near recharge areas, is simulated by mixing rainwater with minerals present in the surficial layers of the aquifer. In the second stage, the Sellers sample, which is halfway between the recharge and discharge points, is modeled by mixing the Clio water with minerals present in these aquifers. In the final stage, the Conway sample is simulated by mixing the Sellers sample and seawater and allowing it to react with minerals present in the deep layers of the aquifers.

A geochemical computer code "NETPATH" (Plummer and others, 1991) was used to evaluate net geochemical reactions between rainwater, seawater, and ground water. The program, using water quality data, examines every possible reaction model for a set of constraints (including mass balance) and speci-

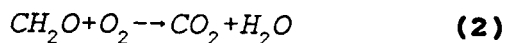
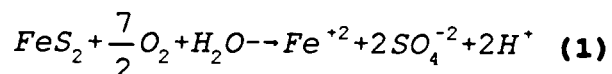
fied mineral phases. Additionally, the WATEQ4F (Ball and Nordstrom and others 1991) code was used to compute ionic balance, partial pressures of gasses, and index of saturation of mineral phases. Results from these models were verified with mineralogical and chemical information available for the ground water system.

## MODELING OF THE CHEMICAL EVOLUTION OF GROUND WATER

**CLIO MODEL.**—In this model the water sample from Clio is considered to have evolved from the mixing of rainwater and seawater. From the Cl, the sample was determined to be composed of 99.99 percent rainwater and 0.01 percent seawater. Table 6 shows that all dissolved ions in the Clio sample have increased with respect to those of rainwater, that silica has considerably increased, and that dissolved oxygen (O<sub>2</sub>) has decreased. The speciation code indicated that the ground water is undersaturated with calcite and slightly supersaturated with silica and amorphous ferric hydroxide and that the partial pressure of carbon dioxide (CO<sub>2</sub>), 10<sup>-2.5</sup> atmospheres (atm), is greater than the atmospheric partial pressure (10<sup>-3.5</sup> atm).

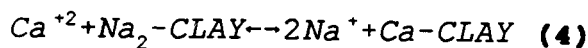
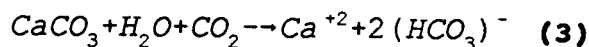
Table 7 lists the results of this geochemical simulation with regard to the evolution of Clio water. In the table, **NET DIFFERENCE** is the mass difference between the **Final water**, CLIO sample, and a mixture of **Initial water 1 (RAINWATER)** and **Initial water 2 (SEAWATER)**. According to the model, a mixture of 99.99 percent rainwater and 0.01 percent seawater could have evolved into the Clio water sample if the reactions listed under **PROCESS** dissolved or precipitated in the given concentrations. **NET GAIN** is the total concentration predicted by model that must be dissolved (+) or precipitated (-) from the solution (mixture) to form the final water. The **NET GAIN**, therefore, must be equal to the **NET DIFFERENCE** if the model is successful.

The model, which is proposed by NETPATH as the most likely, suggests that oxidation of pyrite and organic matter is the dominant process during the first stage of the ground-water evolution. Oxygen, initially present in rainwater, is consumed during the oxidation of pyrite (Eq. 1) and organic matter (Eq. 2), thus, increasing the concentration of ferrous iron and dissolved carbon dioxide (CO<sub>2</sub>) in solution. Organic matter, moreover, is aerobically and anaerobically oxidized.



The specific processes describing the anaerobic oxidation of organic matter, however, were not directly modeled. Instead, an inorganic carbon phase (CO<sub>2</sub>) was assumed to be present in the system to satisfy mass balance. McMahon and Chapelle (1991) suggested that some of this CO<sub>2</sub> is produced in situ by anaerobic processes, such as fermentation.

Other important processes identified by the model are calcite dissolution and cation exchange. Calcium dissolved from a calcium carbonate phase (Eq. 3) is subsequently removed from solution by an exchange process with sodium on the surface of the clay (Eq. 4). The net effect of these reactions is a slight increase in pH and a modest increase in the bicarbonate concentration.



Potassium feldspar dissolution and Na-Mg exchange provided enough potassium, magnesium, aluminum, and silica in solution to achieve mass balance. Finally, some aluminum and silica reprecipitated as kaolinite and gibbsite.

Consequently, the Clio sample is characterized by low pH (6.6) and total dissolved solids. The water is acidic with high CO<sub>2</sub> and a low buffer capacity; therefore, it is aggressive. Pyrite oxidation leads to elevated ferrous iron concentrations, which is common in this region near recharge areas.

**SELLERS MODEL.**—In this model the chemical evolution of a water sample collected from a well located 2 miles southwest of Sellers (Marion County) was simulated. The chemical composition of the sample is listed in Table 6. For modeling, this water sample was considered to have evolved from a mixture of the Clio water sample (99.995 percent) and seawater (0.005 percent).

Table 8 shows the results of the modeling. During this stage of the ground water evolution, carbonate dissolution and Ca-Na exchange are responsible for most of the changes in the chemical composition of the Sellers water. The pH increases to 7.4 and

TABLE 7. Geochemical evolution model of the Clío water sample

SAMPLE	PERCENT	Carbon	Calcium	Sodium	Magnesium	Potassium	Silica	Sulfate	Iron	Aluminum	
Initial water 1	RAINWATER 99.9900	0.0578	0.0162	0.0244	0.0058	0.0028	0.0017	0.0190	0.0000	0.0004	
Initial water 2	SEAWATER 0.0100	2.1310	10.6547	485.4534	55.0703	10.5761	0.0739	29.2608	0.0000	0.0001	
Simple mixture		0.0580	0.0173	0.0729	0.0113	0.0039	0.0017	0.0219	0.0000	0.0004	
Final water	CLIO	0.3299	0.0649	0.1436	0.0740	0.0844	0.1997	0.0625	0.0090	0.0019	
NET DIFFERENCE		0.2719	0.0476	0.0707	0.0627	0.0805	0.1980	0.0406	0.0090	0.0015	
<b>GEOCHEMICAL MODELING</b>											
PROCESS	(mmol/kg)										
CO2(g)	ingassing	0.0273									
(Bacteriologic processes)											
Calcite	dissolution	0.1469	0.1395		0.0073						
(Ca.95Mg.05CO3)											
Ca/Na exchange			-0.0919	0.1838							
Mg/Na exchange				-0.1113	0.0556						
Organic matter	oxydation	0.09777									
(CH2O)											
Gibbsite	precipitation	-0.0354									
(Al(OH)3)											
Pyrite	oxydation	0.0204									
(FeS2)											
K-feldspar	dissolution	0.0806									
(KAlSi3O8)			0.0806	0.2417							
Kaolinite	precipitation	-0.0218									
(Al2Si2O5(OH)4)				-0.04366							
Goethite	precipitation	-0.0114									
(FeOOH)							-0.0114				
NET GAIN		0.2720	0.0476	0.0725	0.0630	0.0806	0.1981	0.0407	0.0090	0.0015	

(-) minus--> removed from solution

TABLE 8. Geochemical evolution model of the Sellers water sample

SAMPLE	PERCENT	Carbon	Calcium	Sodium	Magnesium	Potassium	Silica	Sulfate	Iron
Initial water 1	99.9950	0.3299	0.0649	0.1436	0.0740	0.0844	0.1997	0.0625	0.0090
Initial water 2	0.0050	2.1310	10.6547	485.4534	55.0703	10.5761	0.0739	29.2608	0.0000
Simple mixture		0.3300	0.0654	0.1679	0.0767	0.0849	0.1997	0.0640	0.0090
Final water		1.9870	0.1472	1.6532	0.0535	0.1253	0.5493	0.0354	0.0025
NET DIFFERENCE		1.6570	0.0818	1.4853	-0.0232	0.0404	0.3496	-0.0286	-0.0065
<b>GEOCHEMICAL MODELING</b>									
PROCESS		(mmol/kg)							
CO2(g)	ingassing	0.7927							
(Bacteriologic processes)									
Calcite	dissolution	0.7997	0.7598		0.0400				
(Ca.95Mg.05CO3)									
Ca/Na exchange			-0.6780	1.3559					
Mg/Na exchange				0.1271	-0.0636				
Organic matter	oxydation	0.0646							
(CH2O)									
Silica (a)	dissolution						0.2285		
(SiO2)									
Pyrite	precipitation							-0.0287	-0.0143
(FeS2)									
K-feldspar	dissolution					0.0404	0.1211		
(KAIS3O8)									
Goethite	dissolution								0.0078
(FeOOH)									
NET GAIN		1.6570	0.0818	1.4831	-0.0236	0.0404	0.3496	-0.0287	-0.0065

(-) minus--> removed from solution

the bicarbonate to 112 mg/L, rendering this water neutral and well buffered. The model predicted that organic matter is anaerobically oxidized first by the reduction of ferric iron from goethite and later by the reduction of sulfate.

In contrast with the previous sample, the concentration of dissolved iron is low, probably because ferrous ions in solution precipitated with sulfide ions to form pyrite. Sulfide ions became available to the system from the reduction of sulfate to sulfide during oxidation of organic matter. Finally, note that the concentration of silica is considerably higher than in the other two samples. This sample, as well as others from the Black Creek aquifers in Marion County and Florence County, has the highest concentration of dissolved silica of any sample taken from the Cretaceous aquifers in South Carolina. It has been noted that silica could replace calcite as the cementing material in the aquifer.

**CONWAY MODEL.**—The water sample from Conway was assumed to have evolved from a mixture of 99.44 percent Sellers sample and 0.56 percent seawater. The makeup of the sample is listed in Table 6.

The geochemical models considered thus far assumed that sodium, available from exchange sites on marine clay, exchanged with calcium in solution. This indicates that dissolved calcium concentrations are high enough to support the exchange process. The water sample from Conway, according to the speciation code, is supersaturated with respect to calcite, suggesting that calcium spontaneously precipitates and not dissolves. Consequently, calcium would not be available for exchange with sodium. To resolve this apparent contradiction, two possible solutions are proposed: (a) there is another source of sodium, and (b) the pH of the water sample is incorrect. In the following discussion both solutions are explored.

**ALTERNATE SODIUM SOURCE (MODEL A).**—It was mentioned earlier that in order to explain the observed increase of inorganic carbon, McMahon and Chappelle (1991) proposed a flux of dissolved inorganic carbon from the anaerobic oxidation of organic matter available in the confining beds. To maintain charge balance, however, a companion cation had to be added to this flux. They elected to use a sodium ion. This sodium could then be used to replace the sodium otherwise available for exchange on the marine clay. To account for this sodium flux, a  $\text{NaHCO}_3$  "mineral phase" was assumed to be present in the system.

Model A is one of several possible solutions predicted by Netpath. Differences between the versions of the model were minor. According to model A (Table 9), at least 62 percent of the sodium in solution came as a flux from confining beds and 38 percent from seawater (mixture). The inorganic carbon in solution, mostly present as bicarbonate, came partially as dissolved  $\text{CO}_2$  from confining beds and partially from anaerobic oxidation of organic matter in the aquifer. The model also indicated that kaolinite, phlogopite, and potassium mica precipitated in small quantities.

**SODIUM FROM EXCHANGE PROCESS (MODEL B).**—Model B does not consider the contribution of  $\text{Na}^+$  from the confining beds, but rather it assumes that the sodium originated from the Ca-Na exchange. To proceed with this approach, however, the pH of the sample had to be decreased to 8.1. At this pH, the ground water becomes undersaturated with respect to calcite. It is possible that when the sample was collected, degassing of  $\text{CO}_2$  had taken place, resulting in the pH being increased to 8.5.

According to model B (Table 10), calcium from calcite dissolution is exchanged for sodium from marine clay. This model also requires a source of inorganic carbon, which again is assumed to have come from the anaerobic oxidation of organic matter. Finally, kaolinite, phlogopite, and potassium mica, as in model A, are predicted to have precipitated in small amounts.

## PETROGRAPHIC EVIDENCE OF CHEMICAL PROCESSES

Mineralogic evidence for some of the geochemical processes predicted by the ground water models can be seen in the following Scanning Electron Microscope taken of core samples from the Black Creek aquifers at Myrtle Beach. In Figure 18, the photomicrographs of the 404 ft bls (below land surface) samples show quartz sand of two distinctive sizes, very fine and medium, with good porosity. More interesting, however, is the presence of feldspar grains in the process of dissolution, which the geochemical models predicted on the basis of the chemical composition of the water. Figure 18 also shows photomicrographs from a sample at 550 ft bls. The sand is fine grained and well sorted with good porosity. Also evident is authigenic smectite with authigenic pyrite crystals and pyrite framboids. From the x-ray analysis, smectite and illite were observed in glauconitic grains. Thus, this petrographic information supports the results of the geochemical model regarding

TABLE 9. Geochemical evolution model of the Conway water sample, model A

SAMPLE	PERCENT	Carbon	Calcium	Sodium	Magnesium	Potassium	Silica	Sulfate	Iron
Initial water 1	99.4390	1.9870	0.1472	1.6532	0.0535	0.1253	0.5493	0.0354	0.0025
Initial water 2	0.5610	2.1395	10.6547	485.4534	55.0703	10.5761	0.0739	29.2608	0.0000
Simple mixture		1.9879	0.2061	4.3673	0.3621	0.1839	0.5466	0.1994	0.0025
Final water		11.2663	0.1049	11.5379	0.0206	0.0947	0.2332	0.0354	0.0000
NET DIFFERENCE		9.2784	-0.1012	7.1706	-0.3415	-0.0892	-0.3134	-0.1640	-0.0025
<b>GEOCHEMICAL MODELING</b>									
CO2(g)	ingassing	2.21331							
(Bacteriologic processes)									
Calcite	precipitation	-0.1066	-0.1013		-0.0053				
(Ca.95Mg.05CO3)									
Glauconite	dissolution					0.07948	0.47688		0.07948
(K2Fe2Al6(Si4O10)3(OH)12)									
Pyrite	precipitation								
(FeS2)									
K-mica	precipitation					-0.0566	-0.1699		
(KA13S3O10(OH)2)									
Kaolinite	precipitation								
(Al2Si2O5(OH)4)									
Phlogopite	precipitation								
(KMg3AlSi3O10(OH)2)									
NaHCO3		7.1718		7.1718					
NET GAIN		9.2785	-0.1013	7.1718	-0.3416	-0.0893	-0.3134	-0.1639	-0.0025

(-) minus--> removed from solution

**TABLE 10. Geochemical evolution model of the Conway water sample, model B**

SAMPLE	PERCENT	Carbon	Calcium	Sodium	Magnesium	Potassium	Silica	Sulfate	Iron
Initial water SELLERS	99.4390	1.9870	0.1472	1.6532	0.0535	0.1253	0.5493	0.0354	0.0025
Initial water SEAWATER	0.5610	2.1395	10.6547	485.4534	55.0703	10.5761	0.0739	29.2608	0.0000
Simple mixture		1.9879	0.2061	4.3673	0.3621	0.1839	0.5466	0.1994	0.0025
Final water CONWAY		9.0644	0.1049	11.5379	0.0206	0.0947	0.2332	0.0354	0.0000
NET DIFFERENCE		7.0765	-0.1012	7.1706	-0.3415	-0.0892	-0.3134	-0.1640	-0.0025
<b>GEOCHEMICAL MODELING</b>									
PROCESS		(mmol/kg)							
CO <sub>2</sub> (g) ingassing (Bacteriologic processes)	3.4188	3.41883							
Calcite dissolution (Ca <sub>95</sub> Mg <sub>05</sub> CO <sub>3</sub> )	3.6673	3.6673	3.4839		0.1834				
Ca/Na exchange	3.5852		-3.5852	7.1705					
Glauconite dissolution (K <sub>2</sub> Fe <sub>2</sub> Al <sub>6</sub> (Si <sub>4</sub> O <sub>10</sub> ) <sub>3</sub> (OH) <sub>12</sub> )	0.03973					0.07946	0.47676		0.07946
Pyrite precipitation (FeS <sub>2</sub> )	-0.0820							-0.1639	-0.0820
K-feldspar dissolution (KAlSi <sub>3</sub> O <sub>8</sub> )	0.0063					0.0063	0.0188		
Kaolinite precipitation (Al <sub>2</sub> Si <sub>2</sub> O <sub>5</sub> (OH) <sub>4</sub> )	-0.1421						-0.2841		
Phlogopite precipitation (KMg <sub>3</sub> AlSi <sub>3</sub> O <sub>10</sub> (OH) <sub>2</sub> )	-0.1750						-0.5249	-0.1750	-0.5249
NET GAIN		7.0861	-0.1013	7.1705	-0.3415	-0.0893	-0.3135	-0.1639	-0.0025

(-) minus--> removed from solution

(+) plus --> added to solution

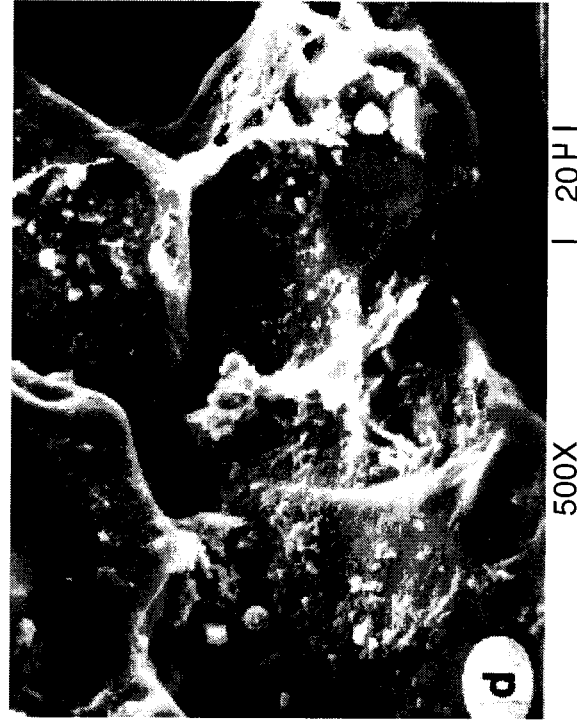
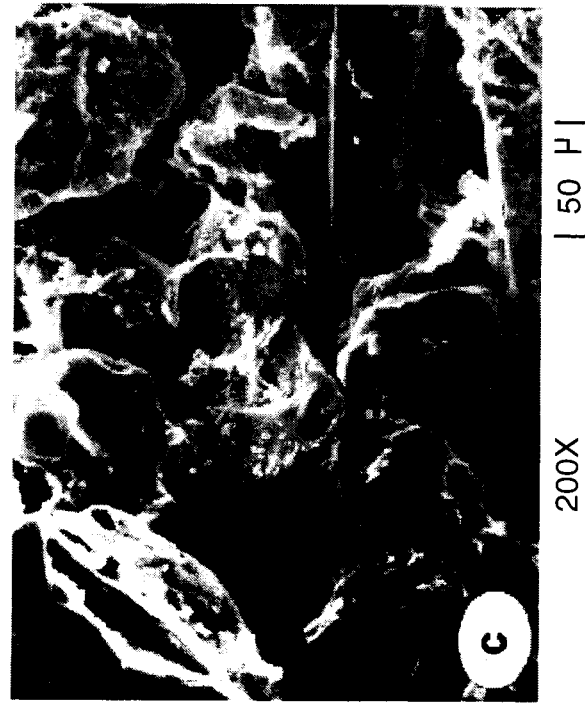
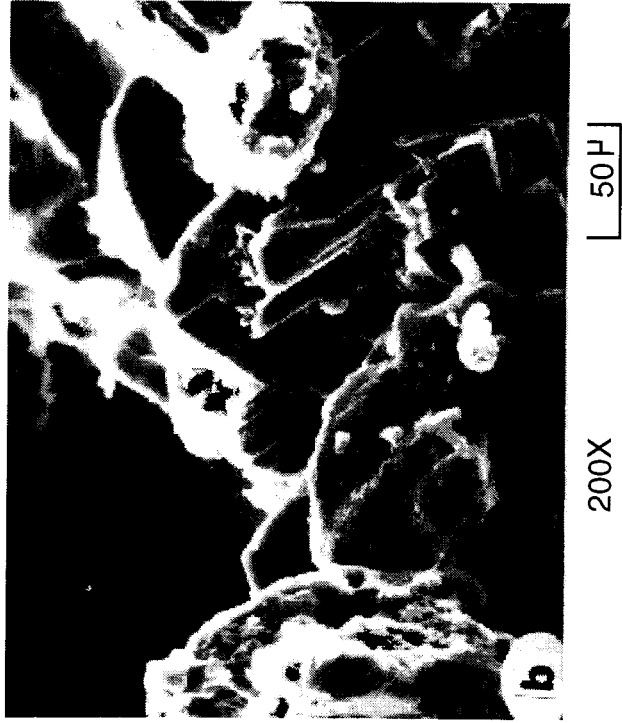
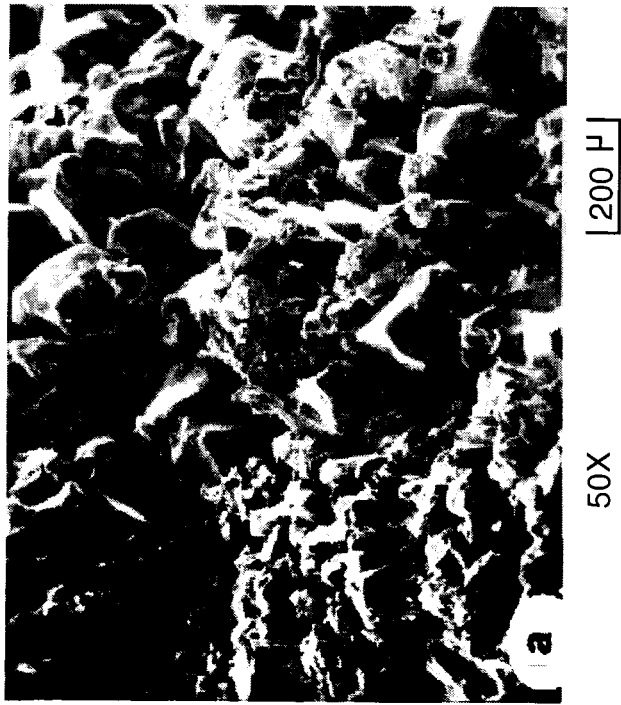


Figure 18. Photomicrographs from SEM analysis at 404 ft bls (a and b) and 550 ft bls (c and d), where (a) shows grain sizes of very fine and medium sand, (b) shows feldspar dissolution, (c) reveals a fine-grained, well-sorted sand, and (d) shows authigenic smectite with authigenic pyrite.



the precipitation of pyrite and the presence of glauconite in the system.

## SUMMARY

In northeastern South Carolina, rainwater that infiltrates surficial soil and moves downward to recharge the water table (about 10 percent of the total precipitation) is characterized by high dissolved oxygen and carbon dioxide, low dissolved solids, low pH, and low alkalinity. In general it is soft and very aggressive. By the time this ground water has reached the deeper aquifers, as in eastern and southern Marlboro County, the dissolved oxygen has been depleted, dissolved solids are about 30 mg/L, the pH 6.6, and it has acquired a high concentration of dissolved iron. The iron ion is a byproduct of the anaerobic oxidation of organic matter and remains in solution until enough sulfide is available to precipitate pyrite.

Farther down the gradient, in Dillon and Marion Counties, where this aggressive water has had a longer

time to react with the aquifer, the ground water is characterized by a dissolved-solids concentration of about 150 mg/L, a pH between 7 and 8, an alkalinity of approximately 110 mg/L, and a sodium concentration around 40 mg/L. By the time the ground water has reached this locale, a significant amount of sulfate has been reduced to sulfide by means of anaerobic sulfate reduction. The sulfide and ferrous ion precipitate to form pyrite, causing the concentration of these ions to decrease. In some wells of this region, water having a black precipitate has been reported, which could indicate the presence of sulfide in the system.

Once ground water reaches Horry County, it can be characterized by a dissolved-solids concentration over 600 mg/L, a pH of about 8.2, alkalinity close to 700 mg/L, and a sodium concentration of about 250 mg/L. Additionally, although not modeled, ground water in the coastal counties has a high concentration of fluoride. The fluoride is believed to have come from fluorapatite (Zack, 1980). Apparently the dissolution of calcite exposes the fluorapatite, making it susceptible to dissolution.

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# WELL EFFICIENCY—ITS IMPORTANCE AND ITS CALCULATION

by

Roy Newcome, Jr.

## Abstract

Efficient water wells cost less to operate than inefficient wells. Well efficiency can be calculated from a pumping test in which the well is pumped at a constant rate and the resulting water-level drawdown is measured. The same data that are used to calculate the aquifer transmissivity are used to calculate the well's specific capacity. The ideal specific capacity of a well producing from a confined aquifer bears a linear relation to the transmissivity. Well efficiency is the ratio of actual specific capacity to ideal specific capacity. One-third of the nearly 500 wells tested in South Carolina's Coastal Plain have well efficiencies greater than 90 percent.

An efficient water well is one in which the water level during pumping is not much deeper than the potentiometric level in the aquifer just outside the well. The more difference there is in these levels, the less efficient is the well. Because the cost of pumping water (power cost) varies directly with the depth from which it is pumped, it behooves well owners to have the most efficient wells they can obtain. This means that the design and development of wells should be based not just on obtaining the required yield (gallons per minute) but on obtaining that yield with the least drawdown of the water level in the well. Wells for which the diameter and length of screen and size of screen openings are properly selected for the aquifer's thickness and grain size, and for which the development process is adequately carried out, are likely to have high efficiencies.

The key element in well yield and well efficiency is the specific capacity. This, simply, is the number of gallons per minute the well yields for each foot of water-level drawdown. It is usually calculated for a 1-day period to provide standard comparative data. The calculated specific capacity can then be compared with the "ideal" specific capacity of the well to ascertain the well efficiency. Ideal specific capacity would be that of a fully (100-percent) efficient well.

The ideal specific capacity is directly related to the aquifer properties of transmissivity and storage and to the well size (diameter). Transmissivity ordinarily is calculated from a pumping test in which drawdown and recovery measurements made in a pumped well and/or observation well(s) during and after a period of constant-rate pumping are plotted against elapsed time. Ascertaining the storage coefficient requires a pumping

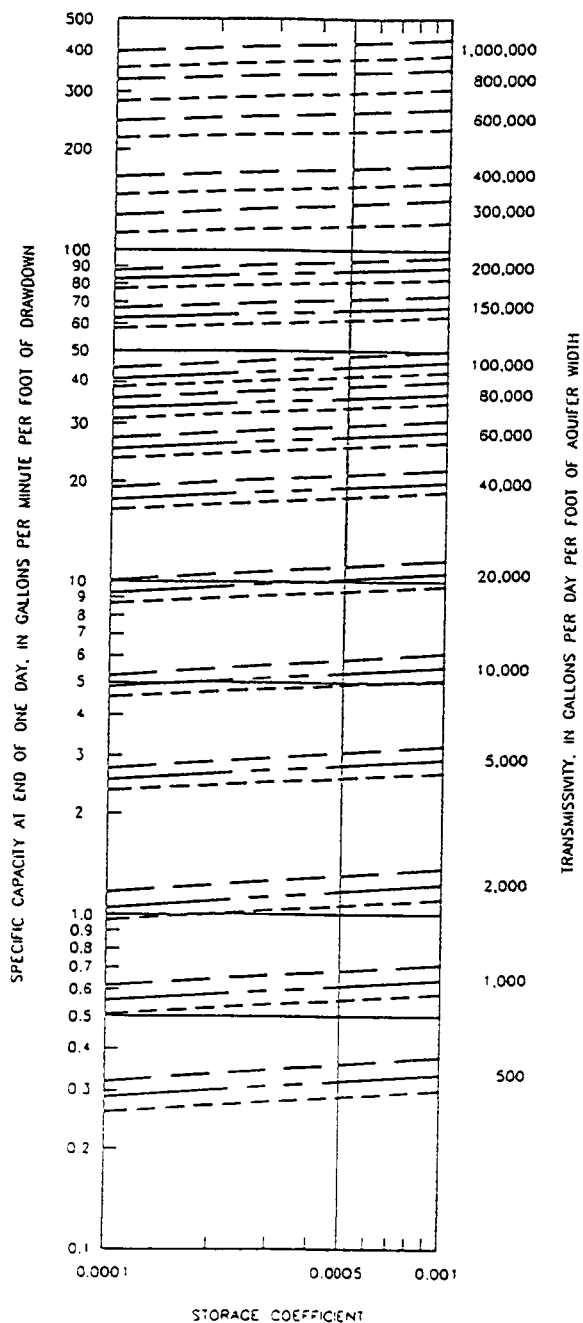
test involving one or more observation wells in addition to the pumped well, but usually it can be reasonably estimated. The diameter of the well screen—or in the case of an unscreened well, as is common in limestone aquifers, the well bore—will be known from the drilling record. Hydraulics equations are available for calculating the ideal specific capacity once the foregoing aquifer and well properties are known. The parameters of transmissivity, storage coefficient, well diameter, and specific capacity have been interrelated in a graphical form (Fig. 19). It can be seen from this graph that, for the common range of confined-aquifer storage coefficients (0.0001-0.001), an average value of 2,000 is a reasonable and easy-to-remember relation between transmissivity, in gallons per day per foot, and specific capacity, in gallons per minute per foot of drawdown. Only confined aquifers are treated in this discussion because, almost without exception, they are the ones tapped by major wells in South Carolina.

The practical application of all this for the well owner or consultant is that by means of a simple pumping test (a pumping test is required by the South Carolina Department of Health and Environmental Control for all public-supply wells) the information can be obtained on aquifer transmissivity and well specific capacity that permits the estimation of well efficiency. Merely divide the transmissivity by 2,000 to obtain the ideal specific capacity and then divide the latter into the well's calculated true specific capacity to obtain the well efficiency. Because of several assumptions (such as storage coefficient, effective well diameter, and the 2,000 rule-of-thumb), it is not practical to quote well efficiency more precisely than the nearest 5 percent. Moreover, if

hydrologic boundaries distort the semilogarithmic graphs of drawdown and recovery (versus time), the specific capacity must be calculated from a projection of the drawdown graph prior to the boundary effect. This is the part of the graph that provides the transmissivity value used to estimate ideal specific capacity (see Figure 20).

Analysis of the efficiencies of nearly 500 sand-aquifer wells indicates that South Carolina's Coastal Plain wells generally have high efficiencies. As shown by the cumulative graph of Figure 21, more than half of the wells tested had efficiencies greater than

70 percent. Only a quarter of them were less than 50-percent efficient. R.C. Heath, of the U.S. Geological Survey, stated in his Basic Ground-water Hydrology (Water-Supply Paper 2220, p. 59) that "Under the best conditions, an efficiency of about 80 percent is the maximum that is normally achievable in most screened wells. Under less than ideal conditions, an efficiency of 60 percent is probably more realistic." It appears that South Carolina wells are somewhat more efficient than Heath suggests as "normally achievable," one-third of those tested being more than 90-percent efficient.



To obtain the specific capacity, find the transmissivity value indicated by the pumping test, follow the appropriate well-diameter line to its intersection with the calculated or assumed storage-coefficient line, and then read horizontally to the left margin. All scales are logarithmic, and this should be considered when interpolating for other values of transmissivity.

This chart is for the normal range of confined-aquifer storage coefficients. It is based on the well being 100-percent efficient.

The chart is adapted from one presented by R. R. Meyer in U. S. Geological Survey Water-Supply Paper 1536-1 in 1963.

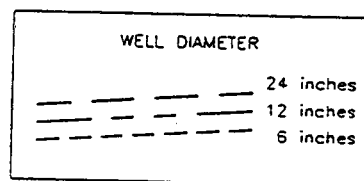
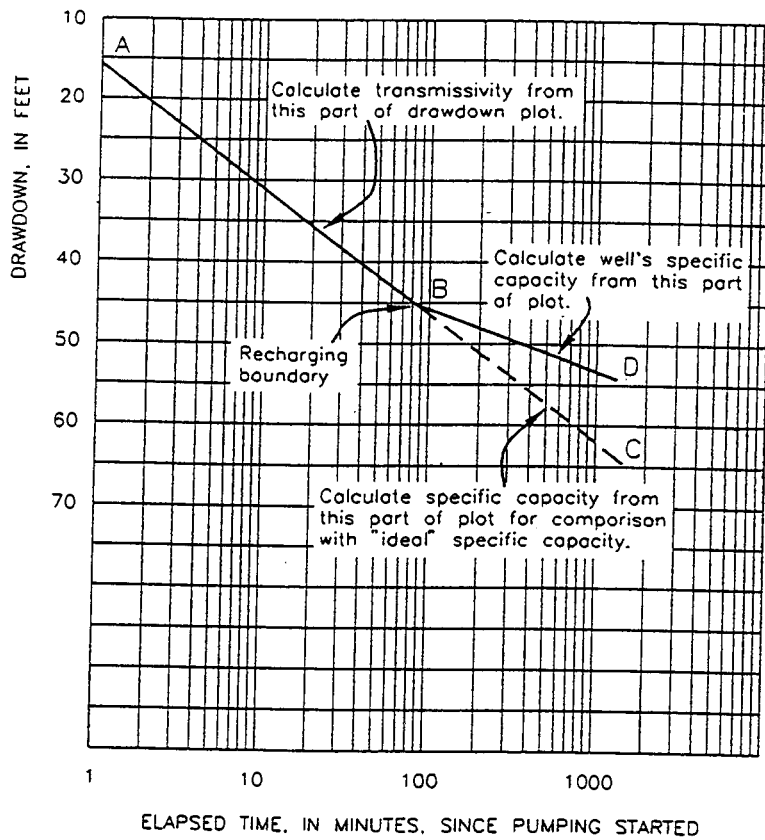


Figure 19. Illustrating the interrelationship among transmissivity, storage coefficient, well size, and specific capacity.



$$\text{Specific capacity of well} = \frac{\text{Pumping rate (gallons per minute)}}{\text{Drawdown indicated by ABD (feet)}} = \text{gallons per minute per foot}$$

$$\text{Ideal specific capacity} = \frac{\text{Transmissivity indicated by AB (gallons per day per foot)}}{2,000} = \text{gallons per minute per foot}$$

$$\text{Well efficiency} = \frac{\text{Specific capacity indicated by ABC}}{\text{Ideal specific capacity}} \times 100 = \text{percent}$$

Figure 20. Illustrating the estimation of well efficiency where a hydrologic boundary is encountered.

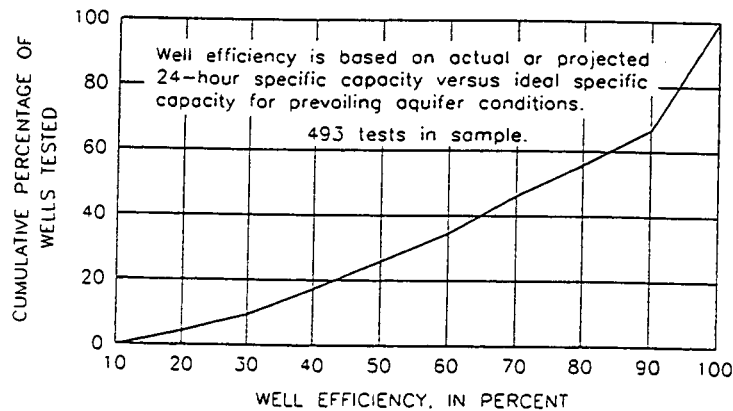


Figure 21. Illustrating the efficiencies of wells in the Coastal Plain sand aquifers of South Carolina.

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# ANALYSIS OF INTERFERENCE-TYPE PUMPING TESTS OF THE COASTAL PLAIN AQUIFERS IN SOUTH CAROLINA

by

Roy Newcome, Jr.

## Abstract

In 76 interference-type pumping tests made in the South Carolina Coastal Plain, drawdown was measured in 61 of the pumping wells. The pumped-well and observation-well data provided corroborating transmissivity values in 50 tests. Examination of the 11 tests in which the transmissivity indicated by pumped and observation wells differed by 23 percent or more provided a definite or likely explanation in all cases. Usually it was a matter of pumped well and observation well not producing from exactly the same aquifer(s). Well efficiency is not relevant in calculating transmissivity from pumped-well data, and storage coefficient probably can be as reliably estimated as calculated from data. Therefore, single-well pumping tests provide results that are likely to be as reliable and useful as those obtained from interference tests, and at much less cost.

The files of the South Carolina Department of Natural Resources, Water Resources Division, as of September 1996, contained about 625 pumping tests made at wells producing water from Coastal Plain aquifers. Only 76 of the tests are interference tests—that is, they measured the pumping effect of a well on the water level in one or more other wells. All the other tests are single-well tests in which pumping effects were measured only in the well being pumped.

The 76 interference tests were made in 17 of the 28 Coastal Plain counties and represent most of the confined-aquifer systems in that region. These tests provide the aquifer-hydraulics parameters of transmissivity and storage coefficient, which are required for the prediction of aquifer and well yields and well interference. The single-well tests provided the transmissivity but not the storage coefficient. A transmissivity value is necessary in predicting well yields and pumping effects, calculating well efficiency, and estimating ground-water flow through an area.

For an interference test to provide reliable values it is essential that (1) all wells involved tap the same and only the same aquifer or aquifers; (2) the observed wells are far enough from the pumped well for laminar flow to prevail; and (3) there are no extraneous pumping effects, such as starting or stopping of other pumps, that could influence the wells of the test.

It appears that some interference tests are made merely because the observation wells are available for measurement. Insufficient records may be at hand to assure that the foregoing requirements are met. If the data obtained from such tests are analyzed and used for

predictive purposes, there is a danger of underestimating or overestimating the amount of water available.

One way to improve the usefulness of interference-test data is to measure the drawdown and recovery in the pumped well itself. This was not done in 15 of the 76 interference tests. With the pumped-well measurements, the transmissivity indicated by the observation wells is verified (or refuted), and the additional values of specific capacity and well efficiency are provided.

In 11 of the 61 tests in which the pumped well was measured, the transmissivity indicated by the pumped well differed substantially (23 percent or more) from that indicated by the observation well(s). A careful examination of these 11 tests and the available well-construction data revealed or suggested, in each case, the reason(s) for the discrepancy. Usually it was a matter of the pumping well and observation well(s) not being completed in the same and only the same aquifer(s). In such situations the interference tests should not have been attempted.

Pumped-well and observation-well data provided corroborating transmissivity values in 50 tests. This was irrespective of the efficiency of the pumped well. For 25 of these tests the well efficiency was less than 100 percent; for several it was less than 50 percent.

Values for the storage coefficient can be calculated only from observation-well data. More than 70 percent of the values determined in the interference tests were in the range of 0.0001-0.0004; a full 30 percent were 0.0002. It is common for the storage coefficient calculated from the data for several observation wells in

the same test to vary through much of the range quoted above. It seems reasonable, therefore, to suggest that an estimated value for the storage coefficient may often be as reliable for predictive purposes as the one calculated from pumping-test data. A median value near 0.00025 would appear to be a safe and useful one that could be used in all except special situations, thus reducing the need for interference tests.

The conclusions that can be drawn from this analysis of the interference tests in the Coastal Plain aquifers of South Carolina are:

- (1) Great care should be exercised in selecting or constructing observation wells.
- (2) Drawdown and recovery should be measured in the pumped well if it is possible to do so.
- (3) Pumped-well measurements provide good transmissivity values. The values are not affected by well inefficiency and should be the same as those indicated by observation wells.
- (4) An estimated storage coefficient probably is adequate for predicting pumping effects, thus obviating the need, in many cases, for interference tests.

# HYDRAULIC CONDUCTIVITY OF THE PRINCIPAL CRETACEOUS AQUIFERS IN SOUTH CAROLINA

by

Roy Newcome, Jr.

## Abstract

Hydraulic conductivity, calculated from transmissivity revealed by pumping tests and thickness shown by electric logs, is generally greater for aquifers in the Middendorf Formation than for the Black Creek Formation in the Coastal Plain of South Carolina. Both formations contain confined aquifers of Upper Cretaceous age that are extensively tapped for water supplies. In both formations the hydraulic conductivity declines seaward.

The hydraulic conductivity of a clastic aquifer may be expressed as the number of gallons (or cubic feet) of water that would each day move through each square-foot vertical section of the aquifer at unit hydraulic gradient. Hydraulic conductivity (symbol K) is greatly influenced by the grain size and sorting and by the "cleanness" of the aquifer—that is, the degree of admixture of clay or other fine material. Temperature also has an influence insofar as it affects viscosity and density of the water.

The confined Cretaceous aquifers constitute the principal source of water supply for much of the Coastal Plain area of South Carolina. Two formations, the Middendorf, which occupies the lower part of the Cretaceous section, and the Black Creek, in the upper part, are extensively tapped in all but the southernmost counties (where the Floridan limestone aquifer is the almost exclusive source of water for wells). In many wells both Cretaceous formations are tapped.

The maps of Figure 22 show where the Black Creek and Middendorf Formations contain freshwater. One or both contain only saline water (dissolved-solids concentration greater than 1,000 milligrams per liter) in the immediate coastal area. The updip extent of the Black Creek Formation is generally 5-20 miles southeast of the Fall Line.

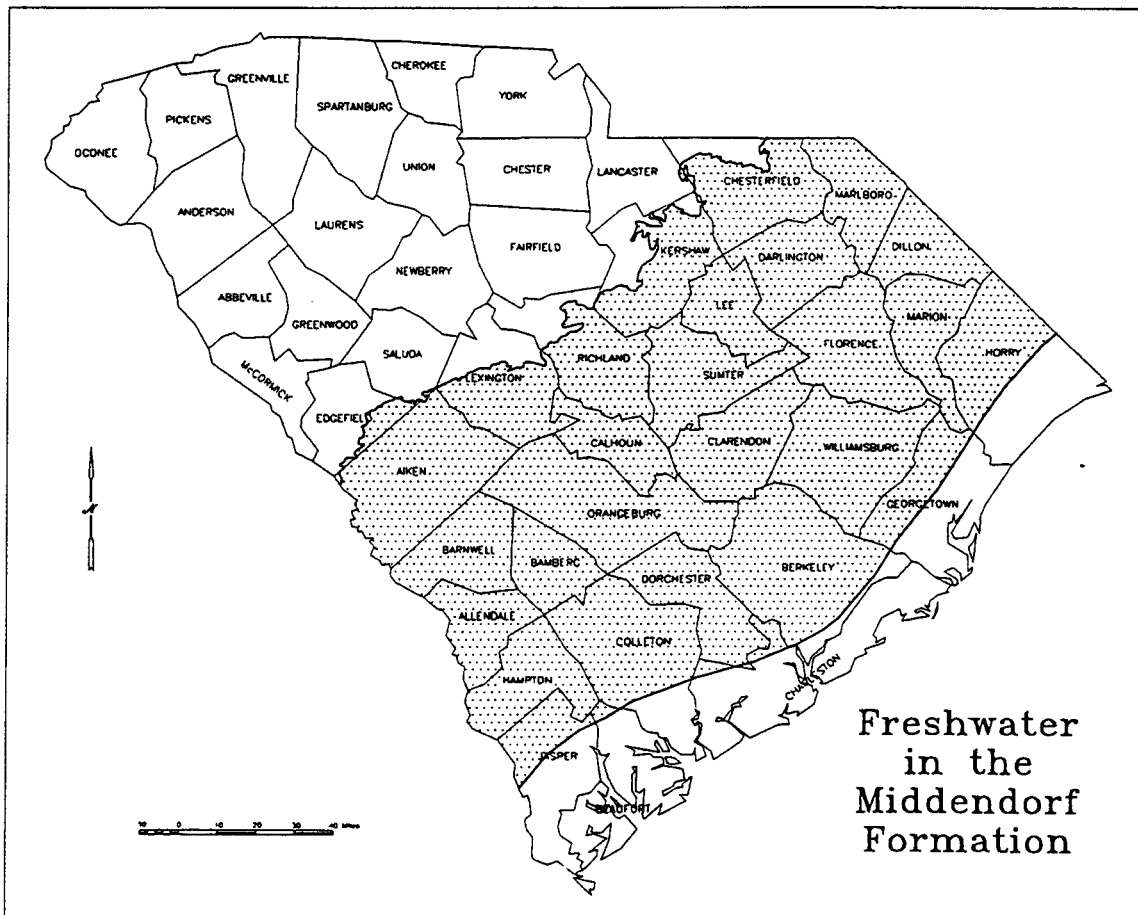
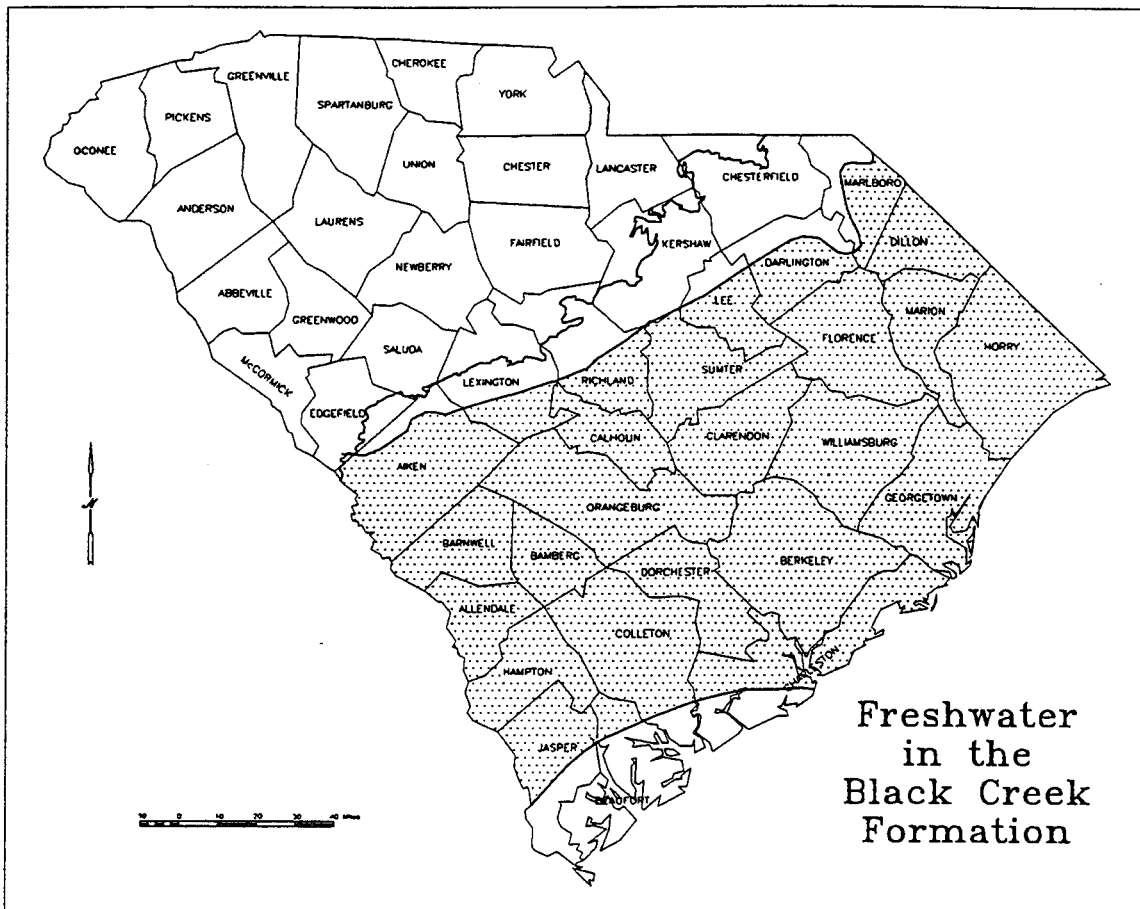
A reliable method of determining K is by means of a pumping test, in which the indicated value for transmissivity tells how much water would move through a 1-foot-wide section of the full aquifer thickness under unit hydraulic gradient. Dividing the transmissivity by the aquifer thickness provides the K. This presupposes that accurate information is available on aquifer thickness, and the best source for this is geophysical logs, specifically electric-resistivity logs.

About 230 pumping tests are available in the Department of Natural Resources files for wells producing water from either the Middendorf Formation or Black Creek Formation and for which reasonably definitive electric logs also are available. These data, studied together, permit the calculation of K and the comparison of these two important water-bearing formations in most of the Coastal Plain. Comparing the hydraulic properties of the two aquifers is more complex than it might seem at first inspection. In only a few counties are both tapped by a sufficient number of wells that the results of pumping tests are reasonably comparable. Most counties have wells in only one of the formations. In the places where both formations are available, most of the major wells tap both. Tests of those wells could not be used in this aquifer-comparison effort. The one county that has a good number of tests for both formations is Florence. There, 13 tests of the Middendorf and 10 tests of the Black Creek provide average K values of 250 gpd/ft<sup>2</sup> (gallons per day per square foot) for the Middendorf and 300 gpd/ft<sup>2</sup> for the Black Creek (Fig. 23). Florence County does not seem to typify the K relationship between the two formations. As indicated on Figure 23, in every other county with values for both formations the Middendorf has the higher average value for K.

Despite the paucity of data, the maps of Figure 23 suggest three important conclusions that apply to the Coastal Plain in South Carolina:

1. Hydraulic conductivity of the Middendorf aquifers is greater, on the average, than that of the Black Creek aquifers.
2. In both formations the hydraulic conductivity declines in a seaward direction.





**Figure 22. Showing where freshwater is available from the aquifers of the Black Creek and Middendorf Formations in South Carolina.**



3. In both formations the greatest hydraulic conductivity is in the northwestern part of the Coastal Plain, most specifically the area including Aiken and Orangeburg Counties, and the least is in the eastern part, mainly in Georgetown and Horry Counties.

Not revealed by the maps is the range in the values from which the average K was calculated for each county. The ranges (following table) illustrate vividly why any but the most general predictions of hydraulic conductivity would be unwarranted. Only the three generalizations enumerated above are of real importance.

Hydraulic conductivity is of critical importance in water-supply availability only as it combines with aquifer thickness to provide transmissivity. It is the last that determines how much water wells can produce. The available data suggest that the

Middendorf Formation has generally thicker aquifers than the Black Creek Formation. This, in turn, translates into higher transmissivity for the Middendorf, on the average.

This discussion has used the names Middendorf and Black Creek to refer to the principal water-bearing formations in the lower and upper parts of the Cretaceous section in South Carolina. In doing so, the writer has followed the division made by Colquhoun and others.<sup>1/</sup>

<sup>1/</sup>Colquhoun, D.J., and others, 1983, Surface and subsurface stratigraphy, structure and aquifers of the South Carolina Coastal Plain: Columbia, S.C., University of South Carolina, Department of Geology, 78 p.

<u>County</u>	<u>Range in K values (number of tests)</u>	
	<u>Middendorf Fm</u>	<u>Black Creek Fm</u>
Aiken	250-900 (6)	260-650 (2)
Allendale	-	250-475 (2)
Barnwell	-	600-670 (2)
Berkeley	250-390 (2)	100 (1)
Calhoun	210 (1)	280-320 (2)
Charleston	-	54-490 (14)
Chesterfield	74-650 (6)	-
Clarendon	380 (1)	300-350 (2)
Darlington	100-840 (11)	-
Dillon	140-350 (5)	-
Dorchester	110-400 (3)	-
Florence	100-500 (13)	80-460 (14)
Georgetown	-	40-190 (19)
Horry	-	46-390 (61)
Kershaw	50-540 (9)	-
Lee	150-700 (4)	-
Lexington	120-830 (8)	-
Marion	-	72-350 (10)
Marlboro	80-620 (4)	-
Orangeburg	640-1500 (3)	580-690 (3)
Richland	240-600 (6)	200 (1)
Sumter	235-400 (3)	130-550 (7)
Williamsburg	80-760 (2)	56-180 (5)