Characterization and Modeling of Land Subsidence due to Groundwater Withdrawals from the Confined Aquifers of the Virginia Coastal Plain

Jason P. Pope

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Thomas J. Burbey, Chair
George E. Harlow, Jr.
Madeline E. Schreiber

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Abstract

Measurement and analysis of aquifer-system compaction have been used to characterize aquifer and confining unit properties when other techniques such as flow modeling have been ineffective at adequately quantifying storage properties or matching historical water levels in environments experiencing land subsidence. In the southeastern Coastal Plain of Virginia, high-sensitivity borehole pipe extensometers were used to measure 24.2 mm of total compaction at Franklin from 1979 to 1995 (an average of 1.5 mm/yr) and 50.2 mm of total compaction at Suffolk from 1982 to 1995 (an average of 3.7 mm/yr). Analysis of the extensometer data reveals that the small rates of aquifer-system compaction appear to be correlated with withdrawals of water from confined aquifers. One-dimensional vertical compaction modeling indicates that the measured compaction is the result of nonrecoverable hydrodynamic consolidation of the fine-grained confining units and interbeds as well as recoverable compaction and expansion of coarse-grained aquifer units. The modeling results also provide useful information about specific storage and vertical hydraulic conductivity of individual hydrogeologic units. The results of this study enhance the understanding of the complex Coastal Plain aquifer system and will be useful in future modeling and management of ground water in this region.

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The final completion of this thesis has taken me much longer than expected, and its effect on my life has been much larger than I would have ever imagined. Nonetheless, I am happy with where this effort has led me. I have had the opportunity to meet and learn from many people to whom I owe my thanks. It has been an honor to work with and learn from so many interesting and intelligent folks. I hope that most of these people know who they are, but I want to use this opportunity to mention a few of them by name.

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[*] Dunn, Stephen, 1989, "The Storyteller," Between Angels, New York: W.W. Norton and Company.

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I. INTRODUCTION

Withdrawals of groundwater from the confined aquifers of the Virginia Coastal Plain have increased dramatically over the past century, resulting in large declines in water levels, associated compaction of the aquifer system, and gradual subsidence of the land surface. Subsidence in this region was first identified in the 1970s through a high-precision releveling survey (Holdahl and Morrison, 1974), which measured several millimeters per year of localized vertical movement of the land surface. In an attempt to learn more about the nature of this subsidence and investigate its relationship to groundwater withdrawals, the United States Geological Survey (USGS) and the Virginia Water Control Board drilled and instrumented two compaction wells with borehole pipe extensometers in southeastern Virginia in the late 1970s and early 1980s. The two extensometers were installed near the cities of Franklin and Suffolk, in close proximity to the centers of the greatest groundwater withdrawals. These instruments provided over 16 years of compaction data before they were removed from service at the end of 1995. Over that period, the extensometers measured small rates of land subsidence (between 1 and 4 millimeters per year) apparently related to water-level changes.

Little attention has been given to the compaction data or the issue of regional land subsidence, perhaps because of its small magnitude, which was not noticed by most observers. However, research in other regions experiencing land subsidence has demonstrated that an understanding of the relations between groundwater withdrawals, hydraulic heads, and compaction in an aquifer system reveals fundamental information about the properties and behavior of the system (Galloway and others, 1999). This information is important to current efforts by the USGS to produce an accurate groundwater flow model for the Virginia Coastal Plain, which will be useful for the management of the limited water resources of this region's large and rapidly-growing population.

The communities of southeastern Virginia depend on large withdrawals of groundwater, which currently exceed rates of natural recharge in many areas and are expected to continue to increase with growing populations. In light of these continued withdrawals, there is also concern in coastal areas about the effects of land subsidence on the local rate of (relative) sea-level rise, which is approximately twice the magnitude measured elsewhere in the eastern United States. As a result, information about the processes of aquifer-system compaction in this region should be useful to those concerned with the proper management of groundwater resources.

Purpose and Scope

The purpose of this research is to thoroughly describe the nature and extent of aquifer-system compaction and land subsidence in the Coastal Plain of southeastern Virginia. This involves the presentation of extensometer data documenting aquifer-system compaction in the Coastal Plain of southeastern Virginia, along with related data on groundwater withdrawals, groundwater levels, and land subsidence. This project also includes the results of numerical simulations and statistical analyses of aquifer-system compaction. Direct measurements and model simulations of compaction in the study area are used to refine previous estimates of properties that govern compaction of the aquifer system and its components. This task includes the accurate calculation of elastic and inelastic specific storage values for individual coarse-grained aquifers and fine-grained confining units or aquitards, as well as estimation of the time responses of confining units to current and future changes in stress.

This study is focused on aquifer-system compaction measurements and groundwater-level data at the extensometer sites at Franklin and Suffolk, Virginia. However, the study area includes all of the southeastern Virginia Coastal Plain because compaction is not confined to the two measurement locations and is likely influenced by groundwater processes and geologic properties throughout the area experiencing declines in groundwater levels. Furthermore, results from this study will be incorporated into regional models of groundwater flow currently being developed and revised by the USGS for the entire Coastal Plain of Virgnia (McFarland, 1998).

More than 15 years of aquifer-system compaction measurements from the unconsolidated Coastal Plain sediments, collected from the two pipe extensometers, were analyzed for this study. Almost 100 years of groundwater hydrographs from numerous wells in the study area were also analyzed, along with withdrawal records for all of the aquifers of interest in the study area. Statistical methods are used to describe the groundwater-level data (stress) and the aquifer-system compaction data (strain), as well as the correlation between the stress and strain records.

The simulations of compaction are based on the established model of aquitard drainage developed by Tolman and Poland (1940). They were carried out using a one-dimensional computer model of time-dependent ground movement due to changes in hydraulic head (a program known as COMPAC) developed by Helm (1975). The COMPAC model is based on the physical properties of aquifers and aquitards and the changes in stress due to the fluctuation of water levels in those units. Groundwater-level data and early land subsidence data were used to calibrate simulations of aquifer-system compaction at Franklin and Suffolk throughout the past century. Simulations for the period 1900-1995 were partially constrained by the availability of historical water-level data, though measured subsidence data from the two extensometers were used to facilitate more precise calibration for recent years (1979-1995).

One-dimensional modeling results were applied to the entire Coastal Plain of Virginia to develop estimates of the area affected by land subsidence due to groundwater withdrawals and to predict which areas are likely to be affected in the future. In terms of cultural and environmental significance, coastal areas are most vulnerable to deleterious effects from land subsidence, and recent data were examined to investigate the possible influence of land subsidence on anomalous rates of relative sea-level rise in coastal areas of southeastern Virginia.

Terminology

Subsidence is most simply defined as sinking or downward settling of the earth's surface (Bates and Jackson, 1984). Of course, subsidence can occur naturally in a sedimentary basin, but it is often induced by human activity such as withdrawal of water or oil. The term compaction is defined by Poland and others (1972) as a decrease in thickness of a sedimentary layer or unit in response to an increase in vertical compressive stress, which can be caused by the withdrawal of fluids. For the purposes of this research, subsidence is considered to be the vertical displacement of the land surface resulting from compaction within an aquifer system.

An aquifer is a body of rock or sediment sufficiently permeable to conduct groundwater and yield significant quantities of water (Bates and Jackson, 1984). In contrast, a confining unit is defined as a body of impermeable or distinctly less permeable material stratigraphically adjacent to an aquifer unit. The lower permeability of a confining unit typically results from the finer-grained sediments of which it is composed. An aquitard is defined as a "leaky confining bed," or one that retards but does not prevent the flow of water to or from an adjacent aquifer (Bates and Jackson, 1984). For the purposes of this report, fine-grained units that are interpreted as separating distinct aquifer units are referred to as confining units or confining beds, while the generally thinner fine-grained layers occurring within aquifer units are referred to as aquitards or interbeds. While aquitards and confining units do not transmit appreciable amounts of water, they may be important in the storage of water in an aquifer system. An aquifer system, such as the one defined in this study, includes a series of aquifers, aquitards and confining units.

Compaction includes both instantaneous and time-dependent deformation (Epstein, 1987). This deformation may be nonrecoverable and inelastic (or virgin), or it may be recoverable and approximately elastic. Nonrecoverable compaction occurs when the past maximum effective-stress (expressed as water-level decline) is exceeded, and it is proportional to the logarithm of that increase (Hanson, 1989). As long as the change in effective stress is less than the previous maximum effective stress (i.e. as long as the previous maximum drawdown is not exceeded), the compaction is recoverable and is called recoverable compaction (Poland et al., 1972). It should be noted that this 'elastic' compaction is not necessarily an instantaneous linear response to a change in effective stress; it is given this designation simply because it is fully recoverable (Hanson, 1989). Total nonrecoverable and recoverable compaction lag behind each increase in effective stress because of the impedance to groundwater outflow as pore space is reduced (Hanson, 1989). The time required for an aquifer unit to reach near (93 percent) equilibrium again after a change in effective stress is known as the time constant for that unit (Riley, 1984).

Location of Study Area and Measurement Sites

The Coastal Plain physiographic province of Virginia encompasses approximately the eastern third of the state (figure 1-1). It is bounded on the west by the Fall Line, which separates the Coastal Plain from the Piedmont province, and it is bounded on the east by the Atlantic Ocean. The spatial scope of this research is limited primarily to the southeastern section of the Coastal Plain in Virginia, which includes an area of approximately 3,850 square miles (10,000 square kilometers) bounded by the James River, the Chesapeake Bay and the Atlantic Ocean, the North Carolina border, and the Fall Line. This specific study area was chosen in order to correspond to previous groundwater modeling efforts by the U.S. Geological Survey (USGS) (Hamilton and Larson, 1988), and to constrain variations in the geologic framework as described by Powars (2000).

Subsidence due to water-level decline is likely occurring throughout the study area, but one-dimensional compaction data are available only at two extensometer sites, located in the cities of Franklin and Suffolk. The City of Franklin is located in the southeastern Coastal Plain of

Virginia, in southern Southamption County and across the Blackwater River from Isle of Wight County. Franklin is approximately 10 miles (16 km) from the North Carolina border and just over 50 miles (80 km) from the Atlantic Ocean. The City of Suffolk is located approximately 20 miles (32 km) north-northeast of Franklin, in the central part of Suffolk City County, approximately 10 miles (16 km) south of the James River and 45 miles (72 km) from the Atlantic Ocean.

Development of Groundwater Resources and Resulting Land Subsidence

Significant withdrawals from the aquifers of the Virginia Coastal Plain began in the late 1800s, as flowing (artesian) wells were tapped for agricultural and industrial uses. Before these earliest developments, water levels at Franklin exceeded 40 feet (12.2 meters) above sea level, according to several sources (Hamilton and Larson, 1988). Continued pumping at rates in excess of natural recharge caused moderate declines in water levels throughout the Coastal Plain, but water levels in the Cretaceous aquifers (see figure 3-2 for stratigraphic and hydrogeologic units) at Franklin still approached 20 feet (6.1 meters) above sea level in 1939 (Cederstrom, 1945). By that time, the growing volumes of water withdrawn at Franklin had caused the development of a small cone of depression in the piezometric surface of the Cretaceous aquifers at that location. However, wells tapping the aquifer continued to flow until the early 1940s (Cederstrom, 1945). Sharp increases in industrial and municipal withdrawals near Franklin, particularly from the Cretaceous aquifers, began in the early 1940s and continued for several decades before leveling off in the 1970s. These withdrawals were accompanied by rapid increases in industrial, municipal and agricultural withdrawals throughout the Coastal Plain (Kull and Laczniak, 1987). Despite increasing withdrawals elsewhere, the large volumes withdrawn from the Cretaceous aquifers in the vicinity of Franklin have dominated the hydrogeology of the Coastal Plain, lowering water levels in the middle Cretaceous aquifer at Franklin to over 180 feet (55 meters) below sea level. Overall, withdrawals from the Cretaceous aquifers in close proximity to Franklin increased from less than 5 million gallons per day in the early 1930s to the current rate of more than 35 million gallons per day (Harsh and Laczniak, 1990). This increase has resulted in a total water-level decline of over 200 feet (61 meters) in the Cretaceous Middle Potomac aquifer and the expansion of a regional cone of depression to the north across the Virginia Coastal Plain and south into North Carolina (figure 1-2).

The pattern of groundwater withdrawals at Suffolk has been similar to that observed at Franklin, though the magnitude of the withdrawals and their effects have been much smaller. Pumping from the Cretaceous aquifers at Suffolk increased from less than 1 million gallons per day in the early part of the century to the current rate of almost 15 million gallons per day. These withdrawals, along with the influence of the Franklin withdrawals, have resulted in a total water-level decline of approximately 100 feet (161 m) in the Middle Potomac aquifer at Suffolk.

Withdrawals from the Cretaceous aquifers at other locations have also contributed to these regional declines in head. Large industrial withdrawals (over 10 million gallons per day) over the last several decades at West Point, in eastern King William County, have signficantly affected regional water levels and influenced the cone of depression. In addition, growing populations across the Virginia Coastal Plain have led to increases in withdrawals for municipal, domestic, and agricultural purposes. Though distributed relatively evenly across a wide area, the collective effects of these withdrawals are substantial.

Land subsidence was first reported in the Coastal Plain of Virginia in 1974 (Holdahl and Morrison), though it had likely been occurring for decades prior to that report. From an analysis of first-order releveling and mareograph data, Holdahl and Morrison found an area of anomalous subsidence centered approximately at Franklin, Virginia. Though most of the Atlantic Coastal Plain is subsiding to some degree due to minor crustal movements, Holdahl and Morrison noticed significant local variations in subsidence in the Chesapeake Bay area (figure 1-3). In particular, they measured a subsidence rate of over 4 millimeters per year near Franklin between 1942 and 1971, which was much greater than the mean rate across the bay area. Holdahl and Morrison did not provide an explanation for their findings, but the extent of the subsiding area reported in their study corresponds well with the cones of depression in the Cretaceous aquifers. In fact, the known pumping centers in the Virginia Coastal Plain correspond with the areas of differential subsidence reported by Holdahl and Morrison, providing further evidence that the subsidence could be the result of groundwater withdrawal.

Since the early 1970s, the rate of groundwater withdrawal from the aquifers of the Coastal Plain has increased much more slowly, with expected yearly and seasonal fluctuations (Kull and Laczniak, 1987). As a result, water levels have stabilized somewhat in the areas of largest historical hydraulic head declines. While water levels vary from year to year due to variations in withdrawal patterns and recharge rates, only minor declines in water levels in the

Cretaceous aquifers have been observed at most locations in recent years. Perhaps as a result, rates of land subsidence appeared to decrease at Franklin in the last few years that measurements were available. However, marked increases in subsidence have been observed when groundwater levels have dropped below past maximum levels, demonstrating increased potential for future subsidence as groundwater resources are further developed. Furthermore, water levels continue to fall at Suffolk and other locations where withdrawal rates are increasing, leading to continuing land subsidence.

Because of the relatively small rates and magnitudes of compaction reported, problems related to subsidence in the Virginia Coastal Plain have gone almost unnoticed until recently. Nonetheless, the effects of compaction in the aquifer system continue to be significant. When the land area affected by subsidence is considered, even a small amount of compaction translates into a large volume of water. As withdrawals continue to increase to sustain future demands for fresh water by growing municipalities, factors related to aquifer-system compaction may affect the development of future supplies due to permanent reductions in aquifer-system storage. These problems will likely be accompanied by higher rates of coastal inundation as rising sea levels have even larger effects on areas experiencing subsidence due to groundwater withdrawals.

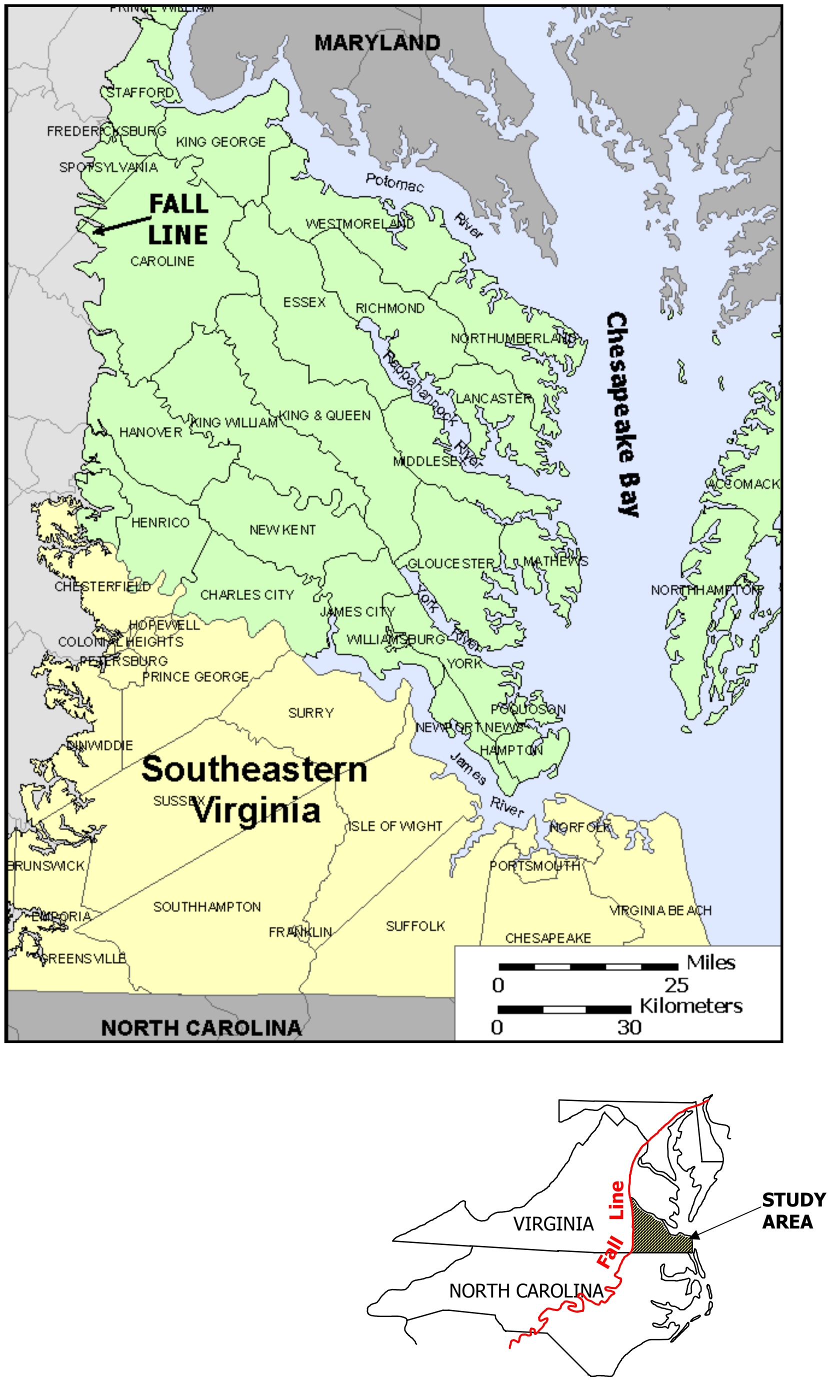


Figure 1-1: Map of study area - the Coastal Plain of Virginia.

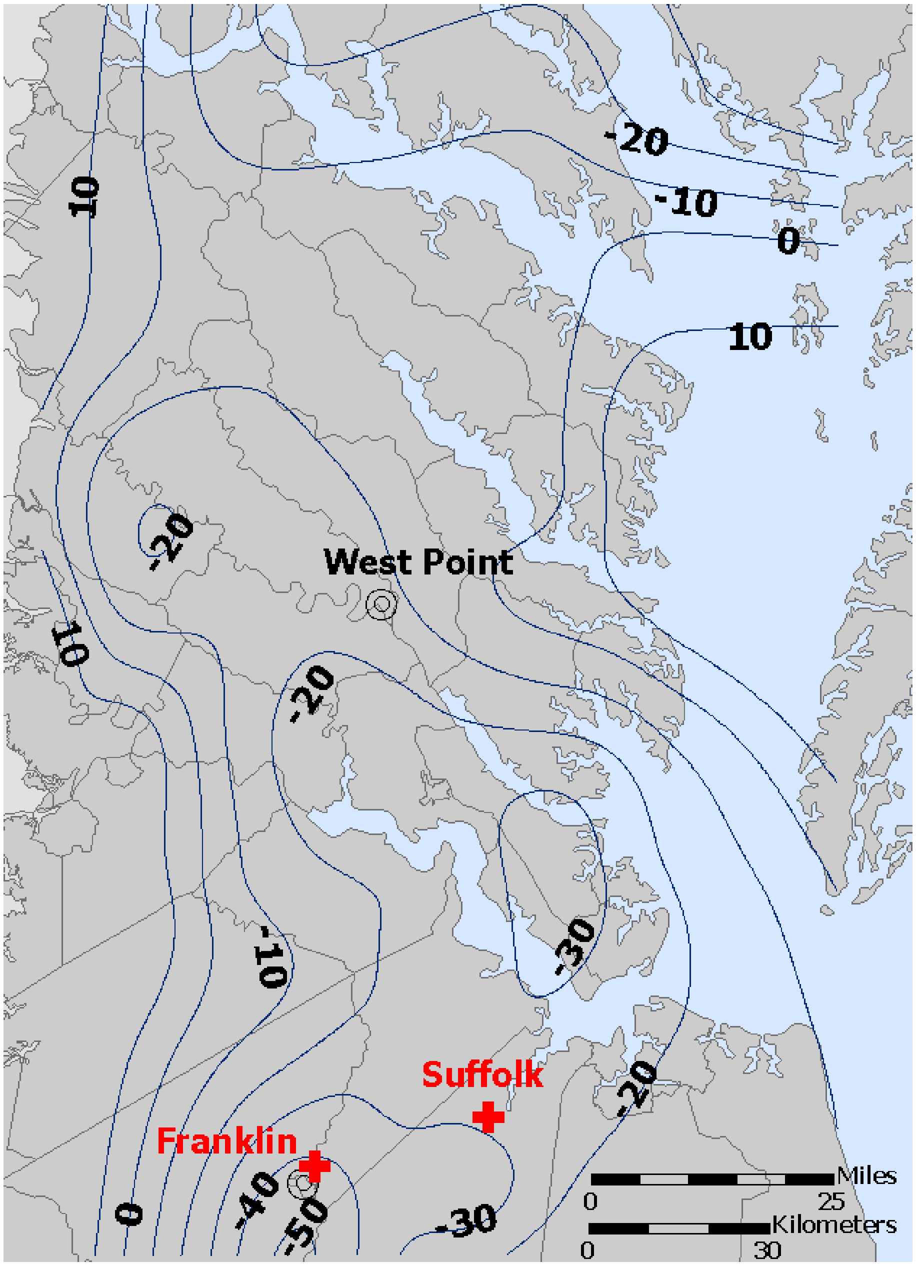


Figure 1-2: Potentiometric surface of the Middle Potomac aquifer in 1998, developed from USGS water-level data. Hydraulic head values have units of meters relative to the sea-level datum. Red crosses indicate extensometer locations.

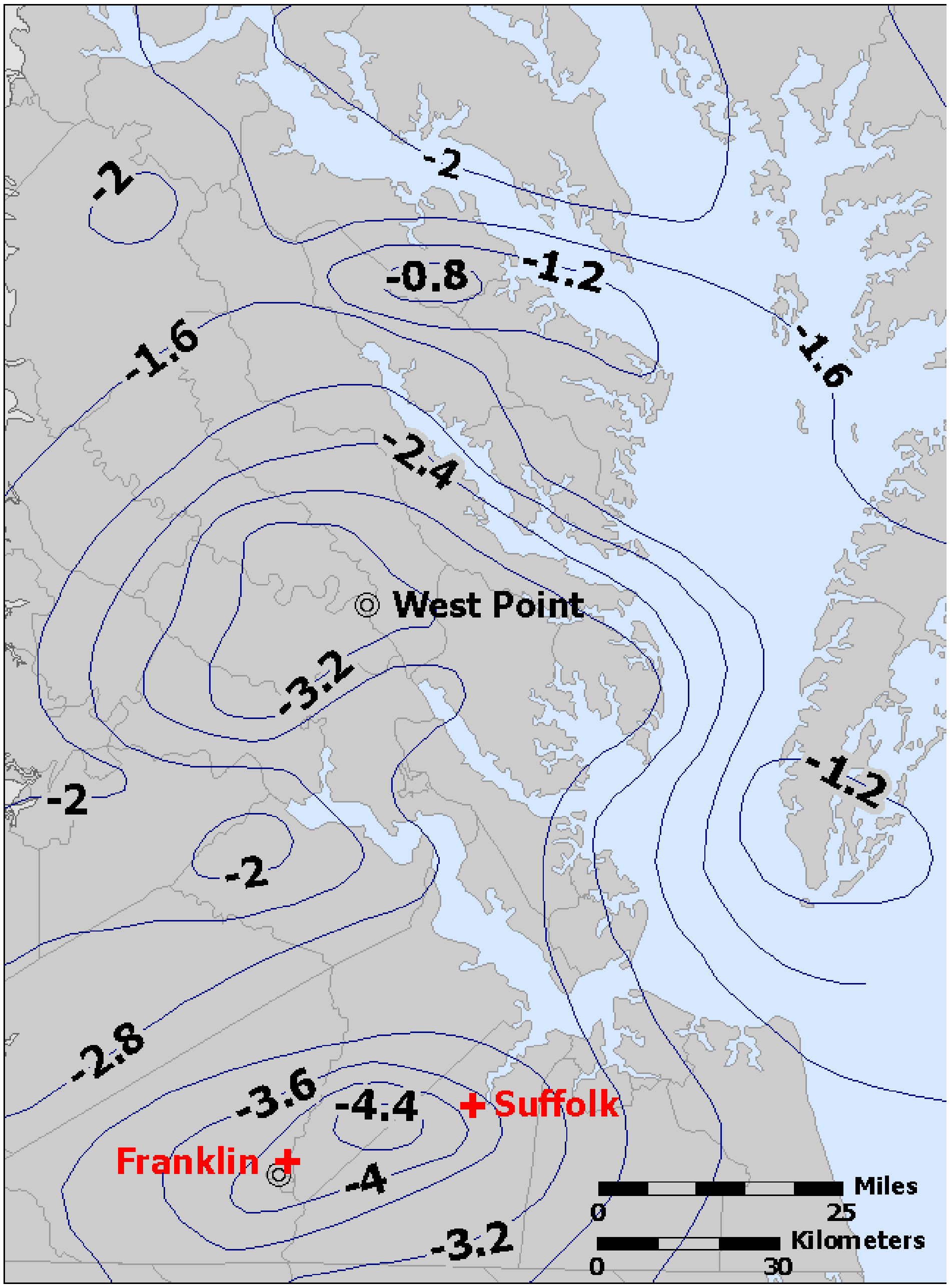


Figure 1-3: Rate of vertical land surface movement (mm/year), adapted from Holdahl and Morrison (1974).

II: ANALYTICAL APPROACH FOR AQUIFER-SYSTEM COMPACTION

Land subsidence due to the withdrawal of groundwater has a number of known causes, but the mechanism considered here is compaction of the aquifer system. The process of aquifer-system compaction is considered to be fairly well understood, although recent research has focused on new aspects of the problem, such as deformation in three dimensions. Most simply, the removal of groundwater from an aquifer or aquifer system causes a reduction in the fluid pressure in the pores of the granular matrix. This loss of fluid pressure results in stress on the aquifer matrix (skeleton), because the weight of the overlying materials supported by the skeleton does not typically change significantly. The increasing stress on the aquifer skeleton causes it to deform, particularly if the formation is unconsolidated or only partially consolidated. The skeleton deforms or compacts in three dimensions, but deformation in the vertical direction is the most noticeable and easiest to measure. This vertical compaction of the aquifer system is usually expressed as subsidence of the land surface. The nature of the compaction and subsidence in an aquifer system is controlled by a number of variables, including the patterns of withdrawals and associated water-level declines, as well as the compressibility, permeability, storage coefficient, and thickness of compacting units. While some recoverable compaction occurs in coarse-grained aquifer units, most nonrecoverable compaction occurs as the result of slow, irreversible consolidation of fine-grained confining units or aquitards. The understanding of this process, known as the aquitard-drainage model, is the foundation of most research in subsidence (Galloway and others, 1999).

Principle of Effective Stress and One-Dimensional Compaction

The development of the theoretical relationship between groundwater levels and compaction of the aquifer system begins with the principle of effective stress originally proposed by Terzaghi (1925, 1948) and represented in figure 2-1. In his theory of consolidation, Terzaghi (1948) developed an expression to describe the vertical stress on a horizontal plane at any depth below land surface in a system at equilibrium. His simple yet powerful relation is expressed as

$$\sigma_T = \sigma_e + P \qquad (2\text{-}1)$$

where σ_T is the total stress due to the geostatic load [N/m^2],

σ_e is the effective or intergranular stress [N/m^2], and

P is the pore-water stress [N/m^2].

Effective stress is defined by Lofgren (1969) as "the grain-to-grain stress which effectively changes the void ratio and mechanical properties of a deposit." Because the analysis of effective stress is the most important consideration in studies of aquifer-system compaction and subsidence, Terzaghi's equation is simply rearranged, by Helm (1975), to solve for effective stress:

$$\sigma_e = \sigma_T - P \tag{2-2}$$

This form of the equation explicitly reveals that the effective stress on the aquifer-system can be increased by increasing the total stress (increasing the overburden) or by removing water and decreasing the pore water stress (Helm, 1975). Of course, both of these phenomena may occur together, and their effects are additive (Helm, 1975). This relation is outlined in greater detail by Hanson (1989), who provides a means for calculating the components of the opposing stresses at any depth z in a vertical column of unit area. The total stress is divided into three separate terms (Hanson, 1989):

$$\sigma_T = \int_0^{z_w} Sn\rho_w g dz + \int_{z_w}^{z} n\rho_w g dz + \int_0^{z} (1-n)G\rho_w g dz \tag{2-3}$$

Where z_w is the depth below land surface to water table [m],

S is the degree of saturation above water table [dimensionless],

n is the average porosity [dimensionless],

ρ_w is the density of water [kg/m^3],

Z is depth below land surface [m],

G is the specific gravity of solid grains in aquifer system [dimensionless].

The first term quantifies the weight of the water above the water table, the second term represents the weight of the water below the water table, and the final term is the weight of the sediments in the column. The pore-water pressure component of the effective stress, which supplies an upward buoyant force in the system, can be expressed as (Hanson, 1989):

$$P = -\int_{z_w}^{z} \rho_w g dz \tag{2-4}$$

This analysis demonstrates why stresses must be considered differently depending on whether water is removed from a confined or unconfined aquifer. The removal of water from an unconfined aquifer typically results in a change in the both the total stress (overburden) and the pore water stress, while the removal of water from a confined aquifer usually results in a change in the pore water stress only, for reasons explained below (Poland and others, 1975; Helm, 1975).

Stress or pressure, typically expressed in units of $ML^{-1}T^{-2}$, can also be expressed in terms of equivalent hydraulic head (pressure head), with units of length L, by the following relation (Sneed and Galloway, 2000):

$$h = \frac{P}{\rho_w g} \qquad (2\text{-}5)$$

This conventional expression of stress as equivalent hydraulic head is valid given the following assumptions: constant gravity and a uniform incompressible (negligibly compressible) fluid (Sneed and Galloway, 2000). Under these standard conditions, one foot of water is equivalent to a pressure of 0.433 lb/in^2. This convention allows a convenient comparison of the magnitudes of various stresses for groundwater systems.

The rise and fall of the water table in an unconfined aquifer causes changes in gravitational stress by removing or supplying fluid (thereby changing the total stress, σ_T) and by increasing or decreasing the hydrostatic pressure in the aquifer (thereby changing the pore-water stress, P). Within the unconfined aquifer, these two forces change with opposite signs. For example, a water-table decline results in a decrease in the total stress in the unconfined aquifer, which tends to decrease the effective stress; however, the same decline reduces the pore pressure in the aquifer, leading to an increase in effective stress. For the unconfined aquifer, the net change is an increase in effective stress (Lofgren, 1968). Specifically, the effective stress change in an unconfined system can be calculated (Epstein, 1987) as change in weight per unit area:

$$\Delta\sigma_e = \Delta w = \Delta h(\gamma_m - \gamma_b) \qquad (2\text{-}6)$$

Where Δh is the change in water table height [m],

γ_m is the unsaturated unit weight of sediments above depth z,

γ_b is the buoyant unit weight of sediments above depth z.

Such an increase in effective stress in the unconfined system may cause some recoverable compaction in the aquifer, and it may also induce a small amount of nonrecoverable compaction in fine-grained interbeds within the unconfined system. However, compaction in unconfined aquifers is usually of relatively small magnitude, because the net stress change is small for a given water-table decline, and unconfined aquifers are usually relatively thin, leading to small water-level declines. Of greater significance to compaction is the effect of rising and falling water tables in an unconfined aquifer on the change in total stress on the aquifer system, including confined aquifers below. This discussion of stresses is included here for completeness, but the unconfined system is negligible, and it is disregarded in the analysis of aquifer system stresses and the resulting compaction in the Virginia Coastal Plain. The stress changes in the confined units are of much greater importance.

In a confined aquifer system, changes in effective stress almost always result from changes in the pore water stress. Unless dewatering (drainage of the aquifer pores) occurs, the lowering of head in a confined aquifer system does not significantly change the gravitational stress, because the amount of water withdrawn per unit area is negligible. The associated change in fluid pressure in the pores of the aquifer system is much more significant (Poland and others, 1972). Thus, the increase in effective stress in a confined aquifer system is approximately equal to the decrease in the pore-water pressure (Poland and others, 1972; Helm, 1975). Furthermore, Terzaghi's equation for effective stress is developed under the assumption that sediment-volume reduction through vertical compaction in an aquifer system is equal to the volume of water expelled (Hanson, 1989). The nature of this relation depends on the distribution of aquifers and confining units and interbeds in the aquifer system.

Lowering the hydraulic head in a confined aquifer results in a decrease in fluid pressure in the pores of the aquifer, which is equivalent to an increase in effective stress. While the total stress on the system does not change, stress is shifted from the pore water to the skeleton as pore pressure is reduced. In incremental terms, this relation can be expressed as

$$\Delta\sigma_e = -\Delta P \tag{2-7}$$

The increase in effective stress results in compaction of the aquifer and a slight reduction in porosity as the granular skeleton is compressed and rearranged. It should be noted that the granular skeleton of the aquifer system is considered to be much more compressible than the individual grains (Helm, 1975), and the compressibility of the grains is therefore neglected in this analysis. Compaction that occurs in an aquifer responds immediately to a decrease in fluid pressure, and it may be recovered quickly with a restoration of fluid pressure (Poland and others, 1972). Furthermore, this compaction is usually relatively small in magnitude, because the coarse, angular grains cannot be placed in a much more compact arrangement.

The situation is much different in a fine-grained confining unit or aquitard adjacent to an aquifer in which hydraulic head is lowered. A decrease in pore-water stress in the aquifer is equal to an increase in effective stress at the aquifer-aquitard boundary (Hanson, 1989), but the dissipation of this stress by the expulsion of pore water is not immediate. Pore fluid pressure can only equilibrate in the confining unit as quickly as water can flow out (Lofgren, 1968). A fine-grained confining unit typically has a low vertical hydraulic conductivity and a relatively high specific storage. As a result, the vertical escape of water and the adjustment of pore pressures from a clay unit is typically very slow compared to the response of an aquifer (Poland and others, 1972). Thus, the increase in stress, and the associated compaction, take effect only as the pore pressure in the confining unit or aquitard slowly declines toward that in the aquifer (Poland and others, 1972). Some of the compaction that occurs in a fine-grained unit is recoverable or reversible, but a large part of it results from an irreversible rearrangement of clay grains (Lofgren, 1968). This mechanism, along with the typically higher porosity of clays, explains why most compaction in a system occurs primarily in the fine-grained interbeds and confining units. This process is the basis of the theory of hydrodynamic consolidation developed by Terzaghi (1925) and illustrated in figure 2-2.

Elastic and Inelastic Compressibility (Specific Storage)

An understanding of the relation between stress (head decline) and strain (compaction) is an integral part of subsidence analysis, as suggested by the discussion above. In fact, analysis of the compaction of a geologic unit relies on the stress-compaction relation, which is best quantified by the specific storage (S_s) of that unit (Jacob, 1940, 1950; Cooper, 1966):

$$S_s = \rho_w g(\alpha + n\beta) \qquad (2\text{-}8)$$

where ρ_w is the density of water [kg/m^3],

g is the acceleration due to gravity [m/s^2],

α is the compressibility of the aquifer matrix [m^2/N],

n is the porosity [dimensionless],

β is the compressibility of water [m^2/N].

Specific storage is the amount of water per unit volume of a saturated formation that is stored or expelled from storage owing to compressibility of both the mineral skeleton and the pore water per unit change in head (Fetter, 1994). Because the compressibility of water is assumed to be negligible compared to the compressibility of the skeleton in most compacting systems, the specific storage equation can be further simplified:

$$S^*_{sk} = \alpha \rho_w g \quad (2\text{-}9)$$

This equation gives the portion of specific storage derived from the expansion or compression of the skeleton and is usually called S^*_{sk} to indicate that it is the skeletal component of specific storage. The asterisk indicates that this is the overall specific storage of the aquifer system, including both aquifers and confining units, following the convention of Sneed and Galloway (2000). As the equation demonstrates, the compressibility term is the important link in relating stress to strain. The coefficient of compressibility (α) can be defined as the empirical relation between the loss of void space and the increase in effective stress (Hanson, 1989). Because the compressibility of water is not considered here, this coefficient of compressibility should formally be known as α^*_k, or the skeletal component of compressibility. The coefficient of compressibility can also be defined as the empirical relation between the loss of void space and the increase in effective stress (Hanson, 1989)

$$\alpha^* = {-\Delta e} \big/ {-\Delta \sigma_e} \quad (2\text{-}10)$$

Where e is the average void ratio (e = volume of pore space/volume of solids).

If the deformation of an aquifer or confining unit is considered to be entirely vertical, the loss of void space can be expressed is terms of thickness. Formally, the relation is given between vertical strain and the change in the average void ratio (Helm, 1975):

$$\frac{\Delta b}{b_0} = \frac{-\Delta e}{1+e} \quad (2\text{-}11)$$

where Δb is vertical compaction [m],

b is the initial confining unit thickness [m],

$\Delta b/b_0$ is the mean vertical strain, and

Δe is the average change in void ratio over the vertical column.

The expression of the change in void space as the change in vertical thickness is important because the vertical change in thickness is the strain parameter most conveniently measured. It has been shown previously that the change in effective stress can be measured as the change in pore pressure, expressed in terms of hydraulic head. As a result, two readily measurable parameters are available with which to characterize the stress/strain relation. In terms of specific storage, this relation can now be expressed as follows (Hanson, 1989):

$$S*_{sk} = \rho g \frac{(\Delta b / b_0)}{\Delta \sigma_e} \tag{2-12}$$

In this discussion of the stress/strain (or water-level/compaction) relation, a distinction has not yet been made between the behavior of coarse-grained aquifer material and the behavior of the fine-grained aquitard or confining unit material. As the earlier discussion indicated, however, this distinction is vital for a full understanding of the behavior of the system. Early researchers in subsidence recognized that the compaction of an aquifer system could be represented as the combined effect of the short-term, recoverable (elastic) response of coarse-grained aquifer material and the delayed, non-recoverable (inelastic) response of fine-grained confining unit or aquitard material. This theoretical approach involves a number of assumptions, and results in some significant limitations (which will be discussed later), but it has been the basis of most subsidence research.

When the stress on an aquifer system is less than any past maximum stress (drawdown) it has experienced, the system responds to water-level changes by compacting and expanding almost instantaneously in a recoverable or elastic manner. On the other hand, when the stress exceeds the past maximum level, the system experiences non-recoverable compaction that is typically of greater magnitude. These two processes are usually represented by the recoverable and nonrecoverable components of skeletal specific storage, or S_{ske} and S_{skv} (Sneed and Galloway, 2000):

$$S*_{sk} = \begin{cases} S*_{ske} = \alpha*_{ke}\ \rho g & when \quad \sigma_e < \sigma_{e\,(\max)} \\ S*_{skv} = \alpha*_{kv}\ \rho g & when \quad \sigma_e > \sigma_{e\,(\max)} \end{cases} \tag{2-13}$$

The 'e' subscript here signifies the recoverable, or elastic component, while the 'v' subscript refers to the nonrecoverable component, sometimes called the virgin component. As before, the asterisks (*) indicate these are average specific storage values across the total aquifer system.

As the previous discussion indicates, however, aquifers and confining units respond quite differently and should be characterized separately. In terms of notation, parameter values for confining units or aquitards are typically indicated by primes ('). Thus, the specific storage of a confining unit is given by (Sneed and Galloway, 2000):

$$S'_{sk} = \begin{cases} S'_{skv} = \alpha'_{kv}\,\rho g & when \quad \sigma_e > \sigma_{e\,(\max)} \\ S'_{ske} = \alpha'_{ke}\,\rho g & when \quad \sigma_e < \sigma_{e\,(\max)} \end{cases} \tag{2-14}$$

where α' is the compressibility of the confining unit only.

For applied stress (drawdown) less than the past maximum, the water released from storage in the confining unit (S'_{ske}) comes from the recoverable deformation of the fine-grained skeleton due to the reduction in pore pressure (Sneed and Galloway, 2000). When the stress exceeds past maximum levels, the water released from storage (S'_{skv}) comes from nonrecoverable compression of the fine-grained confining unit skeleton due to the rearrangement and reorientation of the clay particles. This nonrecoverable component (S'_{skv}) has been found to be 30 to several hundred times larger than the elastic component (S'_{ske}) in typical aquifer systems consisting of unconsolidated to consolidated sediments (Hanson, 1989).

The specific storage value for an aquifer (S_{sk}) is typically indicated by the absence of the prime (') used to signify confining unit parameters. Continuing to neglect the compressibility of water, the water released from storage in an aquifer unit results primarily from recoverable compression and expansion resulting from changing pore water pressure. Very little nonrecoverable compression occurs in aquifer units because appreciable reorientation of the relatively rounded sand grains is not possible. As a result, skeletal specific storage of an aquifer unit may be approximated by the recoverable component alone (Sneed and Galloway, 2000):

$$S_{sk} = S_{ske} = \alpha_{ke}\rho g \tag{2-15}$$

This elastic component of skeletal specific storage (S_{ske}) is typically about an order of magnitude smaller than the elastic skeletal specific storage (S'_{ske}) in a confining unit (Hanson, 1989).

As the above analysis demonstrates, the compaction of an aquifer system is driven primarily by the larger magnitude of compaction of fine-grained confining units and interbeds rather than coarse-grained aquifer units. Aquifer units may contribute small but noticeable amounts of recoverable compaction, but they typically contribute almost no permanent compaction. Not surprisingly, then, the study of aquifer-system compaction and land subsidence is concerned primarily with the strain behavior of confining units and interbeds under changes in stress. This subject is the most important component of the aquitard drainage.

Detailed knowledge of the empirical relation between stress and strain is available from laboratory consolidimeter tests of samples taken from confining units (Helm, 1975). An idealized diagram of the stress-strain behavior of a compacting clay (figure 2-3) provides a graphical representation of this relation, which can also be applied to the behavior of a confining unit or aquitard (Helm, 1975). As the confining unit is subjected to stress (drawdown) greater than any experienced in the past, nonrecoverable strain (compaction) occurs along the solid curve from point A to point B, resulting in a significant and permanent increase in strain represented as the decrease in void space between those points (Helm, 1975). If the water level recovers, resulting in a stress decrease, the strain decreases along the solid curve from point B to C in figure 2-3, resulting in a small increase in void space, or expansion. From the stress level represented by point C, stress less than that experienced previously results in recompaction along the dotted line from C to B. This compaction, however, is reversible within the allowable range of stresses, and the confining unit or aquitard may again expand along the curve from B to C. This hysteresis is often observed both in the laboratory and in the field (Helm, 1975; Riley, 1969). Further nonrecoverable compaction will occur only when the stress level again exceeds that found at point B.

It should be noted that neither S'_{skv} nor S'_{ske} is actually constant. Rather, the actual values of compressibility and S'_{sk} at a particular stress (water) level are given by the slope of the tangent to the compaction curve at that point (Helm, 1975). In practice, the linear approximations represented by the dashed lines from A to B and from C to B in figure 2-3 are used to provide mean compaction and storage values over given ranges of stress because the actual values can not usually be measured in the field. These linear estimates are considered to be reasonably accurate under the ranges of stress and strain usually experienced in compacting aquifer systems (Hanson, 1989).

Delayed Drainage of Confining Units

The pore water stress component of Terzaghi's equation, P, is divided by Helm (1975) into two parts: $P = P_s + U$, where P_s is the equilibrium head in the pore water and U is the transient pore-water stress in excess of the equilibrium. This approach provides a useful understanding of the behavior of the system. As Helm (1975) points out, the excess pore water stress P decays slowly to zero as water flows from an area of higher pressure within a fine-grained unit until the stress equilibrium is approached. The stress equilibrium is reached almost instantaneously in an aquifer, while the process occurs much more slowly in a fine-grained confining unit or aquitard, as explained above. The time τ required for a fine-grained unit to reach equilibrium varies directly with specific storage and with the square of the unit thickness, and it varies inversely with the vertical hydraulic conductivity of the fine-grained unit (Riley, 1969). For a symmetrical, doubly-draining unit, this time constant is defined by Riley (1969):

$$\tau = \frac{S'_{skv}\left(b'_0/2\right)^2}{K'_v} \qquad (2\text{-}16)$$

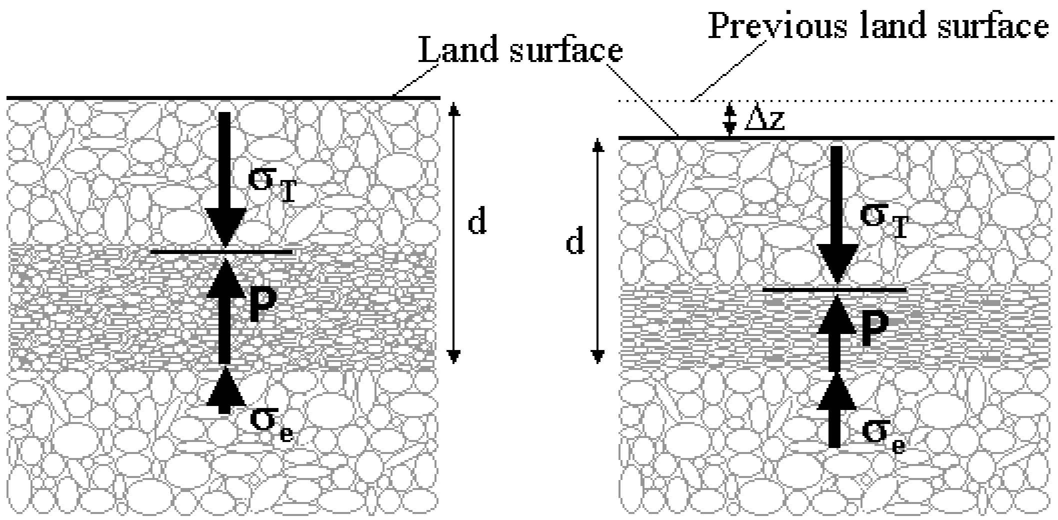
These concepts, which collectively describe both the time-dependent nature of stress (head change) throughout a fine-grained unit and the strain (compaction) response of the unit to the stress, comprise the aquitard drainage model that has been the basis of most land subsidence research.

Discussion of Assumptions

The development of the theory of hydrodynamic compaction and the aquitard drainage model as described here are based on a number of assumptions that are either implied or stated throughout the discussion. The assumption of purely vertical strain merits special consideration here. The definition of the storage coefficient in equation 2-8 (Jacob, 1940) is based on the assumption that the volume reduction resulting from compaction of the aquifer skeleton occurs entirely in the vertical direction. Consequently, this underlying assumption has been the basis of many of the conceptual and numerical models addressing the problem of aquifer-system compaction, but it is known to be invalid (Burbey, 2001). In fact, it has been shown that the loss of void space is the result of strain in three dimensions (Helm, 1974).

Horizontal displacement has been neglected partly because it has been thought to be small, but theories of aquifer mechanics, three-dimensional simulations, and recent field data all indicate that horizontal strain may be significant, especially in conditions of radial flow to a pumping center. While strain in three dimensions is not addressed in depth here, much current research in the field of land subsidence is focused on this subject.

In practice, the equations governing strictly vertical compaction are more manageable, its numerical computation is much simpler, and field measurements of vertical strain are considerably easier. For these reasons, this study and many others have continued to apply the assumption of strictly vertical strain, but the limitations of this approach should be noted.



$$\sigma_e = \sigma_T - P$$

Figure 2-1: The principle of effective stress applied to land subsidence. Modified from Sneed and Galloway (2000).

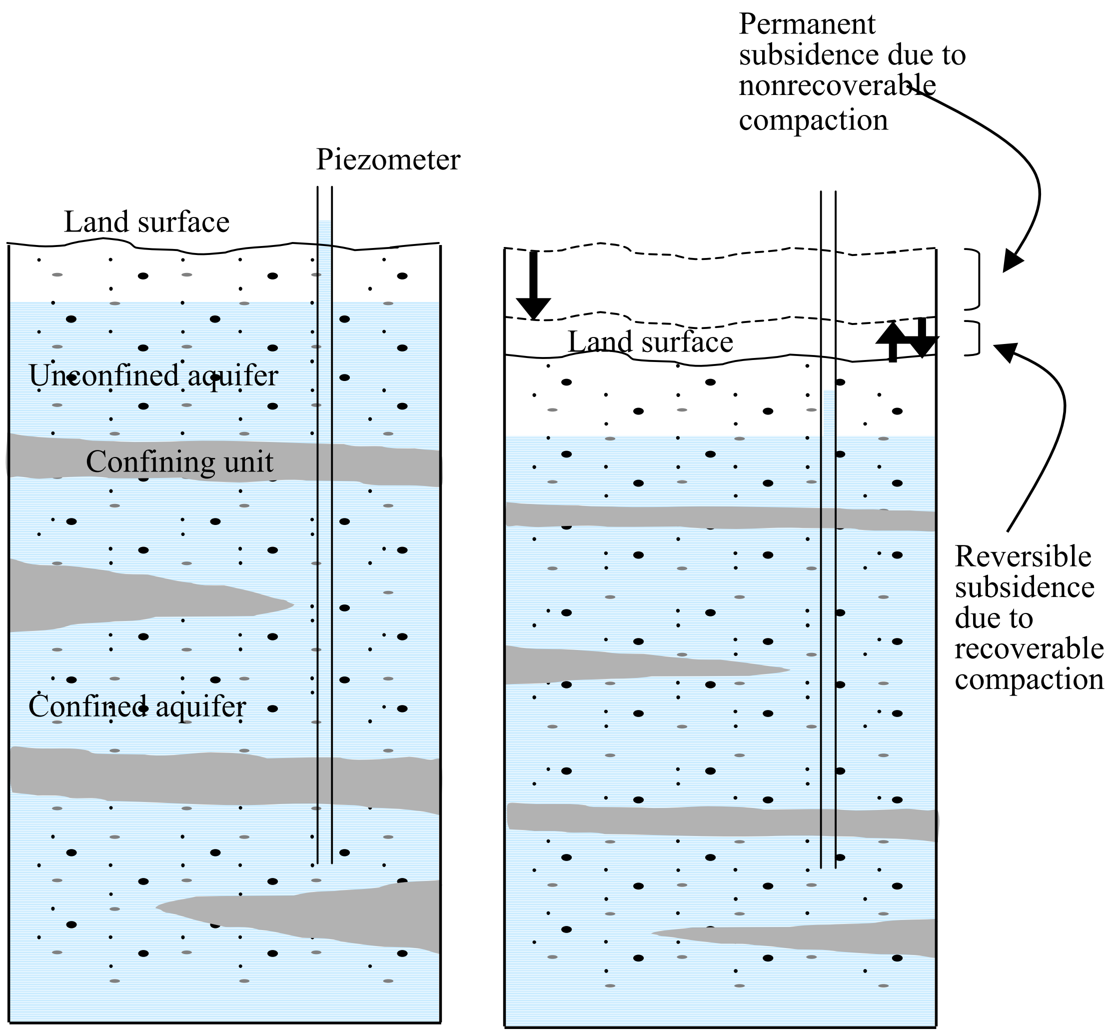


Figure 2-2: Land subsidence due to the change in thickness in a multi-layer aquifer system resulting from a decline in hydraulic head in a confined aquifer.

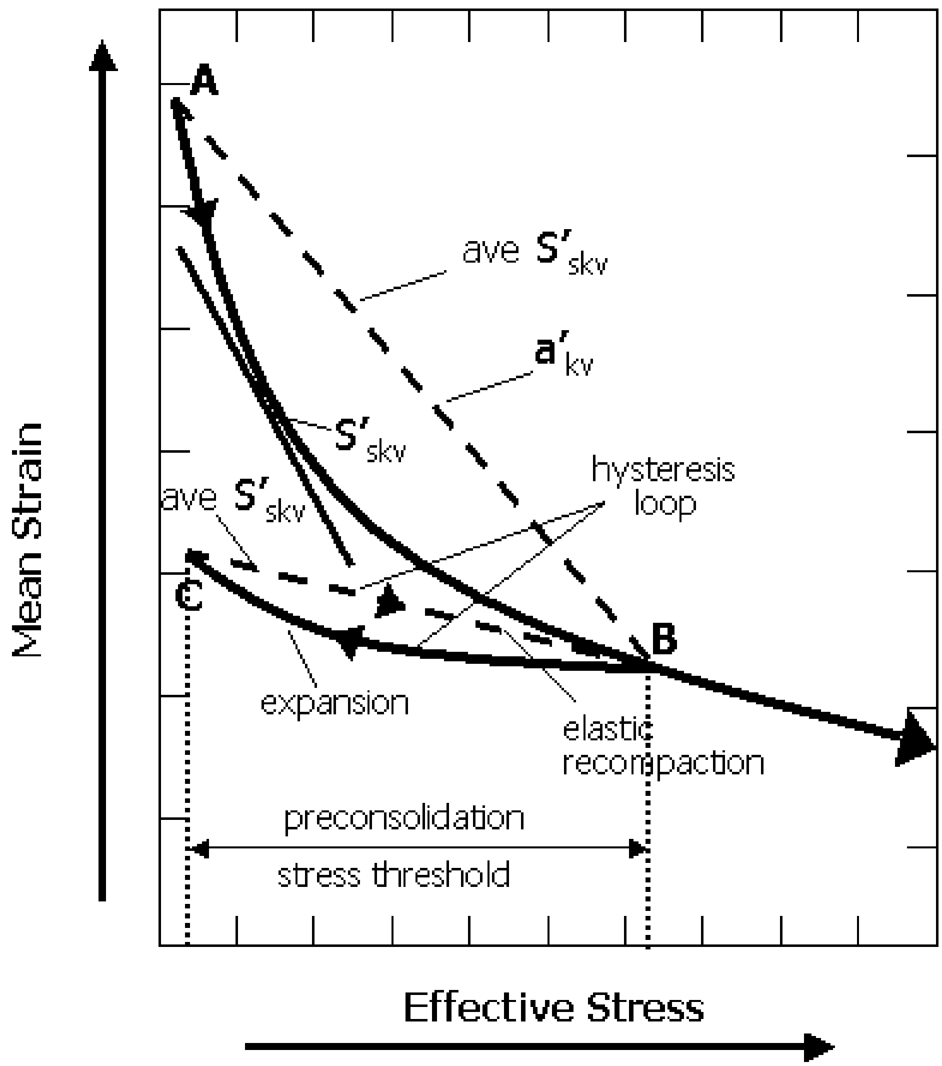


Figure 2-3: Idealized stress-strain relationship in a fine-grained unit. Modified from Helm (1975). See text for full description of diagram.

III: HYDROGEOLOGIC SETTING

Previous Work in the Coastal Plain of Virginia

Clark and Miller (1912) provided the first comprehensive report on the geology of the Virginia Coastal Plain, while Sanford (1913) delivered an early analysis unifying the region's geology and groundwater resources. Cederstrom (1945, 1957) carried out a large amount of work in the Virginia Coastal Plain and provided thorough descriptions of the hydrogeology of southeastern Virginia (1945) and the York-James Peninsula (1957). In addition to describing and correlating hydrogeologic units, Cederstrom's works also supplied a large amount of useful information on groundwater quality, usage patterns, and the history of groundwater development in the Virginia Coastal Plain. A number of subsequent reports have described various aspects of geology and hydrogeology in the Coastal Plain of Virginia, including stratigraphy, geochemistry, and aquifer properties. Reports by Brown and Cosner (1974), Cosner (1975) and Hopkins and others (1981), describing groundwater conditions in the Coastal Plain were of particular relevance to the current work, as were reports on eastern Virginia water resources provided by Richardson and others (1988), Kull and Laczniak (1987), and Laczniak and Meng (1988).

Meng and Harsh (1988) integrated the available data on geology and hydrogeology to create the first complete hydrogeologic framework of the Coastal Plain in Virginia. Recently available geophysical, palynological, and deep core data allowed Meng and Harsh (1988) to refine and expand upon the conclusions of earlier authors. Their study, completed as part of the Regional Aquifer System Analysis (RASA) program of the USGS, remains the definitive work on Coastal Plain hydrogeology in Virginia and was immensely helpful in the completion of the current research. Harsh and Laczniak (1990) expanded on the work of Meng and Harsh with a detailed analysis of the Coastal Plain groundwater flow system in Virginia. Their work included the development of a computer flow model for the Virginia Coastal Plain aquifer system using the framework of Meng and Harsh (1988), and it significantly enhanced the current understanding of the behavior of the system. A related modeling study by Hamilton and Larson (1988) focused on the hydrogeology of the Virginia Coastal Plain south of the James River, an area known as southeastern Virginia. Aside from the detailed hydrogeologic data and analysis it provided for that area, the Hamilton and Larson (1988) report was important because it included a slight revision to the Coastal Plain framework developed by Meng and Harsh, adding one additional aquifer and confining unit to the aquifer system. The Virginia studies were ultimately incorporated into a comprehensive groundwater modeling study of the entire Northern Atlantic Coastal Plain, which was completed as part of the RASA program (Leahy and Martin, 1993).

The recent discovery of a bolide impact crater beneath the Chesapeake Bay area (Poag and others, 1994; Poag, 1997) has led to further reconsideration and revision of the geologic and hydrogeologic framework of the Virginia Coastal Plain (Powars and Bruce, 1999). The 56-mile-wide crater (figure 3-1) was created approximately 35 million years ago by the impact of a large comet or meteorite, which disrupted the pre-impact sediments and rocks in the region (Powars, 2000). The presence of the crater has also apparently affected the geologic structure of the region and influenced the subsequent deposition and erosion of sediments in the surrounding area (Powars, 2000). Thus, while the most significant effects of the impact can be found within the boundary of the crater, the impact and the crater it created appear to have influenced the geologic history of most of the Virginia Coastal Plain (Powars and Bruce, 1999). As a result, Powars and Bruce (1999) and Powars (2000) have proposed revisions to the geology of the Virginia Coastal Plain reflecting these influences.

The framework outlined by Meng and Harsh (1988), with the addition of refinements reported by Hamilton and Larson (1988), has for years been the accepted interpretation of the Coastal Plain aquifer system in Virginia, particularly outside the boundary of the impact crater. While Powars and Bruce (1999) correctly point out that the crater discovery has "revealed the inadequacy of the layer-cake, multi-aquifer model" of the Virginia Coastal Plain, their revisions to the framework focus on the geology and hydrogeology within the crater boundary, which is beyond the scope of this work. Outside the crater boundary, the 'original' framework is somewhat intact, though recent publications have revealed a number of problems with long-accepted interpretations. Powars and Bruce (1999) address revisions to the geologic framework on the York-James peninsula of Virginia, and Powers (2000) addresses related revisions to the geologic framework south of the James. Not surprisingly, these revisions suggest some associated changes in the hydrogeologic framework, which currently does not reflect the known effects of the impact crater. Many of these changes are now being included in a reassessment of the hydrogeologic framework by the Virginia District of the USGS Water Resources Division (McFarland, oral communication, 2001). The comprehensive numerical model originally developed by Harsh and Laczniak (1990) has been repeatedly updated and refined by the USGS in order to incorporate new hydrogeologic data, reflect improvements in model design, and support the more sophisticated application of the model as a water resource management tool (McFarland, 1998). Another revision to the model is currently being undertaken by the USGS and the Virginia Department of Environmental Quality in order to address these issues. Among the many changes, the current model revision will reflect the numerous influences of the impact crater on the hydrogeologic framework and groundwater conditions of the Virginia Coastal Plain (Erwin and others, 1999; McFarland, oral communication, 2001).

The geologic units and related hydrogeologic framework described by the above authors has been adopted as the basis for this work. The geologic history and descriptions of geologic and hydrogeologic units provided here are simply summaries of the information provided in those reports; while some synthesis is attempted, no further interpretations are made here. Instead, this discussion is intended as a concise overview of the hydrogeology of the Virginia Coastal Plain as it is currently understood. Importantly, no published reports yet consider the effects of the impact crater or other revisions to the hydrogeologic framework in Virginia, and such a task is well beyond the scope of this work, though not insignificant to it. As a result, some aspects of the revision currently underway are discussed in this report, based on numerous and lengthy oral communications with USGS hydrologist Randy McFarland (2001).

One further clarification on this subject is necessary. Much of the recent work, published and unpublished, on the geology and hydrogeology of the Virginia Coastal Plain concerns the geologic influence of a structural zone associated with (and approximately parallel to) the James River. The apparent influence of this zone on Virginia Coastal Plain geology is such that interpretations of geologic and hydrogeologic units are significantly different on either side of this boundary (Powars and Bruce, 1999). A detailed analysis of these differences is underway and, again, beyond the scope of this work. Nevertheless, a very brief discussion of the James River structural zone is undertaken here, and some mention is made of the effects of that zone on hydrogeology north of the James River. For the most part, however, the descriptions of the geologic and hydrogeologic units undertaken here concern the units as identified south of the James River, within the explicit study area of this work. These units are outlined in the chart in figure 3-2.

Topography, Geology, and Depositional History

The Virginia Coastal Plain is part of the Atlantic Coastal Plain province, which extends from Cape Cod, Massachusetts to the Gulf of Mexico, as described generally by Heath (1983) and others. While the various parts of the province share similar characteristics, only the Virginia Coastal Plain is described here. The topography of the Virginia Coastal Plain was described succinctly by Meng and Harsh (1988) as, "a series of broad, gently sloping, highly dissected terraces bounded by seaward-facing, ocean-cut escarpments extending generally north-south across the province." Most of the Coastal Plain in Virginia is less than 100 ft in elevation, with the highest elevations found along the Fall Line (Meng and Harsh, 1988).

Geologically, the Coastal Plain in Virginia can be most simply described as a sedimentary wedge of unconsolidated gravels, sands, silts, and clays with variable amounts of shells (Meng and Harsh, 1988). This sedimentary wedge (figure 3-3) thickens to the east, ranging in thickness from almost zero at the Fall Line to over 6,000 ft beneath the Eastern Shore Peninsula (Meng and Harsh, 1988). All units dip approximately to the southeast and strike approximately parallel to the Fall Line, though, as Meng and Harsh (1988) point out, the dip of each successive depositional unit decreases upward. As a result, the oldest and lowermost deposits dip at 40 ft/mi along the basement surface, while the youngest surficial deposits dip less than 3 ft/mi.

The unconsolidated sediments range in age from Early Cretaceous to Holocene (figure 3-2) and indicate a complex history of deposition and erosion outlined by Meng and Harsh (1988) as "a thick sequence of nonmarine deposits overlain by a much thinner sequence of marine deposits." The sedimentary deposits rest on a crystalline rock basement that is a combination of massive igneous and highly deformed metamorphic rocks ranging in age from Precambrian to Lower Paleozoic, along with some unmetamorphosed consolidated sediments and igneous intrusives of Triassic age (Meng and Harsh, 1988).

Meng and Harsh (1988) divide the Coastal Plain sediments of the Virginia Coastal Plain into five principal lithostratigraphic groups based on the mode of deposition. From oldest to youngest, these groups are the Lower Cretaceous and lowermost part of the Upper Cretaceous Potomac Formation; the uppermost Cretaceous deposits; the lower Tertiary Pamunkey Group; the upper Tertiary Chesapeake Group; and undifferentiated Quaternary sediments (Meng and Harsh, 1988) (figure 3-2).

The thick sedimentary sequences of the Early Cretaceous group resulted from the deposition of continental and marginal marine sediments by rivers, which formed large subaerial deltas on the broad underlying rock surface. The fluvial-deltaic nature of this deposition produced interfingering sandy channel deposits and clayey inter-channel deposits. As a result, these early Cretaceous sediments vary laterally and are known to thicken, thin, or pinch out over relatively short distances (Hamilton and Larson, 1988). Even so, they make up approximately 70 percent of the thickness of the Coastal Plain sequence (Hamilton and Larson, 1988). The first marine transgression early in the Late Cretaceous inundated the area with shallow seas and extensive estuaries. After a long period of non-deposition as a result of a marine regression early in the Late Cretaceous, a marine transgression deposited the clays, sandy clays and marls of the uppermost Cretaceous group. The relatively thin but areally extensive sediments that form the two Tertiary groups were deposited by what Meng and Harsh describe as interbasinal marine seas. The Pamunkey Group of the lower Tertiary comprises glauconitic sands, silts and clays, while the Chesapeake Group of the upper Tertiary is composed of shelly clays, silts, and sandy clays. Because of the widespread and uniform nature of the transgressing seas, the Tertiary deposits are relatively homogeneous and uniform throughout the Coastal Plain (Hamilton and Larson, 1988), except where they have been disrupted by the Chesapeake Bay impact crater (Powars and Bruce, 1999).

The impact crater itself was created in the late Eocene, when the coastline was west of the present-day Fall Line (Poag and others, 1994). The comet or meteorite that struck the inner continental shelf disrupted the entire Eocene to Cretaceous sedimentary sequence and penetrated into the underlying crystalline basement rocks of Proterozoic to Paleozoic age. The impact caused the ejection of a large amount of debris and generated a series of large tsunamis that spread the debris over most of what is now the U.S. Atlantic continental shelf (Poag and others, 1994). The result of the impact was a complex crater 56 miles wide and about 1.3 miles deep, underlain and surrounded by a zone of shocked, fractured and faulted crystalline basement rock that extends to a depth of over six miles (Powars and Bruce, 1999). The crater was partially filled almost immediately with collapsed and slumped material from the crater rim, as well as a chaotic mix of sediments from the pre-impact sequence (Powars and Bruce, 1999). The sediments deposited at the time of the impact are divided into two principal depositional units by Poag (1997): the Chesapeake Bay impact crater megablock beds and the overlying Exmore tsunami-breccia deposits. The megablock beds are found directly above the crystalline basement inside the crater rim and are interpreted as consisting primarily of "Lower Cretaceous fluvial-deltaic deposits that slumped into the crater during an early stage of crater filling and covered the floor of the annular trough" (Powars and Bruce, 1999). These megablock beds range from 700 to 2,500 feet in thickness (Powars and Bruce, 1999). The tsunami-breccia deposits overlie either the previously described megablock beds or the crystalline basement within the crater or the pre-impact lower Cretaceous deposits outside the crater rim. The tsunami-breccia unit is described by Powars and Bruce as having "a highly variable lithology," consisting of a fining-upward sequence of sand containing abundant material from all pre-impact sedimentary units, as well as "trace amounts of shocked quartz", "fragments of melt rock," and scattered clasts of crystalline basement," containing deformation features (Poag and others, 1994; Powars and Bruce, 1999). The thickness of the tsunami-breccia is highly variable, but estimated to reach a maximum of over 2,000 feet inside the inner basin of the crater (Powars and Bruce, 1999).

Within the crater, post-impact upper Eocene sediments cap the Exmore tsunami-breccia, but upper Oligocene sedimentary units are the earliest to be preserved without disruption across the crater boundary (Powars and Bruce, 1999). Beginning with this upper Oligocene unit, the crater is buried under hundreds of feet of post-impact sediments resulting from a series of marine transgressions across the mid-Atlantic Coastal Plain in the late Tertiary (Powars and Bruce, 1999). While they are not disrupted by the crater boundary, sedimentary units deposited since the impact tend to thicken into the crater, demonstrating the control on the depositional environment by the structural depression associated with the crater (Powars and Bruce, 1999). However, this structural influence becomes less apparent in later sediments, and is almost absent outside the disruption boundary.

Across the Virginia Coastal Plain outside the crater, the Tertiary deposits are overlain by Quaternary fluvial and marine deposits that are interpreted as reflecting Pleistocene sea-level fluctuations. Cycles of erosion due to sea-level drops and deposition due to sea-level rises produced drowned and infilled river valleys and extensive terraces (Hamilton and Larson, 1988). Overlying the Pleistocene sediments, but included in the same group by Meng and Harsh, is a thin layer of Holocene sediments deposited in lagoons, beaches and tidal flats during rising sea levels since the Pleistocene (Hamilton and Larson, 1988). Within the crater boundary, the Pliocene to Quaternary deposits are thought to be influenced by continued structural deformation associated with the crater (Powars, 2000), but the influence of the crater on Quaternary sediments is otherwise minimal.

A discussion of Coastal Plain geology and hydrogeology would be incomplete without some consideration of the structural setting, which has played a significant role in controlling deposition. The Coastal Plain as a whole is characterized as a regionally downwarped feature controlled by crustal deformation, but local variations in this regional structure are significant. The structural highs and lows associated with the Coastal Plain in Virginia are described by Meng and Harsh (1988), but were previously recognized by a number of authors.

The feature most important to this study is the structural high known as the Norfolk Arch, located near and approximately parallel to the present-day James River (figure 3-2). Its axis, which dips to the southeast, divides the Salisbury embayment to the north and the Albemarle embayment to the south (Meng and Harsh, 1988). The influence of the arch and these two basins on the depositional history of the Coastal Plain is described in detail by Meng and Harsh (1988), and later by Powars and Bruce (1999) and Powars (2000). The later authors provide more detail on the structural complexity of this feature, describing a "James River structural zone" coincident with the north flank of the Norfolk arch (Powars, 2000). Together with this structural zone, the arch forms the northern boundary of the study area for the current work. In short, these structural features are thought to be the controlling factors in the formation of two distinct basins. The influence of this feature on the depositional environment appears to have become less significant with time, as late Tertiary and subsequent deposits do not seem to be affected by the arch and embayment structure (Meng and Harsh, 1988).

Hydrogeologic Framework

Descriptions and hydrogeologic data are given here for nine major confined aquifers, nine associated confining units, and the uppermost water-table aquifer, following the framework as developed by Meng and Harsh (1988), modified by Hamilton and Larson (1988), and outlined by McFarland (1998). Hydrogeologic units are described in order from oldest to youngest, and units characterized include those identified in the most recent modeling efforts (McFarland, 1998). While all units present in the Virginia Coastal Plain are described here for completeness, it should be noted that not all of these units are found in southeastern Virginia, which is the area most important to the current work.

The current framework (figure 3-4) does not yet include the modifications proposed by Powars and Bruce (1999) and Powars (2000), reflecting the hydrogeologic effects of the recently-discovered Chesapeake Bay impact crater. While these recent reports on the effects of the crater have begun to answer some important and formerly unsolved questions about the Coastal Plain aquifer system in Virginia, the analysis of the crater's effects is not yet complete. Powars (2000) argues convincingly for a revision of the current framework, pointing out the problems with attempting to correlate the units of the current hydrogeologic framework with the geologic units inside the crater's disruption boundary. Nonetheless, such a revision has not yet been completed. Furthermore, most of the Coastal Plain aquifer system outside the area of the crater's direct influence can still be described adequately, if not completely, by the framework as it currently exists. Nonetheless, data from the recent studies have been incorporated into this work, and comments concerning the influence of the crater on regional hydrogeology have also been included where appropriate.

As is usually the case, the hydrogeologic units described by the above authors are defined by lithologic and hydrologic properties rather than by their stratigraphic boundaries. Thus, a hydrogeologic unit might be composed of several similar geologic formations, or a single geologic formation with lithologic variations might be divided into an aquifer and a confining unit. Analyses of water-level data were used to confirm and support these decisions (Meng and Harsh, 1988). As Meng and Harsh (1988) point out, it is not uncommon for a hydrogeologic unit to consist of combinations or divisions of geologic formations (1988). Nonetheless, the name of a particular hydrogeologic unit, whether aquifer or confining unit, is based on the name of the geologic formation or formations most closely associated with that unit. For example, the Nanjemoy-Marlboro clay confining unit described below is made up of both the upper Paleocene Marlboro clay and the lower Eocene Nanjemoy formation (Meng and Harsh, 1988). Further data and justification for the differentiation of these hydrogeologic units can be found in the reports that collectively define the Virginia Coastal Plain hydrogeologic framework (Meng and Harsh, 1988; Hamilton and Larson, 1998; Harsh and Laczniak, 1990). For the purposes of this report, figure 3-2 provides an overview of the general correlations of the Virginia Coastal Plain aquifer system units with the geologic units of the same area. The upper Eocene geologic units associated with the Chesapeake Bay impact crater are not depicted in this figure.

The summary of hydrogeologic units given below includes a general overview of each unit, including lithologic and stratigraphic information, inferred depositional environments, inventories of known hydrogeologic properties, and information on water supply issues associated with each of the units. In addition to the referenced publications, extensive oral communication and written correspondence with the staff of the Virginia Division of the USGS Water Resources Division were invaluable in the compilation of this information.

Lower Potomac Aquifer

The Lower Potomac aquifer is the lowermost confined aquifer in the hydrogeologic framework defined by Meng and Harsh (1988). It consists of "thick, interbedded sequences of angular to subangular coarse sands, clayey sands and clays", which range in age from early to middle Early Cretaceous (Meng and Harsh, 1988). This wedge-shaped unit strikes approximately north-south and dips eastward at approximately 40 ft/mi (5.7 m/km), extending across the entire Coastal Plain of Virginia (figure 3-5). It thickens eastward from a thickness of almost zero at the Fall Line to a maximum of over 3,000 ft near the Atlantic coast. Throughout its extent, the Lower Potomac aquifer rests entirely and unconformably on the basement surface and is overlain by the Lower Potomac confining unit.

The sands of the Lower Potomac aquifer are crossbedded and contain a high amount of interstitial clays, as well as many clay lenses and beds. The sands themselves have commonly been described as coarse and gray. More specifically, they are composed of medium to very coarse subangular quartz, with abundant weathered potassium feldspar and some plagioclase (Meng and Harsh, 1988). The "light to drab-colored" well-bedded clays are typically composed of "mixed-layer illite/smectite," according to lithostratigraphic descriptions, while the "interstitial and laminated clays" are composed primarily of kaolinite (Reinhardt and others, 1980). Electric resistivity logs show patterns that are "blocky in profile, indicating massively bedded sequences with relatively sharp lithologic contacts among sands, clayey sands and clays" (Meng and Harsh, 1988).

An understanding of the depositional environment which formed this unit is helpful in explaining what Meng and Harsh (1988) call "the lithologic heterogeneity and discontinuous nature" of the Lower Potomac aquifer sediments. The sediments here are known to be poorly sorted and contain a large amount of coarse material. This evidence, along with bedding characteristics, has led several studies to suggest that these sediments represent the development of a continental delta in a broad alluvial plain dominated by braided streams and occasionally inundated by marine seas (Meng and Harsh, 1988). This mode of deposition explains why the clay beds found within the unit and in the overlying Lower Potomac confining unit are “not continuous or areally extensive,” but consist instead of a series of “interlensing clayey deposits” (Hamilton and Larson, 1988).

The extremely variable and heterogeneous nature of the Cretaceous clays has become increasingly apparent with recent work and has caused some geologists and hydrogeologists to question the approach of defining distinct confining units in the Cretaceous sediments (McFarland, oral communication, 2001). Instead, some have suggested that the Early-to-Middle Cretaceous sediments of the Potomac Formation might better be represented as one large aquifer with a number of discontinuous clay interbeds. While revisions to the hydrogeologic framework currently in progress will likely take such an approach, all recent publications define several distinct Cretaceous aquifers, and water-level and withdrawal data have been developed based on that level of discretization. As a result, this report follows the approach taken in the recent literature, but an attempt has been made to note the areas where the published work may not be consistent with current understanding. While a detailed treatment of the potential revisions and their development is beyond the scope of this work, an outline of the concepts behind them is critical to a complete understanding of Coastal Plain hydrogeology and may be important in understanding and analyzing land subsidence in this region.

The thickness of the Lower Potomac aquifer sediments, along with relatively high hydraulic conductivity throughout, gives the aquifer large transmissivity values, generally increasing in the down-dip direction. Aquifer tests at a variety of locations in the Coastal Plain yielded mean transmissivity values of 12,000 to 19,000 ft^2/d, or 1,115 to 1,765 m^2/d, while calibration of a groundwater flow model produced maximum transmissivity values of just over 12,000 ft^2/d, or 1,115 m^2/d (Harsh and Laczniak, 1990). Estimates and model calibrations yielded average lateral hydraulic conductivity values from 25 to 42 ft/d, or 7.6 to 12.8 m/d (Harsh and Laczniak, 1990). Aquifer storage coefficients, derived from the same aquifer tests, range from 6.0×10^{-4} to 1.1×10^{-3}. Storage coefficient values from flow models ranged from 8.0×10^{-6} to 2.5×10^{-4} (Hamilton and Larson, 1988).

As these high numbers indicate, the Lower Potomac aquifer is known to be capable of producing large quantities of water. Yields from individual wells may exceed 1,500 gallons/min, though the common range of yields is 100 to 800 gallons/min (Hamilton and Larson, 1990). However, the depth of the aquifer below land surface, and the associated drilling and pumping expenses, preclude use by all but the largest industrial applications, and only a handful of wells penetrate this aquifer. Furthermore, increasingly high chloride concentrations in the down-dip direction also restrict the use of groundwater from this aquifer (Meng and Harsh, 1988). Even so, large industrial users make the Lower Potomac aquifer the third largest supplier of water in the Coastal Plain, and withdrawals at several major pumping centers have caused tremendous drawdowns in groundwater levels near those pumping centers and large associate cones of depression. For example, water levels at the pumping center at Franklin, Virginia once rose over 20 ft above land surface (Cederstrom, 1945) but have fallen to 150 ft below sea level in recent years (White and Powell, 1996).

Lower Potomac Confining Unit

The Lower Potomac confining unit, which overlies the Lower Potomac aquifer, consists of sequences of brown, gray, or dark-green carbonaceous clay, interbedded with thin, sandy clay (Hamilton and Larson, 1988). While it is described as a confining unit, recent studies emphasize the highly discontinuous and heterogeneous nature of the clays, pointing out that this supposed unit demonstrates little lateral continuity. Revisions currently in progress will likely describe the clays of this unit as another interbed in one large Potomac aquifer, as noted previously (McFarland, oral communication, 2001).

The sediments of this unit are generally middle Early Cretaceous in age (Meng and Harsh, 1988). Like the underlying Lower Potomac aquifer, this confining unit is thinnest along the Fall Line and thickest toward the coast. It attains a maximum known thickness of 173 ft (52.7 m) in the northeastern part of the Coastal Plain (Meng and Harsh, 1988) and a maximum thickness in the southeastern Virginia study area of approximately 80 ft (24.4 m) near the city of Virginia Beach (Hamilton and Larson, 1988). The Lower Potomac confining unit rests atop the Lower Potomac aquifer throughout the study area, except where the later crops out in stream valleys near the Fall Line (Meng and Harsh, 1988). This confining unit is overlain by the Middle Potomac aquifer.

The sediments of the Lower Potomac confining unit are referred to colloquially as "hard and tough," while lithologic studies describe the sequence as "a massive, clay-dominated interval composed of thick sequences of finely laminated, carbonaceous clays interbedded with thin sandy clay beds" (Meng and Harsh, 1988). More specifically, the thickly bedded clays of this interval are identified as "mixed-layer illite/smectite" with a high percentage of expandable clays, and the laminated carbonaceous clays found here are identified as "predominately kaolinitic" (Reinhardt and others, 1980). Well logs indicate both "the sharp lithologic contacts between the thickly-bedded confining clays and the overlying and underlying aquifer sands," and the "massively bedded nature" of these clays (Meng and Harsh, 1988). The lithology of this unit distinguishes it from the underlying interbedded clays of the Lower Potomac aquifer, as the aquifer clay units contain significantly more sands (Meng and Harsh, 1988).

Due to the depositional environment, the clay beds of this confining unit consist of a series of "interlensing clayey deposits," rather than a "continuous, areally extensive layer" (Meng and Harsh, 1988). In this way, the unit is much like the interbedded clay units of the underlying aquifer. Nonetheless, the "predominance of finer-grained clayey materials and their associated bedding characteristics" in the Lower Potomac confining unit has caused several studies to hypothesize a change in depositional environment from that of the Lower Potomac aquifer. In general, the change is described as a progression from a "high-gradient to a lower-gradient fluvial flood plain," and the paleoenvironment is described as a "shallow, discontinuous back-swamp basin" (Meng and Harsh, 1988).

While the clay deposits are not continuous, water-level observations indicate that the interlensing causes the deposits to behave regionally as a single confining unit, and they are represented as such in groundwater flow models (Harsh and Laczniak, 1990). Laboratory analysis of the clays of this confining unit yielded a vertical hydraulic conductivity of 1.9×10^{-6} ft/d, or 5.8×10^{-7} m/d (Harsh and Laczniak, 1990), though it should be noted that measurement artifacts and field-scale heterogeneities generally cause laboratory measurements to underestimate the actual values. One initial estimate gave a vertical hydraulic conductivity value of 8.5×10^{-4} ft/d, or 2.6×10^{-4} m/d. (Harsh and Laczniak, 1990). Calibrations of groundwater flow models yielded a range of 3.28×10^{-5} ft/d, or 1.00×10^{-5} m/d (Harsh and Laczniak, 1990) to 4.32×10^{-5} ft/d, or 1.32×10^{-5} m/d (Hamilton and Larson, 1988). Very little data are available for calculating accurate specific storage values for this confining unit or others, and current models do not consider the effects of water released from confining-unit storage (Harsh and Laczniak, 1990).

Middle Potomac Aquifer

The Middle Potomac aquifer is the second lowest confined aquifer in the hydrogeologic framework. Present throughout the Virginia Coastal Plain, it overlies the Lower Potomac confining unit and underlies the Middle Potomac confining unit. It is confined throughout, except for a narrow outcrop along the Fall Line (Harsh and Laczniak, 1990). It consists of "interlensing medium sands, silts, and clays of differing thickness," which are late Early Cretaceous in age (Meng and Harsh, 1988). Like the Lower Potomac aquifer, this aquifer is wedge-shaped and extends across almost the entire Coastal Plain, striking north-south and dipping to the east at 15 to 25 ft/mi, or 2.8 to 4.7 m/km (Meng and Harsh, 1988) (figure 3-6). From its western limit near the Fall Line, the Middle Potomac aquifer thickens to a maximum of almost 1000 feet in the eastern part of the Coastal Plain (Meng and Harsh, 1988). Its similarity with the Lower Potomac aquifer has caused some to suggest that it may be best defined as part of that unit, though water levels indicate definite vertical gradients across the two units (McFarland, oral communication, 2001).

The sediments of the Middle Potomac aquifer are described by early studies as a thick sequence of interbedded silty clays and fine to medium sands (Meng and Harsh, 1988). A later lithostratigraphic study more specifically describes this unit as a "sand-dominated interval characterized by distinct fining-upwards sequences interbedded with laminated or massive clays" (Reinhardt et al., 1980). The sands are described as being composed of "coarse to fine, angular to subangular quartz, and some plagioclase," and are also "commonly micaceous and contain abundant heavy minerals" (Reinhardt et al., 1980). The "laminated and massive clays" interbedded within this sequence are composed of "mixed kaolinite and highly expandable illite/smectite" (Reinhardt et al., 1980). Well logs for this aquifer provide further information that distinguishes this unit from the otherwise similar Lower Potomac aquifer. In particular, resistivity logs show high-resistance values for the sands of the Middle Potomac aquifer, reflecting both the clean sands of this unit and the relatively low concentrations of dissolved solids in the water from this unit (Meng and Harsh, 1988). Furthermore, the resistivity log patterns here are somewhat different from those of the Lower Potomac, showing "triangular and saw-toothed" profiles rather than the blocky profiles of the Lower Potomac (Meng and Harsh, 1988). According to Meng and Harsh, the triangular profiles indicate "the fining-upwards sequences characteristically associated with the aquifer sands," while the saw-toothed profiles indicate "the extensively interbedded sequences of sands, silts, and clays also characteristic of these sediments."

Of course, the lithologic character of the Middle Potomac aquifer sediments is closely related to the environment of deposition, which consisted of "a low-gradient, subaerial, fluvial flood plain dominated by meandering streams," according to interpretations by Meng and Harsh (1988) of earlier works by others. More specifically, these sedimentary deposits "are dominated by channel sands, point bars, levees, flood plains, and backswamps, according to Meng and Harsh (1988). A lithostratigraphic study (Reinhardt, 1980) notes the absence of glauconite in these sediments and suggests that they are distinctly continental and represent a "more landward assemblage," than the sediments of the Lower Potomac aquifer (Meng and Harsh, 1988). The sediments of the Middle Potomac aquifer, together with the underlying lower Potomac aquifer sediments have been generally interpreted as representing the development of a continental delta (Reinhardt and others, 1980).

As with the Lower Potomac aquifer, the large thickness and high hydraulic conductivity throughout the Middle Potomac gives the aquifer relatively high transmissivity values. While this aquifer unit thickens to the east, however, the transmissivity values generally decrease seaward due to increasing amounts of silt and clay (Harsh and Laczniak, 1980). Aquifer tests across the Virginia Coastal Plain produced average transmissivity values ranging from 2,000 ft^2/d in the north-central Virginia Coastal Plain to a maximum of about 19,000 ft^2/d or 1,765 m^2/d in southeastern Virginia (Harsh and Laczniak, 1988). A measured transmissivity value of 56,800 ft^2/d has been reported (Hamilton and Larson, 1988), but model calibrations support maximum values of less than 19,000 ft^2/d (Harsh and Laczniak, 1988; McFarland, 1998). Estimates of average lateral hydraulic conductivity from model simulations ranged from approximately 25 to 52 ft/d, or 7.6 to 15.8 m/d (Harsh and Laczniak, 1988). Storage coefficients reported by Harsh and Laczniak from a number of sources range from 2.0×10^{-4} to 1.5×10^{-3}, though Hamilton and Larson (1988) reported storage coefficients ranging from 1.6×10^{-6} to 9.3×10^{-3}, demonstrating a wide variation in the measurements.

The Middle Potomac aquifer provides the largest source of groundwater in the Virginia Coastal Plain, and it is the primary source for large water users (Harsh and Laczniak, 1988). Yields from wells penetrating this aquifer in southeastern Virginia have been reported to exceed 750 gal/min (Brown and Cosner, 1974), though common yields are reported to range from 20 to 160 gal/min (Hamilton and Larson, 1988). The aquifer is used by most municipal and industrial water users in the western half of the Virginia Coastal Plain, but the depth of the aquifer makes use prohibitively expensive for all but large industrial water users in the eastern half (Meng and Harsh, 1988). As a result, the number of wells penetrating the aquifer tends to decrease toward the coast. Withdrawals from the Middle Potomac aquifer began early in the 20th century and now exceed 50 million gallons per day (Harsh and Laczniak, 1988). They have continued to increase at a moderate rate (Harsh and Laczniak, 1988). These withdrawals have resulted in tremendous declines in measured water levels, particularly over the decades following World War II. This aquifer once produced flowing wells (Cederstrom, 1945), but groundwater levels have declined to over 150 feet below sea level in recent years (White and Powell, 1996). As with the Lower Potomac aquifer, high chloride concentrations restrict the use of water from the eastern part of the Middle Potomac aquifer. Otherwise, the water quality is generally good, with relatively low concentrations of dissolved solids (Meng and Harsh, 1988).

Middle Potomac Confining Unit

The middle Potomac confining unit is defined by Meng and Harsh (1988) as “the major clayey strata directly above the middle Potomac aquifer.” The confining unit overlies the middle Potomac aquifer throughout its extent, except for a narrow outcrop of the aquifer along the Potomac River in the northwestern part of the Virginia Coastal Plain (Harsh and Laczniak, 1990). It is overlain by the upper Potomac aquifer over most of its extent, except where it is overlain by the Aquia in the western part of the Coastal Plain (Meng and Harsh, 1988). The sediments of the confining unit are of late Early Cretaceous in age and comprise the middle part of the Potomac formation (Harsh and Laczniak, 1990). Like the lower Potomac confining unit, this unit is thinnest near the Fall Line and thickens toward the coast, obtaining a maximum known thickness of just over 200 ft at a well on the Eastern Shore Peninsula (Meng and Harsh, 1988). The maximum known thickness in southeastern Virginia is 132 ft, measured in the city of Chesapeake (Hamilton and Larson, 1988). Several sources describe the thickness of the middle

Potomac confining unit as highly variable, but it is commonly the thickest confining unit or interbed found in the Potomac formation (Meng and Harsh, 1988). The average thickness used in flow modeling is 40 ft (Harsh and Laczniak, 1990).

The sediments of the middle Potomac confining unit are characterized by "a thick sequence of brightly-colored, highly variegated, plastic clays," commonly used by drillers to identify the unit (Meng and Harsh, 1988). These multicolored clays are also known as "mottled clays," and include red, purple, gray, brown, olive, and yellow units (Meng and Harsh, 1988). The clays of the middle Potomac confining unit are similar in composition to those of the lower Potomac confining unit described earlier; they contain primarily a mixture of kaolinite and illite/smectite known to be highly expandable (Reinhardt and others, 1980). Another description of this unit refers to the composition similarly as predominantly montmorillonitic red clays (Harsh and Laczniak, 1988). Sources agree that the clays of this unit are usually massive and thickly bedded, mottled, highly oxidized, and highly fractured, but they can also be finely laminated in some places, in which case they are usually silty, sandy, micaceous, and carbonaceous (Meng and Harsh, 1988; Harsh and Laczniak, 1990).

Like the lower Potomac confining unit, the middle Potomac confining unit is not composed of a single, continuous layer, but can be characterized as "a series of interfingering deposits" that together act as a single, regional confining unit (Meng and Harsh, 1988), albeit a rather leaky one in places. These interfingering sedimentary units are the result of the environment of deposition in which they formed. This environment is described concisely as "broad, low-gradient fluvial-deltaic plains containing extensive flood plains and swampy interfluves," by Meng and Harsh (1988). Furthermore, the clay-dominated zones are specifically described as "overbank deposition that was modified by weathering and diagenesis," and are considered to be "distinctly continental in origin" (Meng and Harsh, 1988).

While the middle Potomac confining unit is not a single, continuous layer, it is represented as one in regional groundwater flow models (Harsh and Laczniak, 1990). Vertical hydraulic conductivity of the unit was measured from laboratory analysis of a single core, producing a value of 3.4×10^{-6} ft/d, or 1.0×10^{-6} (Harsh and Laczniak, 1990). Of course, a single measurement like this would be expected to produce an artificially low value given the discontinuous nature of this unit, and groundwater models confirm this expectation. Computer simulations of this system produced what is probably a more realistic calibrated average value of

4.06×10^{-5} ft/d for the middle Potomac confining unit as a whole (Harsh and Laczniak, 1990). An average specific storage value for this confining unit was estimate via model calibration as 1.0×10^{-6} ft^{-1}, though it should be noted that groundwater simulations have not yet considered the effects of the release of water from compressible deposits (Harsh and Laczniak, 1990).

Upper Potomac Aquifer

The Brightseat-upper Potomac aquifer was originally defined by Meng and Harsh (1988) as a continuous wedge of sandy sediments overlying the middle Potomac confining unit and overlain by the upper Potomac confining unit (figure 3-7). The wedge-shaped aquifer of Late Cretaceous age defined by Meng and Harsh thickened from its western edge roughly 20 to 30 miles east of the Fall Line to a maximum of over 400 feet in eastern Virginia, and it was interpreted as representing the effects of the "first major marine transgression" across the underlying continental delta complex (1988). Over a decade of further study since the work of Meng and Harsh has revealed the geology and the associated hydrogeology of this interval to be significantly more complex than what they described. Much of the complexity in the interpretation of this interval apparently results from the transitional nature of the depositional environment, which has been further complicated by the influence of the Norfolk arch and James River structural zone, as well as the disruption attributed to the impact crater. Evidence for this increased complexity is documented by Powars and Bruce (1999) and Powars (2000) and suggests that the Brightseat-upper Potomac aquifer may be further subdivided in the future to reflect spatial variations in the depositional environment and resultant hydrogeologic properties (McFarland, oral communication, 2001). Consequently, some of the increased complexity described by various authors is outlined here, particularly as it pertains to the hydrogeology of southeastern Virginia.

In Virginia, no sediments of Late Cretaceous age have been found west of the Chesapeake Bay and north of the James River (Powars and Bruce, 1999), so the upper Potomac aquifer is no longer thought to exist in that area (McFarland, oral communication, 2001). Instead, these sediments are known to be either Early Cretaceous or Paleocene in age (Powars and Bruce, 1999). Of course, the age of a sedimentary interval should not necessarily determine its hydrogeologic delineation, but in this case the age relation is thought to be somewhat indicative of hydrologic continuity, and these sediments are therefore classified as belonging either to the

Middle Potomac aquifer of continental origin or the Aquia aquifer of marine origin (McFarland, oral communication, 2001). In any case, this issue is not particularly relevant to the hydrogeology of southeastern Virginia. Similarly, the Brightseat Formation of Early Paleocene age is no longer thought to exist in Virginia, so the Brightseat is no longer thought to be a part of the upper Potomac aquifer or confining unit (Hamilton and Larson, 1988).

South of the James River, the sediments originally described as the upper Potomac aquifer are of Late Cretaceous age and predominantly marine in origin, and the descriptions of Meng and Harsh (1988) may be more or less accurate, if not complete. A complete characterization of the increased complexity of this interval is again beyond the scope of this work, but some discussion is included here because the increased detail provides a somewhat better representation of spatial variability in hydrologic properties and behavior. The Upper Cretaceous deposits south of the James were subdivided by Powars (1992, 2000) into three lithologic units. He described (1) upper Cenomanian marine and deltaic deposits overlain by (2) a loose glauconitic sand unit and (3) "oxidized red-bed fluvial-deltaic deposits that include multiple paleosols" (Powars, 2000). The vertical relations between these units can be evaluated from figure 3-3 and the chart in figure 3-2. Hydrogeologically, the upper Cenomanian beds correspond with the upper Potomac aquifer described by Meng and Harsh (1988); the glauconitic sand unit corresponds with the Virginia Beach aquifer described by Hamilton and Larson (1988); and the red beds correspond roughly with either the upper Potomac confining unit of Meng and Harsh (1988) or the Virginia Beach confining unit of Hamilton and Larson (1988). For clarification, it should be noted that Hamilton and Larson recognized an eastward part of the upper Potomac confining unit of Meng and Harsh as the separate Virginia Beach aquifer, resulting in the arrangement depicted in figure 3-2. The Virginia Beach aquifer is described separately on following pages.

As described by Meng and Harsh (1988) deposition of the upper Potomac aquifer "occurred in everwidening estuaries and intertidal basins," as the seas gradually and progressively inundated the area (Meng and Harsh, 1988). Specific environments of deposition postulated by other authors include a littoral, possibly tidal flat environment with a protected shoreline, or a "marginal marine outer-delta environment with a vegetated, swampy shoreline" (Meng and Harsh, 1988). Because of the change in depositional environment, revisions in progress will likely distinguish the sediments of this aquifer with a name distinct from the other

Potomac aquifers, which belong to the Potomac formation of continental origin (McFarland, oral communication, 2001).

The sandy units of the upper Potomac aquifer, as described from driller's logs by Meng and Harsh (1988), are "characteristically white, micaceous, very fine to medium quartz, and commonly contain carbonaceous material." The interbedded clays are described by drillers as dark, silty, and highly micaceous, commonly containing carbonaceous material. Available lithologic reports of core samples indicate that the upper Potomac aquifer consists of very fine to medium, "yellow to greenish-gray" micaceous quartz sands that are "poor to well sorted," and "subrounded to subangular," commonly containing wood fragments and "interstitial clays (Meng and Harsh, 1988). Lithologic descriptions of interbedded silty-clay intervals describe them as "yellow green to dark greenish gray, glauconitic, calcareous, micaceous, plastic," and locally sandy, commonly containing shell fragments (Meng and Harsh, 1988). Drillers commonly note that these sediments are easily penetrated, according to Meng and Harsh. Geophysical logs through the Upper Potomac aquifer show patterns similar to those found in the middle Potomac aquifer, but they are "more massive and rounded in profile," indicating more massively bedded sands and thinner interbedded clays (Meng and Harsh, 1988). As in the middle Potomac aquifer, the interlensing nature of the sands and clays and the difficulty in correlating the individual layers cause authors to qualify given elevation and thickness measurements as approximate. Nonetheless, the layers of the upper Potomac aquifer are described as easier to correlate spatially than those of underlying units, probably due to changes in the depositional environment from generally continental to marine (Meng and Harsh, 1988).

For the purposes of this study and the several groundwater modeling studies of the area, the upper Potomac aquifer can be thought of as overlying the middle Potomac confining unit, but it is not always present. The overlying unit varies spatially across the Coastal Plain of southeastern Virginia. It is usually the Paleocene Aquia aquifer, but the overlying unit may also be the Peedee aquifer or the Virginia Beach aquifer, depending on location. Collectively, the predominantly marine sediments of the unit referred to as the upper Potomac aquifer by Harsh and Laczniak reach a maximum thickness of approximately 425 ft in the northeastern part of the Coastal Plain of Virginia (1990).

The upper Potomac aquifer is characterized hydrogeologically by high transmissivity throughout most of its extent, though values decrease somewhat toward the coast. The decrease in transmissivity was originally attributed to a facies change from sands to silts and clays, but it may also be due to the effects of the impact crater boundary. Regardless of the cause, maximum transmissivities can be found in southeastern Virginia, where values from aquifer tests range from 1,500 to 13,000 ft^2/d (Harsh and Laczniak, 1990). In other areas of Virginia, reported values range from a minimum of about 150 ft^2/d in the north-central Coastal Plain to 2,000 ft^2/d near the York River at West Point (Harsh and Laczniak, 1990). Flow-model calibrations produced a transmissivity range of 330 to over 4,000 ft^2/d (Harsh and Laczniak, 1990), while revised model values approach 9,000 ft^2/d in some locations (McFarland, 1998). The only available value for the storage coefficient of this aquifer is provided by Harsh and Laczniak (1990) in their modeling study of the Virginia Coastal Plain. Their reported mean value of 5×10^{-4} is calculated from several earlier aquifer tests. The same source estimates the average lateral hydraulic conductivity of the upper Potomac aquifer to be approximately 25 ft/d.

The upper Potomac aquifer is the principal source of water for municipal, industrial, and agricultural use in the eastern part of the Virginia Coastal Plain (Harsh and Laczniak, 1990). Wells completed in the aquifer are reported to yield from 25 to 350 gal/min, according to several sources, and numerous wells tap the aquifer (Harsh and Laczniak, 1990). With total reported withdrawals from the upper Potomac aquifer in Virginia currently approaching 20 million gallons per day, the aquifer ranks second behind only the middle Potomac as a source of water for the Coastal Plain. As a result of this continued development, water levels in the upper Potomac aquifer have declined from historic highs of over 20 ft above sea level to current lows of 20 to 30 ft below sea level (Harsh and Laczniak, 1990). Continued withdrawals are expected to lead to further declines in water levels.

Upper Potomac Confining Unit

As with the upper Potomac aquifer, the Brightseat-upper Potomac confining unit is no longer thought to include the Brightseat Formation (Hamilton and Larson, 1988; McFarland, oral communication, 2001). The Brightseat aquifer and confining units, described by Meng and Harsh (1988), were once thought to occur across the northeast quarter of the Virginia Coastal Plain. Throughout their extent, they were described as overlying the upper Potomac aquifer and confining unit, but these sediments are now thought to belong to the Aquia aquifer, or perhaps the underlying upper Potomac. Furthermore, the upper Potomac confining unit described by

Meng and Harsh (1988) as extending across the Virginia Coastal Plain is now thought include sediments of Early Cretaceous age north of the James River.

In other words, the proper “upper Potomac confining unit of Late Cretaceous age” only exists south of the James River, while sediments of at a similar depth interval belong to the Middle Potomac confining unit of Early Cretaceous age” north of the James. Regardless of origin or name, these sediments together form the hydrogeologic model unit referred to by Harsh and Laczniak (1990) as the Brightseat-upper Potomac confining unit, which consists of sediments of possibly late Early Cretaceous to early Paleocene age directly underlying the Paleocene sediments of the Aquia aquifer across the Coastal Plain of Virginia. Perhaps the revisions currently underway will sort out this confusion. For now, this interval will be collectively described as presented by Meng and Harsh (1988) and Hamilton and Larson (1988).

The upper Potomac confining unit varies in thickness throughout the Virginia Coastal Plain, but it attains its maximum known thickness of 192 ft in southeastern Virginia (Hamilton and Larson, 1988). The estimated average thickness of this unit used in groundwater models is 35 ft (Harsh and Laczniak, 1990). In general, this unit thickens and dips to the east. As described by Hamilton and Larson, this clay unit continues across almost the entire Coastal Plain in Virginia, giving the rather large combined areal extent (1988).

The sediments of the upper Potomac confining unit are collectively described as “dark-green to black, highly micaceous silty clay with oxidized red to yellow thin clay” (Harsh and Laczniak, 1990). Other descriptions of core note that this clay unit is also calcareous, slightly glauconitic, and sandy (Meng and Harsh, 1988). More specifically, the clays consist of “highly expandable silty-clay to clayey-silt mixed-layer illite and montmorillonite, and minor amounts of kaolinite” (Meng and Harsh, 1988).

As with the lower and middle Potomac confining units, the clays of this confining unit form a series of interlayered deposits rather than a single, continuous layer (Meng and Harsh, 1988). However, these layers and lenses of clay are more extensive than those found in the lower confining units, so they can generally be more easily correlated between wells (Meng and Harsh, 1988). The deposition of these clays is interpreted as the result of the “first major marine transgression in the sedimentary section,” and it probably occurred in a “broad, low-lying outer delta” environment (Meng and Harsh, 1988). It now appears that this description may be most accurate for the part of the unit found south of the James River, which may be correlated with the upper Cenomanian beds described more recently by Powars (2000).

Laboratory analyses of core from this confining unit in Maryland yielded a vertical hydraulic conductivity value of 8.64×10^{-4} ft/d, as well as a specific storage value of 7.4×10^{-5} ft^{-1} (Harsh and Laczniak, 1990). The average vertical hydraulic conductivity value for this unit obtained from calibration of model simulations was about an order of magnitude lower, at 4.41×10^{-5} ft/d, but the specific storage value from the calibrated model is similar to the value from field measurements, at 1.0×10^{-6} ft^{-1} (Harsh and Laczniak, 1990).

Virginia Beach Aquifer

Composed of unnamed Upper Cretaceous sediments, the Virginia Beach aquifer was first recognized and named by Hamilton and Larson (1988). This aquifer unit is present only south of the James River in Virginia and extends southward into northeastern North Carolina, where it is equivalent to the Black Creek Formation (Hamilton and Larson, 1988) (figure 3-8). The Virginia Beach aquifer pinches out to the west approximately 30 miles east of the Fall Line. It's northern edge, apparently controlled by the southern flank of the Norfolk arch, follows a line extending from Virginia Beach southwestward to Suffolk, then just south of Franklin (Hamilton and Larson, 1988). As a result, it is uncertain whether this unit is present at the location of the subsidence wells at Franklin and Suffolk. In Virginia, the Virginia Beach aquifer dips to the east and reaches a maximum thickness of approximately 110 feet at the city of Chesapeake (Hamilton and Larson, 1988). Throughout its extent in Virginia, this aquifer and its confining unit are overlain by the upper Paleocene Aquia aquifer. In northeastern North Carolina, the overlying unit is the Peedee Aquifer.

The fine-grained to medium-grained glauconitic sands of the Virginia Beach aquifer are interbedded with "thin clay layers and indurated zones," and commonly contain shell material (Hamilton and Larson, 1988). The sediments of the Virginia Beach aquifer described by Hamilton and Larson (1988) correlate approximately with the Upper Cretaceous glauconitic sand unit more recently described by Powars (2000). Powars describes the glauconitic sand unit as loose, well-sorted, fine to very coarse glauconitic marine sand with very little clay or silt (2000), which indicates that this glauconitic sand may be only part of the aquifer described by Hamilton and Larson (1988).

Very little published data are available regarding the hydrogeologic properties of the Virginia Beach aquifer. However, the Virginia Beach aquifer is part of a horizon that was previously thought to be part of the upper Potomac, which is also composed of marine sands. Consequently, the hydrologic properties outlined for the upper Potomac aquifer are probably representative of the Virginia Beach aquifer.

The Virginia Beach aquifer is described by Hamilton and Larson as capable of producing "moderate to abundant quantities of generally good quality" for domestic and industrial uses (1988), though only from its up-dip area. In the thickest eastern parts of the aquifer, the water is too brackish to be used as a water supply (Barry Smith, USGS, verbal communication, 2001).

Virginia Beach Confining Unit

Little data are available regarding the character of the Virginia Beach confining unit, which is described by Hamilton and Larson as a series of clay, silty clay and sandy clay beds overlying the Virginia Beach aquifer throughout its extent (1988). According to their description, this unit ranges in thickness from less than 10 feet along its updip edge to almost 30 feet near the city of Virginia Beach (Hamilton and Larson, 1988). It now appears that this unit is likely part of a geologic unit defined by Powars (2000). He identifies Upper Cretaceous red beds overlying the glauconitic sand unit in southeastern Virginia (Powars, 2000). These red beds are described as having a heterogeneous lithology of interbedded oxidized clay, silty fine sand and coarse sand (Powars, 2000). Powars also describes fining-upward sequences that include clayey paleosols (2000). These red beds and the underlying glauconitic sands form the Virginia Beach confining unit and aquifer; this relation can be seen in Figure 3-2. The red beds are described as having a "fluvial to upper-delta-plain depositional environment" because of their extensive development (Powars, 2000).

The Virginia Beach confining unit includes sediments that were formerly defined as part of the upper Potomac confining unit. As a result, the hydrogeologic characteristics of this unit are likely very similar to those of the upper Potomac, but very little specific data are available on the characteristics of the Virginia Beach confining unit. In general, the extent of the Virginia Beach aquifer and confining unit are not particularly well defined. As Meng and Harsh point out, it is relatively easy to distinguish the Cretaceous sediments of marine origin from the underlying fluvial-deltaic sediments, using geophysical logs and available lithologic data (1988). On the other hand, it appears to be much more difficult to distinguish the marine sediments from each other. This has resulted in some misunderstanding of the relation between the upper Potomac and Virginia Beach aquifers and confining units. Figure 3-3 and the chart in figure 3-2 outline the current understanding of the system.

Peedee Aquifer

The Peedee aquifer and overlying Peedee confining unit are present only in the northeastern corner of North Carolina and are not found in Virginia (figure 3-9). These units are Late Cretaceous in age and stratigraphically overlie the Virginia Beach aquifer and confining unit throughout their extent. In thickness and hydrogeologic properties, these units are very similar to the Virginia Beach units (Hamilton and Larson, 1988). They are mentioned here only to provide completeness to the correlation of hydrogeologic units presented in Figure 3-2. More detailed information is provided by Hamilton and Larson (1988).

Aquia Aquifer

The Aquia aquifer is described by Meng and Harsh as the “predominantly sandy facies of the Aquia Formation” of late Paleocene age (1988). It extends across most of the Coastal Plain of Virginia and correlates with similar units in Maryland and North Carolina. The Aquia aquifer overlies either the upper or middle Potomac confining unit and is overlain by the Nanjemoy-Marlboro confining unit throughout its extent (Meng and Harsh, 1988). The Aquia aquifer attains a maximum thickness of almost 150 feet; values range from about 40 to 70 feet throughout much of its extent (Meng and Harsh, 1988), though lower values are reported for most locations in southeastern Virginia (Powars, 2000), (figure 3-10). It extends to a thin edge just east of the Fall Line, and it crops out in many stream valleys in the western part of the study area. Like most of the other units of the Virginia Coastal Plain, the Aquia aquifer dips to the east. Unlike many of the other Coastal Plain aquifers, however, the Aquia aquifer is described as thinning to the east as the result of a facies change in the late Paleocene sediments (Meng and Harsh, 1988). It is now known that the Aquia aquifer is also truncated to the east by the boundary of the Chesapeake Bay impact crater (Powars, 2000). Within the crater boundary, sediments once thought to be part of the Aquia aquifer are now known to be part of the Exmore tsunami-breccia deposits (McFarland, USGS, verbal communication, 2001).

The Aquia aquifer is described as a "continuous, elongate-lenticular sand body" (Meng and Harsh, 1988). It is a "massively bedded unit composed of very fine to medium glauconite and quartz sands" with minor amounts of shells and clay (Meng and Harsh, 1988). The sands are described in more detail by various authors as "greensands and greensand marls," and "very well-sorted, medium to dark green, massive, fine to medium glauconitic sand (Meng and Harsh, 1988). Most authors recognize sparse shelly intervals, carbonate cemented intervals, some thin clay intervals, and a basal section containing sand, pebbles, and fish bones and teeth (Meng and Harsh, 1988). Mica, pyrite, and lignitic material are also reported (Powars, 2000). The sediments of the Aquia aquifer are often described colloquially as running or caving sands due to their soft nature (Meng and Harsh, 1988). Unlike many of the underlying units, the Aquia aquifer can be easily correlated across large distances due to distinctive geophysical log signatures as well as very distinctive clay units which mark the top and bottom of the Aquifer sands on drillers' logs (Meng and Harsh, 1988). Most sources suggest that the depth to the top of the aquifer can be predicted with confidence between control wells due to its consistency (Meng and Harsh, 1988; McFarland, verbal communication, 2001).

The consistent lithology of the Aquia aquifer is due to its relatively consistent depositional environment. According to the summary provided by Meng and Harsh, various authors suggest that the unit was deposited "in a shallow, inner shelf marine basin, below wave base, with slight fluctuation of water depths" from about 100 to 300 ft (1988). Similarly, Hamilton and Larson describe the environment of deposition as "shallow marine waters" (1988).

The transmissivity of the Aquia aquifer has been measured from single-well and specific capacity tests, and values are reported to range from 125 to 1000 ft^2/d. Of course, values vary with thickness, though an eastward facies change was also invoked to explain lower transmissivities in that direction (Harsh and Laczniak, 1988). These variations are now thought to be almost certainly the effects of the Chesapeake Bay impact crater. No data on the storage coefficient are available for the Aquia aquifer in Virginia, but a range of 1.0×10^{-4} to 4.0×10^{-4} is reported for the Aquia aquifer in Maryland (Harsh and Laczniak, 1990). Estimated and modeled hydraulic conductivity values range from about 25 to 40 ft/d (Harsh and Laczniak, 1990).

The Aquia aquifer is a consistent and accessible source of water for a variety of small-volume users. Numerous wells penetrate the aquifer across the Coastal Plain and within the study area, though use of water from the Aquia is more common north of the James River, where the aquifer is thicker and more sandy (Meng and Harsh, 1988). Wells in the Aquia are reported by various authors to yield between 20 and 110 gallons per minute (Harsh and Laczniak, 1990), though they are reported to exceed 550 gal/min (Kull and Laczniak, 1987). The Aquia aquifer provides a water source for light industrial, small municipal and industrial users. Withdrawals total a few million gallons per day for the Coastal Plain, just a few percent of groundwater withdrawals from the Virginia Coastal Plain (Kull and Laczniak, 1987). Small but consistent increases in withdrawals have been observed in recent decades.

Nanjemoy-Marlboro Confining Unit

The Nanjemoy-Marlboro confining unit is composed of the clay deposits of two distinct formations that together overlie the Aquia aquifer throughout its extent, except where the aquifer crops out along its western edge. The Marlboro Clay of late Paleocene age directly overlies the Aquia and is itself overlain by the Nanjemoy Formation of early Eocene age (Meng and Harsh, 1988). Together, these formations make up a generally wedge-shaped unit that thickens from near zero along its western edge near the Fall Line to over 150 ft on the northern part of the Eastern Shore Peninsula (Meng and Harsh, 1988). Its maximum thickness in southeastern Virginia is 62 feet (Hamilton and Larson, 1988). The Nanjemoy-Marlboro confining unit is described as very continuous, and its top elevation can usually be predicted with some confidence between control wells (McFarland, oral communication, 2001). Furthermore, the sediments of this unit are easily recognized on both resistivity and drillers' logs because of their characteristic thick clay pattern and distinctive colors; the contacts with the overlying and underlying sandy aquifers are also distinctive and easily recognized (Meng and Harsh, 1988).

While the confining unit as a whole is described as easy to identify, the two units that compose it are sometimes difficult to distinguish from each other, and there appears to be some uncertainty regarding their areal extents. Meng and Harsh (1988) describe the Marlboro Clay as pinching out against the northern flank of the Norfolk arch, but Powars (2000) finds this unit in analysis of several core holes in southeastern Virginia. On its eastern edge, this unit is truncated by the Chesapeake Bay impact crater and its infilled sediments.

The Marlboro Clay is composed of a distinctive red to gray mottled clay composed primarily of kaolinite and illite (Meng and Harsh, 1988). It is described as massively bedded and extensively burrowed, with irregular lenses of fine silt (Reinhardt and others, 1980). Drillers often refer to it as “slick or sticky” and “pink, gray or white (Meng and Harsh, 1988). It ranges in thickness from about 2 to 20 ft across the Coastal Plain (Meng and Harsh, 1988). The Nanjemoy Formation is composed of a “variably clayey and shelly, micaceous greensand,” which is dark-green to brown-green and massively bedded (Meng and Harsh, 1988). Extensive burrows give the unit what is described as a mottled appearance (Meng and Harsh, 1988). The clay content is mostly illite and ranges from 15 to 80 percent (Meng and Harsh, 1988). Drillers commonly note the presence of shells and black sands (Meng and Harsh, 1988).

Distinct environments of deposition for the upper Paleocene Marlboro clay and the lower Eocene Nanjemoy clay have been inferred from intensive analysis of core (Reinhardt et al., 1980). The Marlboro clay is thought to have formed in a brackish, low-energy basin such as an estuary or lagoon (Meng and Harsh, 1988). The Nanjemoy, on the other hand, was deposited in an environment of somewhat higher energy, such as a protected inner middle marine shelf (Meng and Harsh, 1988).

The average vertical hydraulic conductivity value for this unit from calibration of model simulations was 5.36×10^{-5} ft/d. Vertical hydraulic conductivity values as determined from laboratory analysis of core range from 2.0×10^{-5} to 2.2×10^{-6} (Harsh and Laczniak, 1990). This confining unit is known to be much more areally extensive than some of the underlying units already described, so laboratory analyses would be more likely to produce somewhat better values for the behavior of the unit as a whole. Model calibrations used an estimated value of 1×10^{-6} ft^{-1} for the specific storage of this confining unit, but no other storage data are available.

Chickahominy-Piney Point Aquifer

Definition of the Chickahominy-Piney Point aquifer (figure 3-11) is not possible without some consideration of the role of the bolide impact and the crater it formed. The aquifer, as described by Meng and Harsh, was thought to be composed of the sandy sediments of the middle Eocene Piney Point Formation and the overlying Chickahominy Formation (1988). Meng and Harsh recognized changes in hydrogeologic properties to the east and postulated an eastward facies change from sand to clay (1988). Some amount of change in facies may be apparent in the undisturbed Coastal Plain sediments, but it is now known that the crater boundary is primarily responsible for the observed changes in hydrogeologic properties (Powars and Bruce, 1999; Powars, 2000). Furthermore, recent work by Powars and Bruce (1999) and Powars (2000) identified the Chickahominy Formation only within the boundary of the impact crater; it is described as the first post-impact deposit.

The Chickahominy Formation as it is now described is approximately adjacent to and continuous with the Piney Point formation at the crater boundary, but the change in 'facies' is now known to be relatively abrupt rather than gradational. Outside of the crater boundary, it is probably more correct to refer to this unit simply as the Piney Point aquifer, though Meng and Harsh note that the aquifer also contains some sediments of late Oligocene and early Miocene age, collective known as the Old Church Formation. These younger sediments are primarily fine-grained, white quartzose sands with some glauconite and shells (Meng and Harsh, 1988). Within the crater, the Chickahominy functions mostly as a hydrologic barrier, and would most properly be called the Chickahominy confining unit.

Like most of the Coastal Plain units, this aquifer is described as wedge-shaped in cross section; it thickens from its western edge near the Fall Line to a maximum known thickness of approximately 165 ft in eastern Virginia (Meng and Harsh, 1988). It overlies the undisturbed Nanjemoy-Marlboro Clay confining unit and is "overlain and transgressed by" the Calvert confining unit (Meng and Harsh, 1988). However, this description is probably most accurate for the Piney Point aquifer north of the James River; south of the James, this aquifer unit thins considerably and becomes much more difficult to identify and correlate (Powars, 2000). In fact, the only sediments of this unit identified in core holes south of the James are the thin Oligocene-to-Miocene sands known as the Delmarva beds; the sands of the Piney Point Formation are not identified in these cores, though they are found in thin intervals in other locations south of the James (2000). The Piney Point sands formerly identified in southeastern Virginia as middle Eocene in age are now inferred to be sands of the underlying Aquia Formation (Powars, 2000), but this issue transcends the scope of this research.

The Piney Point Formation is thought to be the result of a marine transgression, and the environment of deposition is described as "a shallow, inner to middle marine shelf dominated by longshore currents" (Hansen, 1972; Meng and Harsh, 1988). The Chickahominy Formation is now thought to be post-impact crater fill, and what little is known of the bolide impact provides some understanding of the extreme disruption to the geology of this area (Powars, 2000).

Estimated and modeled hydraulic conductivity values range from about 25 to 35 ft/d for the central part of the Piney Point aquifer (Harsh and Laczniak, 1990). Transmissivity values, calculated from aquifer tests and specific-capacity tests, vary significantly; the median value for southeastern Virginia is reported as just under 3,000 ft^2/day (Hamilton and Larson, 1988). A specific storage value of 1 x 10^{-4} ft^{-1} was assumed for this unit, but no other storage data are available (Hamilton and Larson, 1988). It is interesting to note that the changes in hydrogeologic properties resulting from the crater boundary were partially revealed by model simulations long before the existence of the crater was discovered (McFarland, USGS, verbal communication, 2001).

The Chickahominy-Piney Point aquifer is reported to be a relatively small but important source of water for domestic, small industrial and municipal supplies in the central part of the Virginia Coastal Plain (Kull and Laczniak, 1987). Water use from this aquifer for the entire Coastal Plain totaled just over 3 million gallons/day in 1983, and usage has increased only slightly since that time (Virginia DEQ water-use database, 2001). Wells typically yield 10 to 150 gal/min, though in some cases yields may exceed 350 gal/min (Kull and Laczniak, 1987). Due to the thin nature of the sands, this aquifer is rarely used in southeastern Virginia; both the underlying Aquia aquifer and the overlying Yorktown-Eastover aquifer provide much more dependable water supplies.

Calvert Confining Unit

The Calvert confining unit is composed of the fine-grained sediments of the Calvert Formation, the oldest member of the Miocene to middle Pliocene Chesapeake Group (Meng and Harsh, 1988). These early to middle Miocene sediments overlie the Piney Point aquifer throughout most of its extent and are overlain primarily by the Saint Marys confining unit, except for where they are overlain by the Aquia along their western edge, and perhaps by the Saint Marys aquifer in the far northeast part of the state. The Calvert crops out along its western edge in numerous stream valleys near the Fall Line (Meng and Harsh, 1988). Like the rest of its group, this unit is wedge-shaped in cross section and thickens and dips to the east from a thin edge, reaching a maximum thickness of about 350 ft on the northern part of the Eastern Shore peninsula of Virginia (Meng and Harsh, 1988). It is much thinner on average, with a thickness ranging from 30 to 70 feet throughout most of its extent (Meng and Harsh, 1988). The Calvert confining unit's composition is typically described as interbedded shelly sandy clay, silty clay, and diatomite, which have a "dark grayish-green" color (Meng and Harsh, 1988). From analysis of core, the unit is described in more detail as a fine-textured gray sediment with an upper section of very fine quartz sand in a silt to clay matrix overlying a lower interval composed of a thin diatomite layer and a basal clay intermixed with quartz sand (Meng and Harsh, 1988). A distinctive basal lag deposit is composed of quartz sand, phosphate pebbles, sharks' teeth, shells, and bones; it marks the contact with the underlying Piney Point aquifer (or, in the crater, the Chickahominy) (Meng and Harsh, 1988; Powars and Bruce, 1999).

The depositional environment of the Chesapeake Group in general is described by Meng and Harsh as "a shallow, low-energy inner-shelf marine basin that was below wave base," which explains the predominance of silts and clays in this interval (1988). The Calvert confining unit is the most fine-grained of the units composing this group, indicating that it was formed in a very low-energy environment. Its particular depositional environment is inferred to be a "siliceous, inner to middle-marine shelf that oscillated between semiprotected embayment to open ocean circulation" (Reinhardt and others, 1980; Meng and Harsh, 1988). Probably due to a relatively consistent depositional environment, the Calvert and the others of its group are described by various authors as being easy to identify on geophysical logs and of predictable thicknesses across long distances between control wells.

The average vertical hydraulic conductivity value for this unit from calibration of model simulations was 1.12×10^{-4} ft/d. The estimated vertical hydraulic conductivity used in model analysis of southeastern Virginia was 3.89×10^{-5} ft/day for the Calvert confining unit (Hamilton and Larson, 1988). In contrast, the vertical hydraulic conductivity of a single core taken from Isle of Wight County was to 9.2×10^{-6} ft/d (Harsh and Laczniak, 1990). The higher model values probably result from the effects of buried paleochannels along river valleys, which incised this confining unit (and others above it) in the Pliocene and subsequently filled those incisions with coarse-grained sediments (Harsh and Laczniak, 1990).

Saint Marys-Choptank Aquifer

The Saint Marys-Choptank aquifer is made up of sandy sediments of the middle-Miocene Choptank Formation and the overlying middle-to-late Miocene Saint Marys Formation (Meng and Harsh, 1988). These sediments lie almost entirely beneath the Eastern Shore peninsula of Virginia (Fig. 3-12), and are not found elsewhere in the state. This wedge-shaped unit reaches a maximum known thickness of over 160 feet near its eastern limit, but it is much thinner throughout most of its extent and pinches out quickly to the west and to the south (Harsh and Laczniak, 1988). Very little well data exists for this aquifer, so much of its extent is estimated and inferred (Meng and Harsh, 1988).

The fine to medium quartzose sands of this aquifer often contain shells and layers of clay and silt (Meng and Harsh, 1988). Their deposition is inferred to be the result of a delta building outward into the shallow inner shelf of the Salisbury embayment (Gibson, 1982; Meng and Harsh, 1988).

There are no known users of the Saint Marys-Choptank aquifer in Virginia, as much better sources of water are located at lesser depths (Harsh and Laczniak, 1988). The water of this depth on the Eastern Shore is too high in chloride for most uses, and few wells penetrate this deep. Estimated and modeled hydraulic conductivity values range from about 10 to 15 ft/d, but data for this aquifer are very limited (Harsh and Laczniak, 1990).

Saint Marys Confining Unit

The Saint Marys confining unit comprises fine-grained sediments of the Saint Marys Formation and the lower part of the overlying Eastover Formation, both of Miocene age (Hamilton and Larson, 1988). This confining unit overlies the Calvert confining unit, except for the extreme northeastern section, where it overlies the Saint Marys aquifer (figure 3-12). It is overlain by the Yorktown-Eastover aquifer throughout its extent (Hamilton and Larson, 1988). Wedge-shaped in cross section, this unit dips eastward from its western edge 30 to 40 miles east of the Fall Line, reaching a maximum known thickness of over 300 feet in the northern part of the Eastern Shore peninsula (Meng and Harsh, 1988). Both Meng and Harsh (1988) and Powars (2000) report thicknesses of 10 to over 100 ft in southeastern Virginia, but Hamilton and Larson (1988) do not map this unit south of the James (probably defining the clay as part of the Calvert Formation).

The sediments of the Saint Marys confining unit consist of interbedded silty and sandy clay, with some reported shelly material, including a basal shelly layer (Meng and Harsh, 1988; Powars, 2000). The Saint Marys formation is described as dominantly clay in composition and bluish-gray to gray in color, while the lower part of the Eastover formation is said to be composed of poorly sorted sandy clay and is greenish-gray in color (Meng and Harsh, 1988).

The upper contact of this unit with the overlying Yorktown-Eastover aquifer is marked by a gradual transition to more sandy sediments (Meng and Harsh, 1988).

The sediments of this confining unit belong to the upper part of the Chesapeake Group described previously and in southeastern Virginia form the uppermost section of a thick series of predominantly fine-grained sediments; the very thin Piney Point is the only sandy interval in a sequence dominated by the Nanjemoy-Marlboro confining unit, the Calvert confining unit, and the Saint Marys. The environment of deposition for the Saint Marys itself is described as a broad, shallow inner-shelf area, which was likely open marine to partially embayed (Meng and Harsh, 1988).

The average vertical hydraulic conductivity value for this unit from calibration of model simulations was 4.15 x 10^{-4} ft/d, while vertical hydraulic conductivity values as determined from laboratory analysis of core range from 2.0 x 10^{-5} to 2.8x 10^{-6} ft/d (Harsh and Laczniak, 1990). Again, this discrepancy probably reflects the effects of Pleistocene incisions that produced areas of higher vertical hydraulic conductivity along buried paleochannels (Harsh and Laczniak, 1990). Very little storage data are available for this or other confining units, which were simulated with an average specific storage of 1 x 10^{-6} ft^{-1}, and none of the simulations thus far account for the effects of confining unit storage (Harsh and Laczniak, 1990).

Yorktown-Eastover Aquifer

The Yorktown-Eastover aquifer is made up of the sandy sediments of the Eastover Formation of Miocene age, the Yorktown Formation of early Pliocene age, and the Bacons Castle Formation of late Pliocene age (Meng and Harsh, 1988). Together, these sediments form a wedge-shaped aquifer that thickens and dips eastward from a thin western edge, reaching a maximum thickness of almost feet in southeastern Virginia at Virginia Beach (Hamilton and Larson, 1988) (figure 3-13). It is generally thinner and more dissected in the north, west and central parts of the Virginia Coastal Plain, where it has been more significantly affected by erosion (Meng and Harsh, 1988). Due to the effects of erosion, the western boundary of the aquifer is highly irregular (figure 3-13). The Yorktown-Eastover aquifer overlies either the Saint Marys or Calvert confining unit and is overlain by the Yorktown confining unit except where it is unconfined. The aquifer is generally unconfined in the western part of the Coastal Plain; it crops out in a broad strip along the western uplands of the Coastal Plain and also in the stream valleys of the central and southeastern parts (Meng and Harsh, 1988). Throughout the eastern part of the Virginia Coastal Plain the aquifer ranges in thickness from about 100 to 200 ft (Meng and Harsh, 1988). The Yorktown-Eastover aquifer is relatively continuous to the north and south, and is said to correlate with aquifers in Maryland and North Carolina (Meng and Harsh, 1988).

The general lithology of the Yorktown-Eastover aquifer is described as “interlayered, thick to massively-bedded shelly sands separated by thinner clay beds,” but its lithology varies significantly across its extent (Meng and Harsh, 1988). Thin and areally discontinuous along it westward edge, the clay beds become “massive and extensive” toward the south and east, dividing the aquifer into three distinct units (Meng and Harsh, 1988). These clay beds are said to be extensive and easily correlated between wells (Meng and Harsh, 1988). Each of the three formations that make up the Yorktown-Eastover aquifer consists of a fining-upward sequence of sediments, with a basal facies of coarse sand and gravel, a middle fine to medium shelly sand facies, and an uppermost very fine silty clay (Meng and Harsh, 1988). The sands and clays are described as gray to yellow in color, and some hard shell beds are reported (Meng and Harsh, 1988).

The facies changes described above are inferred to represent marine transgressions resulting in shallow bays (Meng and Harsh, 1988). Various authors cited by Meng and Harsh describe a depositional environment consisting of a large, shallow, embayed shelf alternately exposed and submerged by regressing and transgressing seas (1988).

Estimated and modeled hydraulic conductivity values range from about 15 to 20 ft/d (Harsh and Laczniak, 1990). Transmissivity values in the Yorktown-Eastover, calculated from several sources, range from 200 to 3000 ft^2/d, generally increasing with thickness toward the east (Harsh and Laczniak, 1990). Storage coefficients for the aquifer are reported to range from 6.0×10^{-3} to 7.7×10^{-4} (Harsh and Laczniak, 1990).

Due to its shallow depth and relatively high thickness, the Yorktown-Eastover aquifer is an accessible and important water supply throughout the Coastal Plain of Virginia, and it is used for domestic, commercial and light industrial purposes (Hamilton and Larson, 1988). It is especially important in coastal areas of southeastern Virginia, where it is the largest source of non-brackish water. In fact, even the lower part of this unit tends to be high in chloride in coastal areas (Harsh and Laczniak, 1990). Withdrawals from the Yorktown-Eastover aquifer have consistently comprised about 5 percent of the total from confined aquifers of the Coastal Plain in recent years, and they have remained relatively steady of late, ranging from about 5 to 8 million gallons/day (Kull and Laczniak, 1987). Wells typically yield from 5 to 500 gallons/min, though yields may exceed 1000 gallons/min in some cases (Kull and Laczniak, 1987).

Yorktown Confining Unit

The Yorktown confining unit comprises the fine-grained sediments of the upper parts of the Yorktown and Bacons Castle Formations of Pliocene age (Meng and Harsh, 1988). This clay unit overlies the Yorktown-Eastover aquifer, except where it is unconfined in the uplands and stream valleys discussed previously (Harsh and Laczniak, 1990). In some of these upland areas, the Yorktown confining unit has been removed by erosion, and the Quaternary sediments of the surficial Columbia aquifer are in direct contact with the Yorktown-Eastover aquifer (Harsh and Laczniak, 1990). Throughout most of its extent, however, this confining unit thickens to the east from its narrow western outcrops to a maximum of about 80 feet near the Atlantic Coast; its maximum thickness in southeastern Virginia is 56 feet, near the city of Virginia Beach (Meng and Harsh, 1988; Hamilton and Larson, 1988).

The sediments of the Yorktown confining unit consist of very fine sandy to silty clays that range in hue from multicolored to blue-gray to gray (Meng and Harsh, 1988; Harsh and Laczniak, 1988). These clays are further described as well-bedded and massive, and they have been observed to contain some shells (Hamilton and Larson, 1988). Unlike most of the underlying tertiary units, the clay layers of this confining unit are described not as a single, areally extensive confining unit, but as a “series of coalescing clay layers” capping the fining-upward sequences of the Yorktown-Eastover aquifer (Meng and Harsh, 1988). The sediments represent a marine transgression, and the inferred depositional setting was a lagoon or bay area on a broad, shallow marine shelf (Meng and Harsh, 1988).

The average vertical hydraulic conductivity value for this unit from calibration of model simulations was 8.64×10^{-4} ft/d, while vertical hydraulic conductivity values as determined from laboratory analysis of core range from 3.9×10^{-3} to 5.9×10^{-4} ft/d (Harsh and Laczniak, 1990). Modeling sensitivity tests and calibrations yielded a specific storage value of 1×10^{-6} for this confining unit and all others in the system (Harsh and Laczniak, 1990). However, the release of water from confining units was not considered in these simulations.

Columbia Aquifer

The Columbia aquifer is the uppermost aquifer in the Virginia Coastal Plain system. Consisting of sandy surficial deposits overlying the Yorktown confining unit, the Columbia aquifer is typically unconfined, though as Meng and Harsh (1988) point out, clayey lenses may produce confined or semi-confined conditions locally. The sediments of the Columbia aquifer are typically Pleistocene to Holocene in age, but some sandy sediments overlying the Yorktown confining unit are also included in this unit (Meng and Harsh, 1988). The thickness of this unit varies significantly with location and topography, but it generally thickens eastward across the Coastal Plain (Meng and Harsh, 1988). More specifically, the saturated thickness ranges from approximately 15 ft near the Fall Line to about 80 ft in southeastern Virginia (Harsh and Laczniak, 1990).

Most of the sediments of the Columbia aquifer were deposited as the result of a series of marine transgressions in the Pleistocene, and they include fluvial, estuarine, marginal-marine, and nearshore-shelf sediments (Powars, 2000). Due to the variable nature of their deposition, these Pleistocene sediments consist of a number of geologic formations that have been identified locally (Meng and Harsh, 1988). These Pleistocene formations are undifferentiated here, but more information about the individual units, including lithology and spatial distribution, is provided by other authors (Mixon and others, 1989). In general, the Pleistocene deposits have been described as "a step-like succession of terraces and intervening scarps" paralleling the coast and major streams and decreasing in age and elevation towards those boundaries (Powars, 2000). The overlying Holocene deposits include estuarine, marsh, swamp, dune, alluvial, and colluvial sediments, according to recent investigations (Powars, 2000). In addition, some channel fill deposits can be found cutting across other units in the Virginia Coastal Plain (Powars, 2000).

The Pleistocene and Holocene formations are characterized by fining-upwards sequences ranging from very coarse gravel to fine sands, silts and clays (Powars, 2000). They are described as gray to oxidized variegated interbedded sands, gravels, silts and clays, including shells and peat in some areas (Powars, 2000).

Previous authors report significant spatial variations in hydrologic properties of the Columbia aquifer, but point out that data are not available to adequately define these variations (Harsh and Laczniak, 1990). One example aquifer test from Chesapeake, Virginia was reported to yield a transmissivity value of 250 ft^2/d, while another author reported a specific yield of 0.15 for this aquifer, calculated from aquifer testing on the Eastern Shore Peninsula (Harsh and Laczniak, 1990). Other available data are sparse, but the properties of the surficial aquifer are thought to be fairly unimportant to a study of compaction and subsidence.

Withdrawals from the Columbia aquifer typically supply small, domestic users (Meng and Harsh, 1988). This aquifer is easily exploited, and withdrawal data are not readily available. However, net withdrawals are thought to be small, as most withdrawals are probably returned as surficial recharge (Harsh and Laczniak, 1990). This aquifer is considered to be an important source of recharge to the underlying confined flow system of the Coastal Plain (Harsh and Laczniak, 1990)

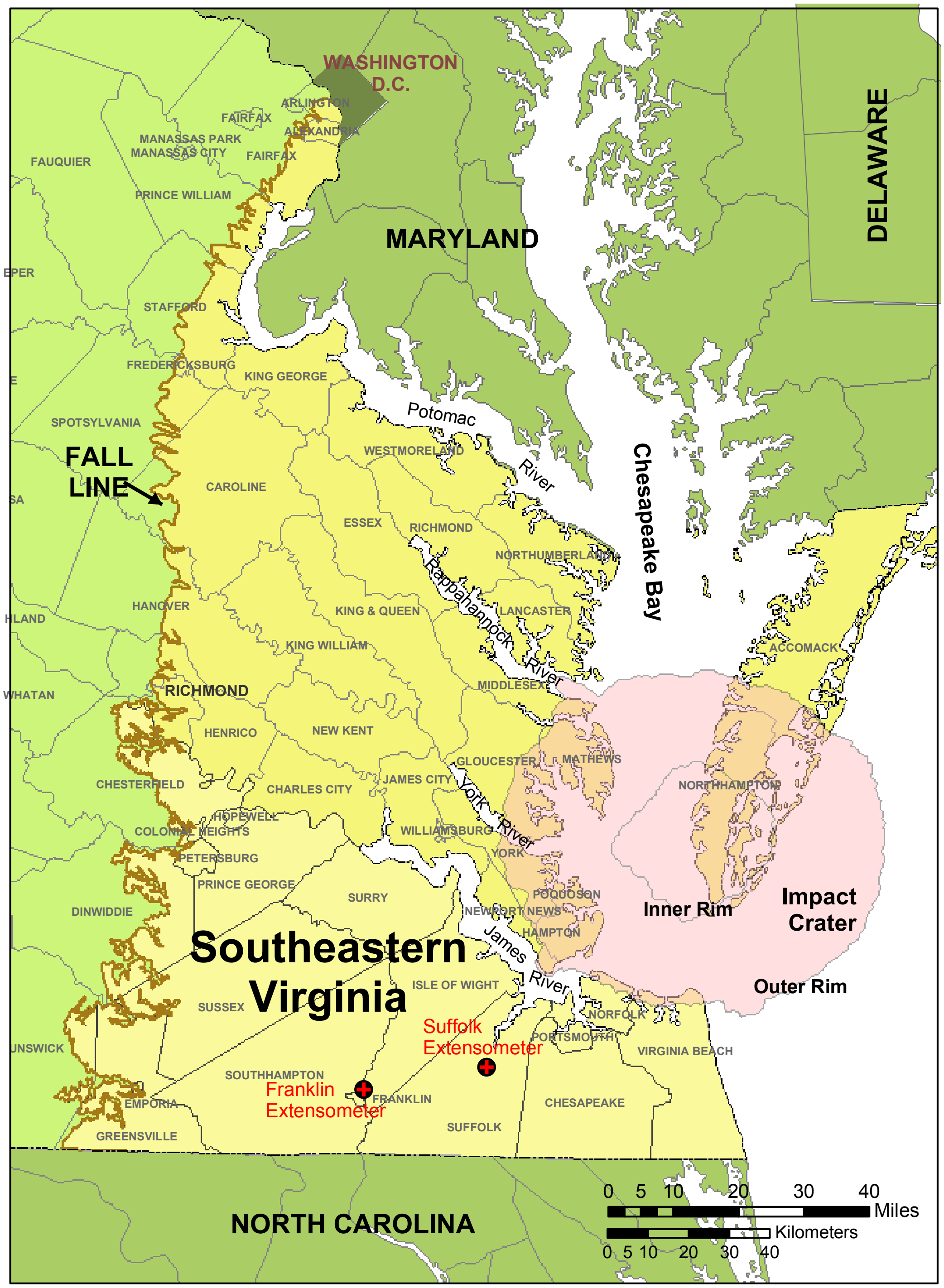


Figure 3-1: Map of the Virginia Coastal Plain and adjacent areas depicting the location of the Chesapeake Bay impact structure and other relevant features.

Period	Epoch		Geologic Units	Hydrogeologic Units
Quaternary	Holocene		Holocene Deposits	Columbia Aquifer
	Pleistocene		Pleistocene Deposits	
Tertiary	Pliocene		Bacons Castle Formation	Yorktown Confining Unit
		Chesapeake Group	Yorktown Formation	Yorktown-Eastover Aquifer
	Miocene		Eastover Formation	St. Marys Confining Unit
			St. Marys Formation	St. Marys-Choptank Aquifer
			Choptank Formation	
			Calvert Formation	Calvert Confining Unit
	Oligocene		Old Church Formation	Piney Point Aquifer
	Eocene	Pamunkey Group	Chickahominy Formation (Crater)	Chickahominy Confining Unit
			Piney Point Formation	Piney Point Aquifer
			Nanjemoy Formation	Nanjemoy-Marlboro Confining Unit
	Paleocene		Marlboro Clay	
			Aquia Formation	Aquia Aquifer
			Brightseat Formation	Brightseat Confining Unit
				Brightseat Aquifer
Cretaceous	Late Cretaceous		Peedee Formation (of North Carolina)	Peedee Confining Unit
				Peedee Aquifer
			Red Beds (Powars, 2000)	Virginia Beach Confining Unit
			Glauconitic Sands (Powars, 2000)	Virginia Beach Aquifer
			Upper Cenomanian Beds (Powars, 2000)	Upper Potomac Confining Unit
			Potomac Formation	Upper Potomac Aquifer
	Early Cretaceous			Middle Potomac Confining Unit
				Middle Potomac Aquifer
				Lower Potomac Confining Unit
				Lower Potomac Aquifer
Jurassic-Triassic	Rift-basin Rocks			
Paleozoic and Proterozoic	Basement Rocks			

= Not present in study area

Figure 3-2: Correlation of hydrogeologic and stratigraphic units in southeastern Virginia. Modified from Meng and Harsh (1988), Hamilton and Larson (1988), and Powars (2000).

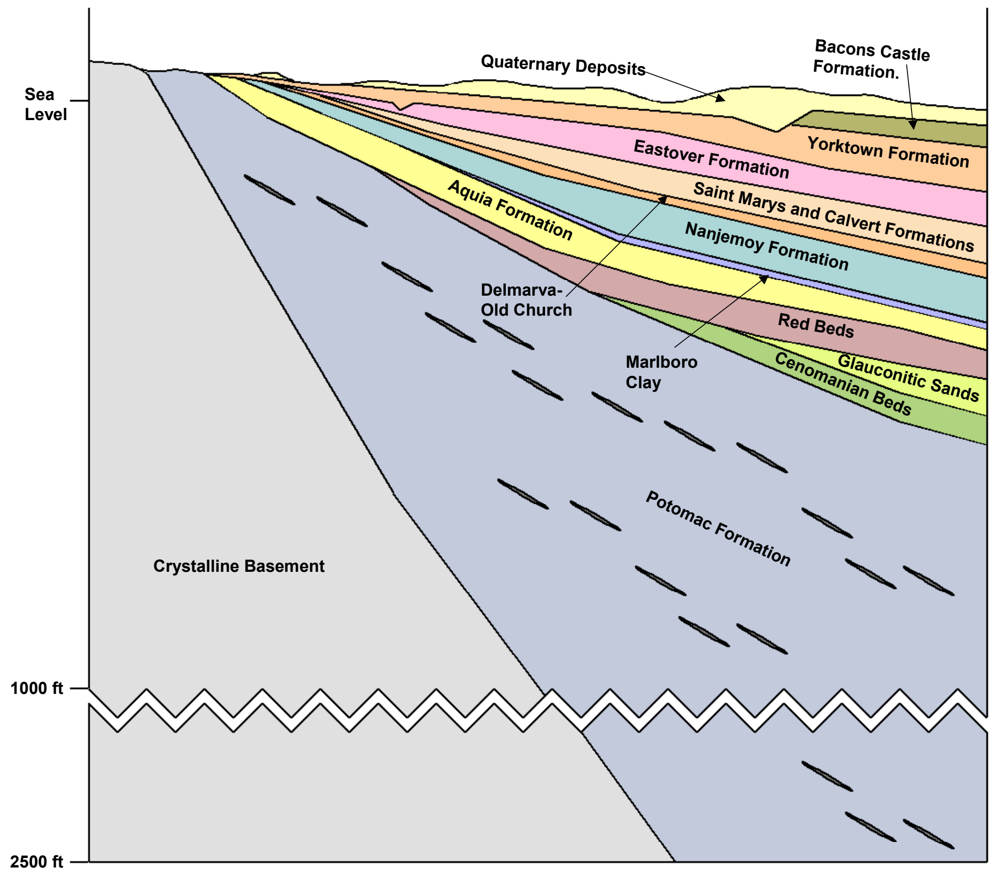


Figure 3-3: Generalized geologic cross section of southeastern Virginia from the Fall Line to the coast. Modified from Powars and Bruce (1999), based on verbal communication with McFarland (2001).

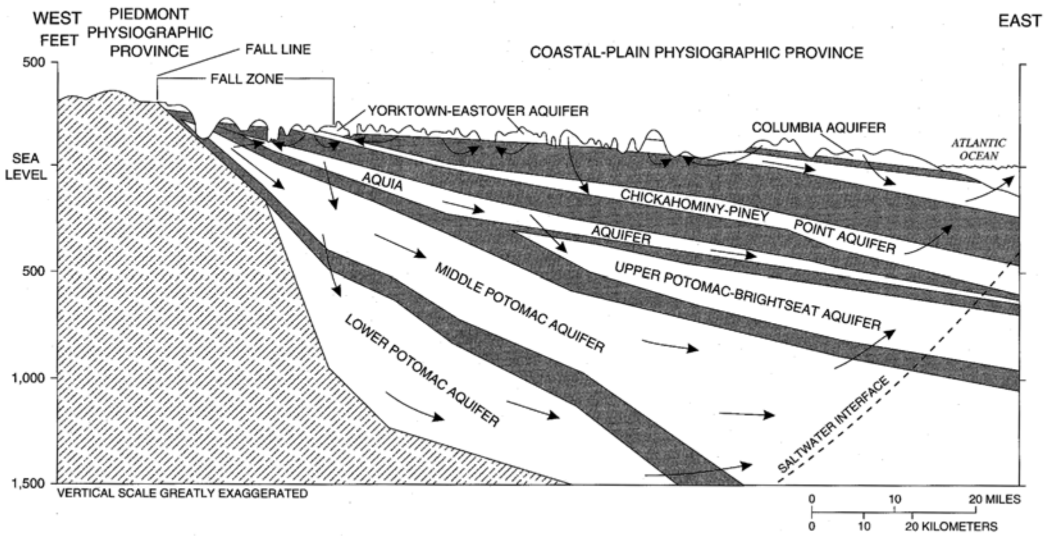


Figure 3-4: Generalized hydrogeologic cross section of southeastern Virginia from the Fall Line to the coast (McFarland, 1998). Arrows show directions of groundwater flow under pre-development conditions.

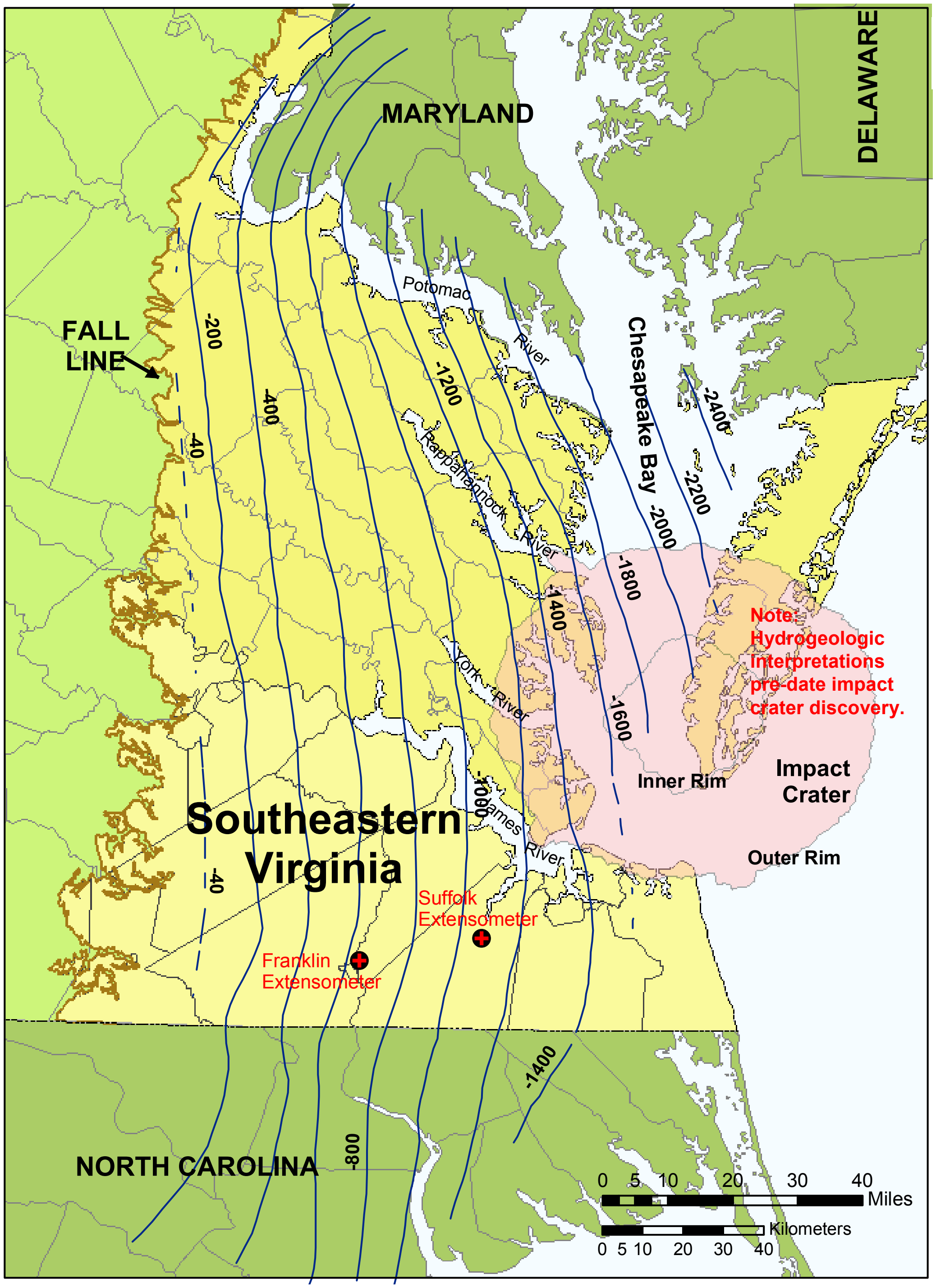


Figure 3-5: Map showing the extent of the Lower Potomac aquifer with elevation of unit top in feet. Modified from Meng and Harsh (2000), Hamilton and Larson (1988), and Focazio and Samsel (1993).

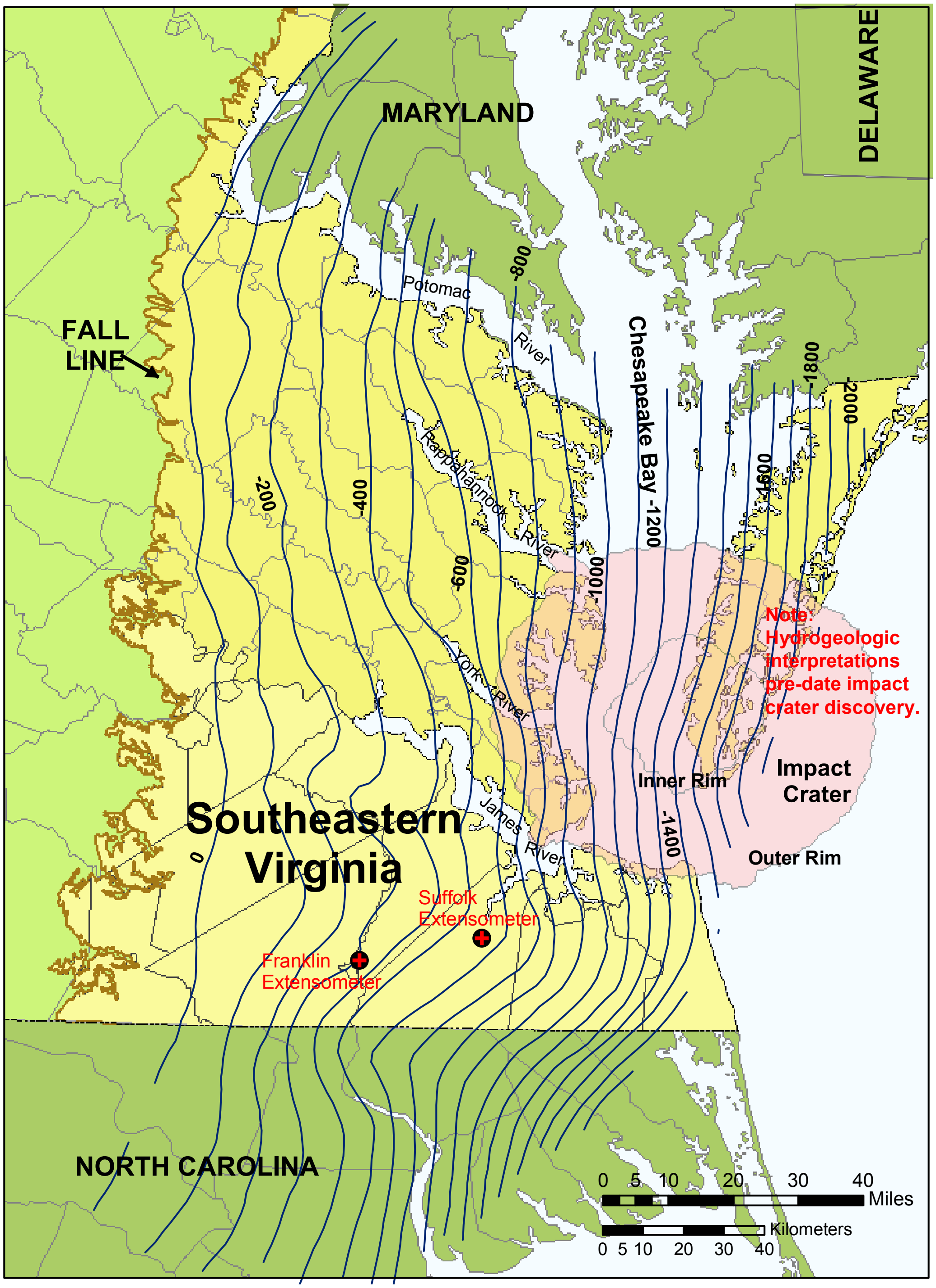


Figure 3-6: Map showing the extent of the Middle Potomac aquifer with elevation of unit top in feet. Modified from Meng and Harsh (2000), Hamilton and Larson (1988), and Focazio and Samsel (1993).

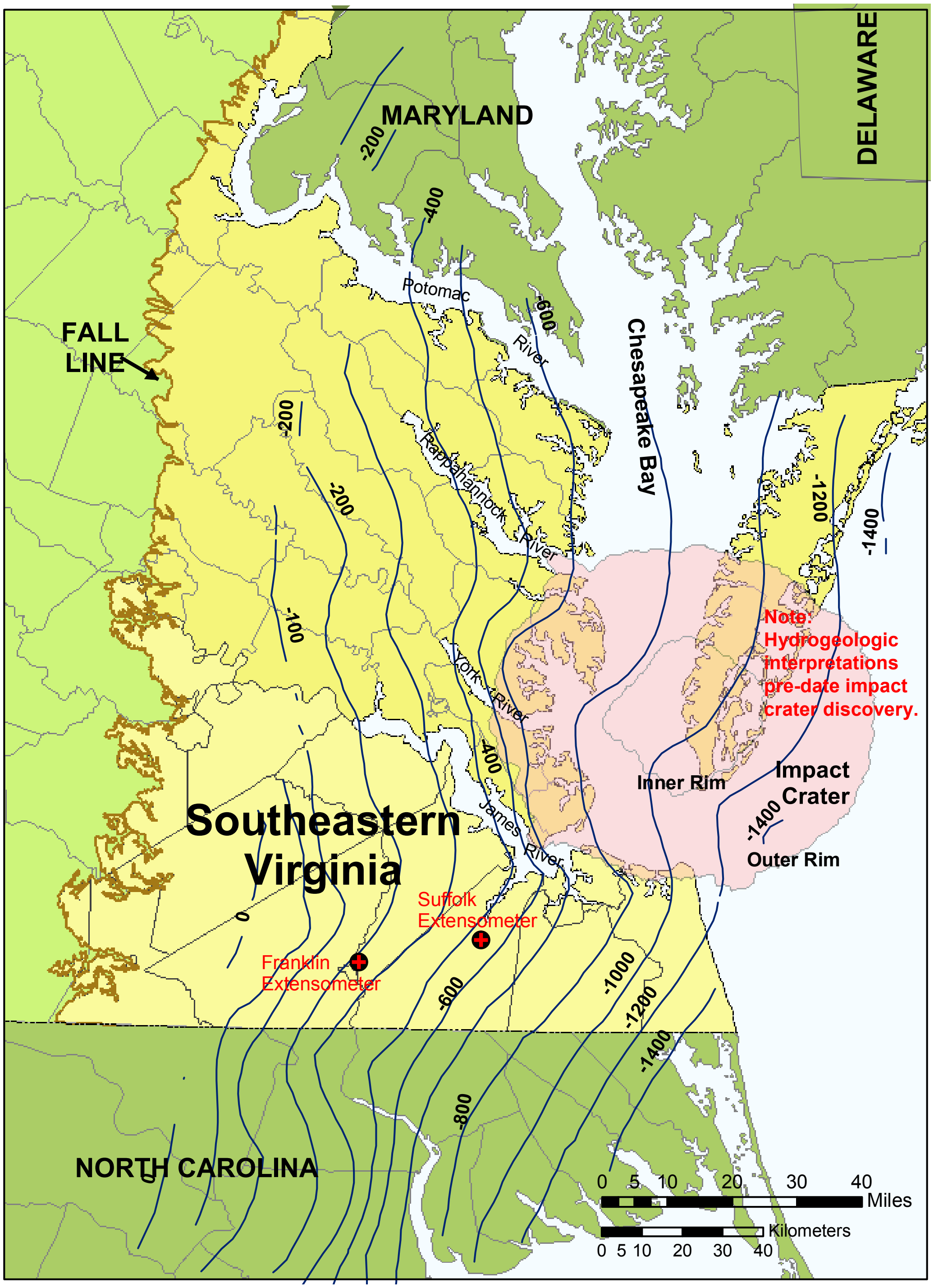


Figure 3-7: Map showing the extent of the Upper Potomac aquifer with elevation of unit top in feet. Modified from Meng and Harsh (2000), Hamilton and Larson (1988), and Focazio and Samsel (1993).

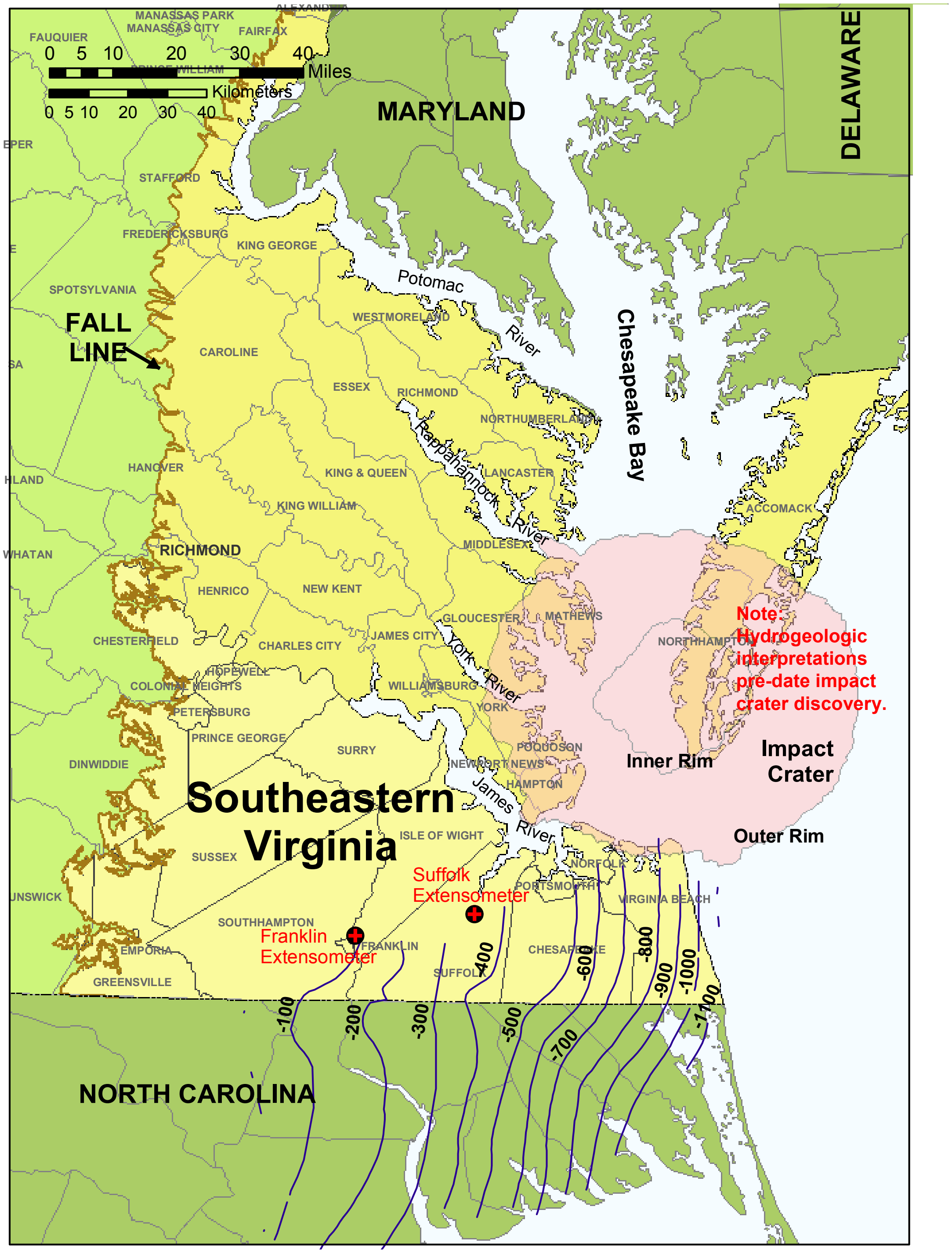


Figure 3-8: Map showing the extent of the Virginia Beach aquifer with elevation of unit top in feet. Modified from Meng and Harsh (2000), Hamilton and Larson (1988), and Focazio and Samsel (1993).

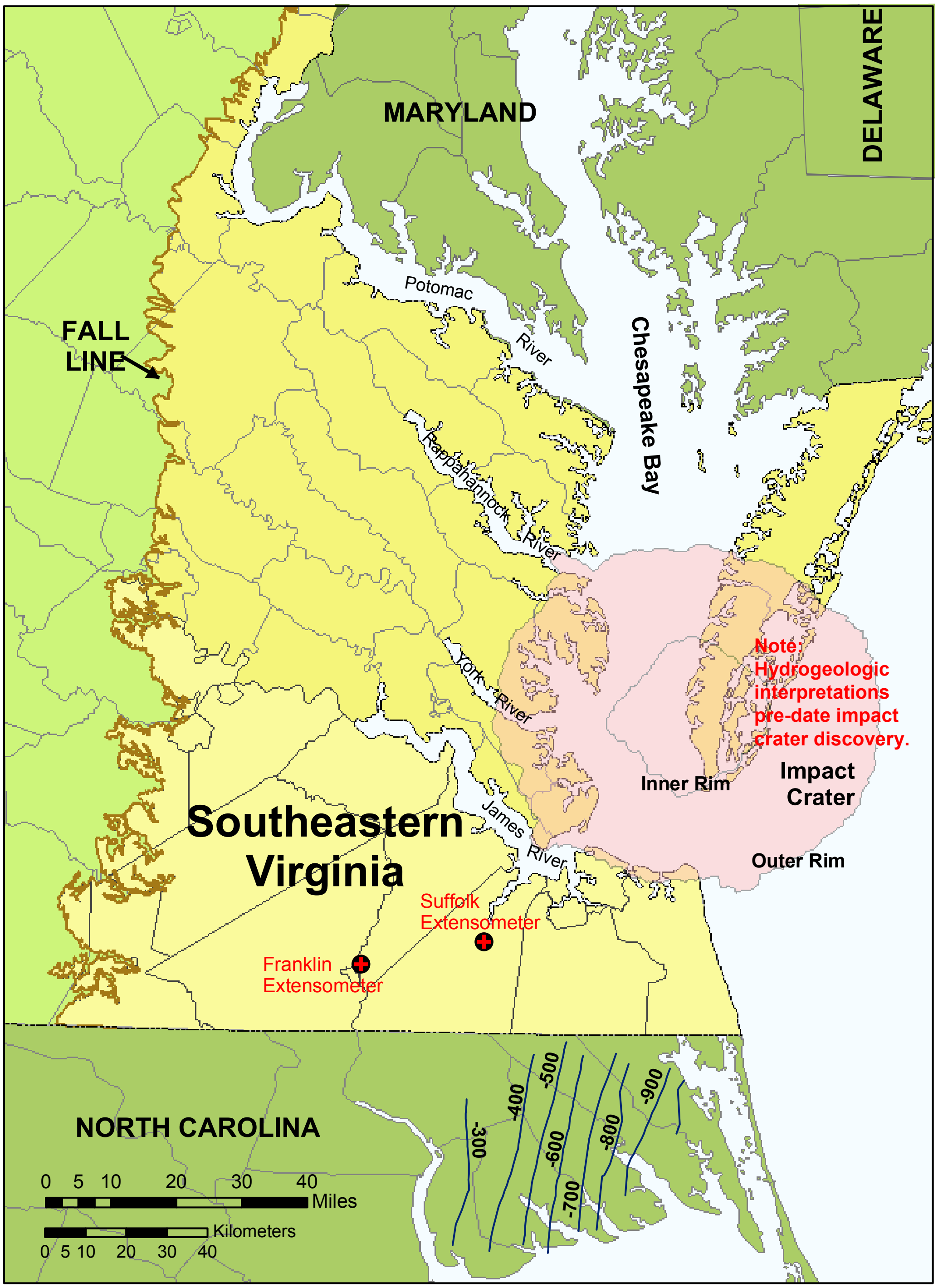


Figure 3-9: Map showing the extent of the Peedee aquifer with elevation of unit top in feet. Modified from Meng and Harsh (2000), Hamilton and Larson (1988), and Focazio and Samsel (1993).

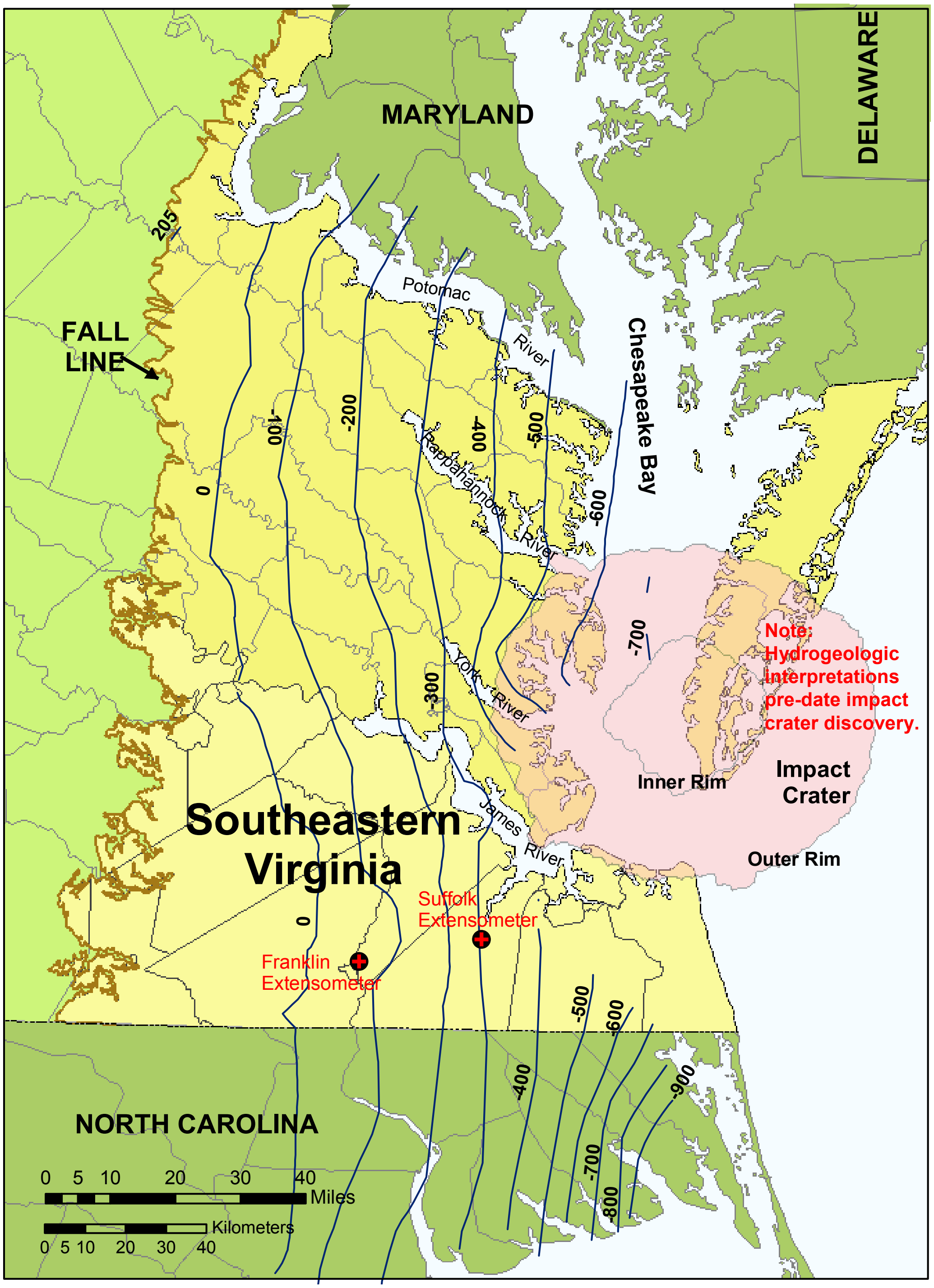


Figure 3-10: Map showing the extent of the Aquia aquifer with elevation of unit top in feet. Modified from Meng and Harsh (2000), Hamilton and Larson (1988), and Focazio and Samsel (1993).

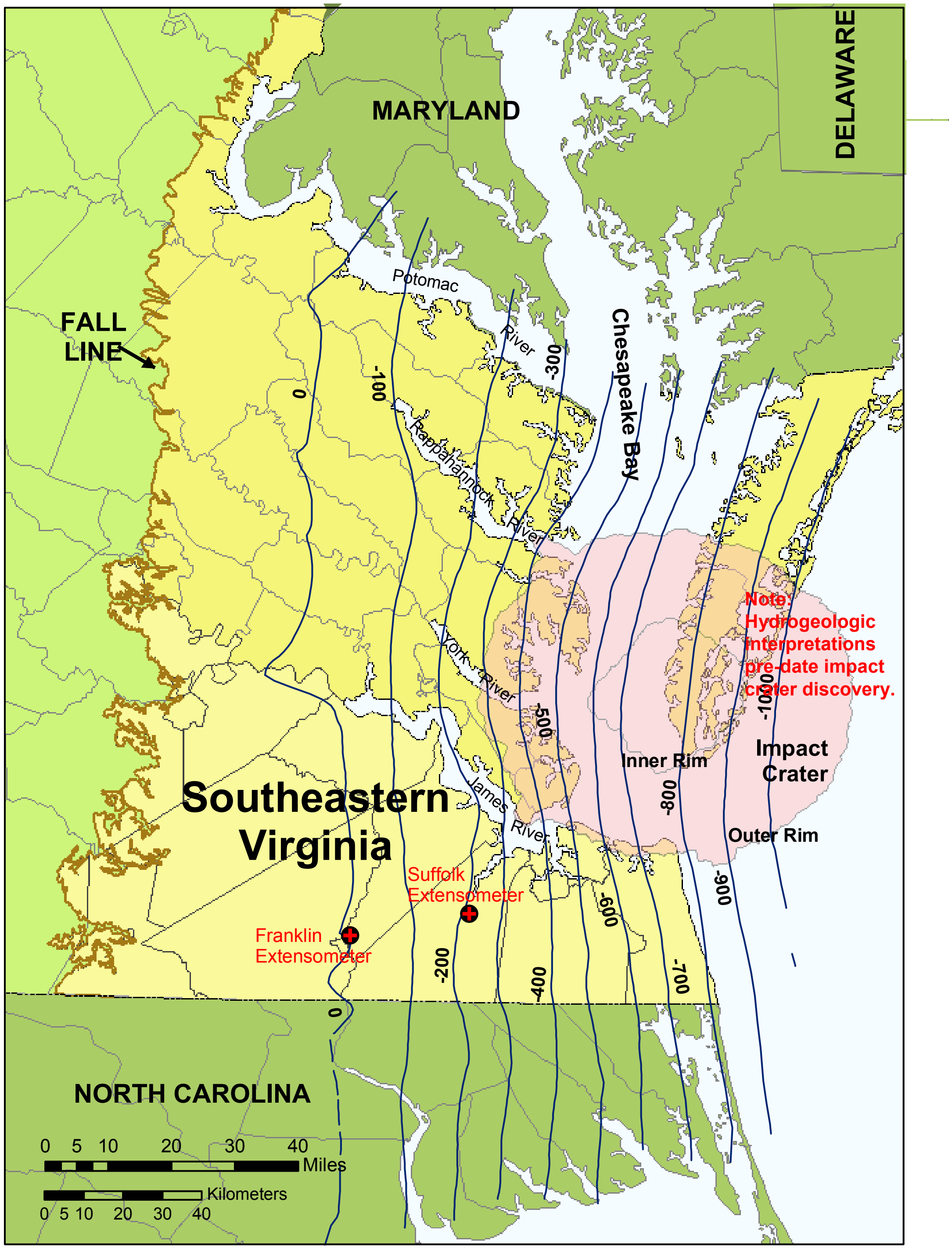


Figure 3-11: Map showing the extent of the Chickahominy-Piney Point aquifer with elevation of unit top in feet. Modified from Meng and Harsh (2000), Hamilton and Larson (1988), and Focazio and Samsel (1993).

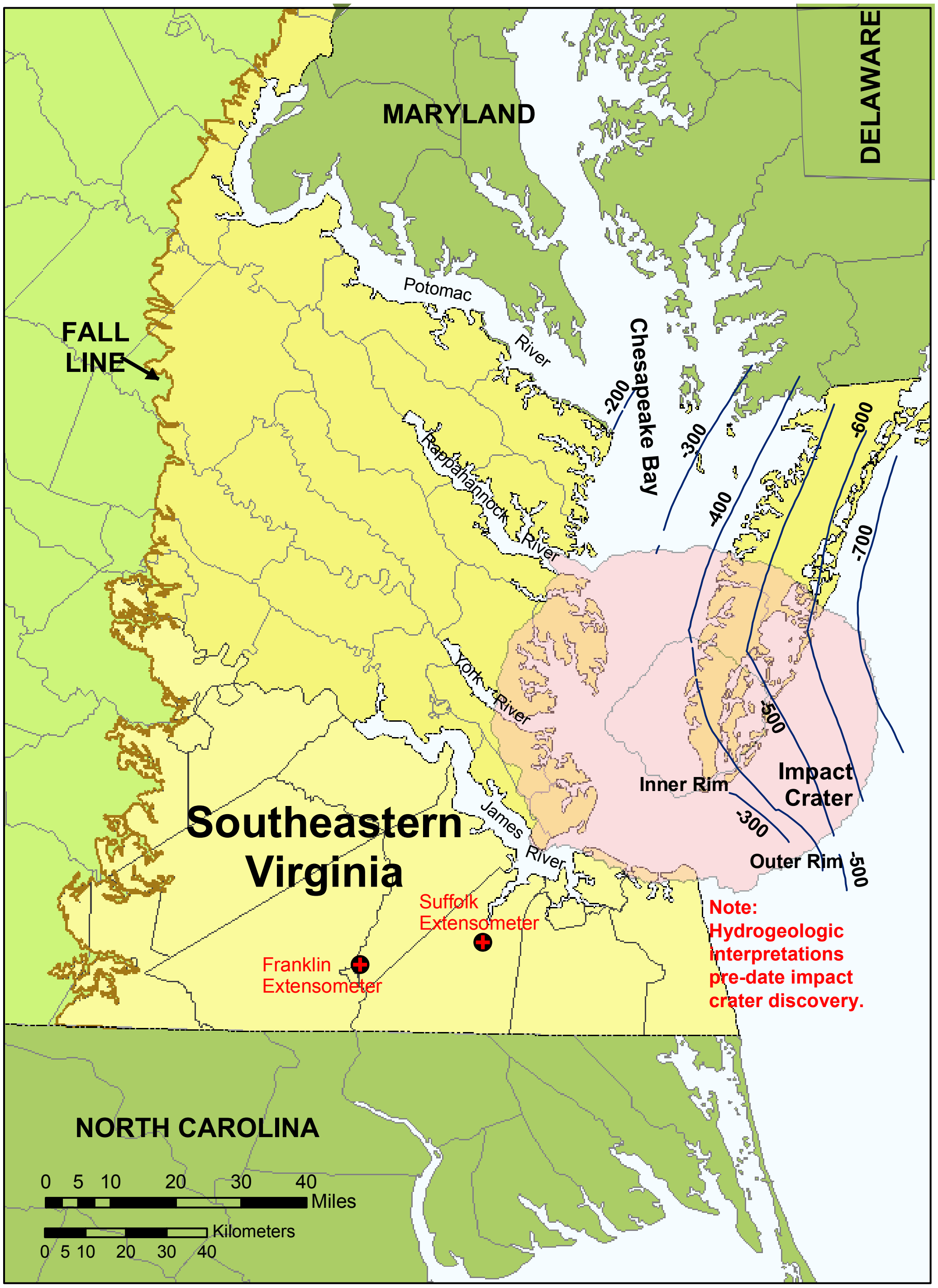


Figure 3-12: Map showing the extent of the Saint Marys aquifer with elevation of unit top in feet. Modified from Meng and Harsh (2000), Hamilton and Larson (1988), and Focazio and Samsel (1993).

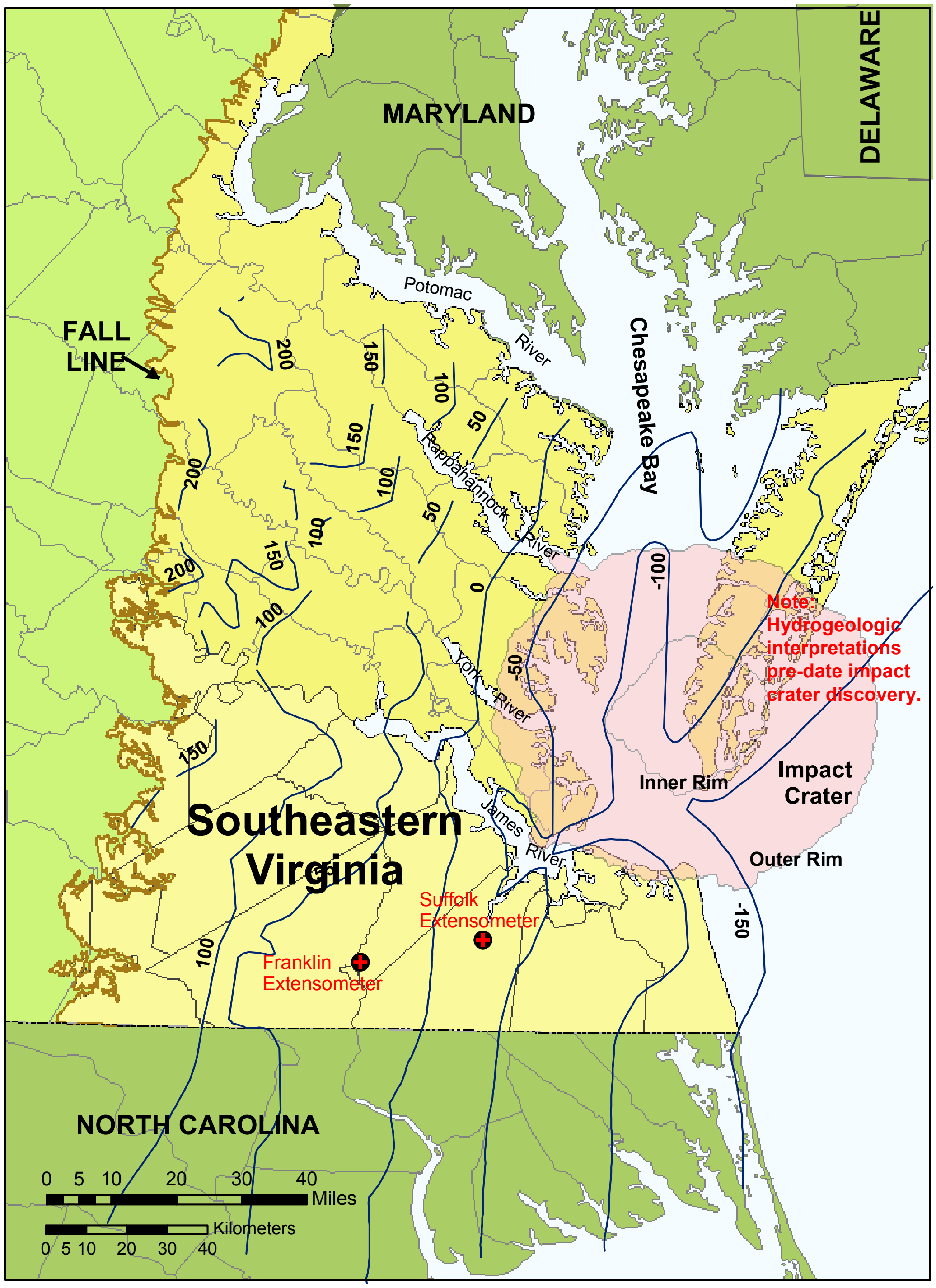


Figure 3-13: Map showing the extent of the Yorktown-Eastover aquifer with elevation of unit top in feet. Modified from Meng and Harsh (2000), Hamilton and Larson (1988), and Focazio and Samsel (1993).

IV: MEASUREMENTS AND ANALYSIS OF HYDROGEOLOGIC DATA FOR THE COASTAL PLAIN OF SOUTHEASTERN VIRGINIA

The analysis of land subsidence due to groundwater withdrawals requires the collection, analysis, and synthesis of data from a variety of sources. Aquifer-system compaction measured with extensometers is most meaningful in light of the water-level changes that induced the compaction. Because these water-level changes are the result of withdrawals from the various aquifers, measurement of withdrawals is necessary for an understanding of the relation between pumpage and hydraulic heads.

A full analysis of aquifer-system compaction and land subsidence for the Coastal Plain of Virginia would require an extensive database of withdrawals and water levels, as well as a three-dimensional model. This approach is well beyond the scope of this research, but a large amount of hydrogeologic data were necessary for this project simply to understand the compaction behavior of this complex aquifer system in the vicinity of the extensometers. This section of the document presents the data collected for this report in the Franklin and Suffolk areas, shown in detail on the maps of figures 4-1 and 4-2. Included here are groundwater withdrawal data, water level data, borehole data used for hydrogeologic interpretations. In addition, the extensometer data are presented and analyzed.

Withdrawal Data

The accurate measurement and of withdrawals from the Coastal Plain aquifers in Virginia is a very difficult undertaking. Reporting requirements vary by location across this region, and by withdrawal category. In general, large industrial and municipal withdrawals are well known, especially in regions designated by the Commonwealth of Virginia as groundwater management areas, but agricultural withdrawals have not been quantified. Even less is known about domestic withdrawals from private wells. The result of these disparities is a fairly large error in the quantification of withdrawals. Fortunately, withdrawal information was not needed for in this study for simulation of one-dimensional compaction, which relates aquifer-system compaction to changes in hydraulic head. Even so, withdrawals were briefly considered for completeness and for the purposes of comparison with measured compaction data.

Historically, groundwater withdrawals from the confined aquifers of the Virginia Coastal Plain (figure 4-3) totaled less than 10 million gallons per day at the start of the twentieth century and increased at a moderate rate prior to the World War II era (Kull and Laczniak, 1987). Significant growth in population and industry in the Hampton Roads area at that time was accompanied by increasing rates of withdrawal from many of the confined aquifers. Withdrawal rates experienced increased tremendously in the 1950s and 1960s, due primarily to large industrial and municipal withdrawals at a few large pumping centers. In the 1970s, the increase in withdrawals began to level off as efforts were begun to limit new large wells in this region. A regulatory approach adopted during that time continues to require proposed new wells in designated groundwater management areas to be permitted based on their estimated impact on the flow system. The permitting process, as well as new public water supply projects importing surface water to the Hampton Roads area, helped to maintain relatively constant withdrawal rates through the 1970s and 1980s. In the 1990s, however, growing demand for additional municipal supplies led to the installation of new wells withdrawing brackish water from the Potomac aquifers in the Hampton Roads area. These new withdrawals post-date the publication shown in figure 4-3, but they represent noticeable increases in the withdrawal rates and have contributed to additional declines in hydraulic head in southeastern Virginia.

The unusual distribution of withdrawals in the Coastal Plain of Virginia is perhaps more interesting than their magnitudes, particularly with regard to the land subsidence problem. Withdrawals from large industrial and municipal pumping centers dominate the Coastal Plain system, in many cases even reversing the directions of regional groundwater flow. As a result, total withdrawals are not evenly distributed across the region. The very skewed distribution of withdrawal rates is depicted qualitatively in figure 4-4, which shows a raised-relief map of withdrawals by county. In this figure, the large withdrawal rates from pulp and paper mills at Franklin and West Point stand out very clearly from the withdrawals in all other areas. The 35 to 40 million gallons per day (mgd) at Franklin and 14 to 18 mgd at West Point together have comprised almost half of the rate of measured withdrawals for the entire Coastal Plain of Virginia in recent years (Harsh and Laczniak, 1987).

The consideration of all Coastal Plain withdrawals was well beyond the scope of the present work, but withdrawals were examined on a more localized basis for the Franklin area. A graph of the withdrawals near Franklin in recent years (figure 4-5) reveals that pumpage magnitudes from the Middle and Lower Potomac aquifers differ somewhat, but the two records exhibit similar patterns. It was initially expected that these withdrawals would vary seasonally, but no seasonal pattern is evident in this graph. Instead, the industrial component overwhelms any seasonal municipal or industrial component of the withdrawal record. Furthermore, withdrawal data were available only from monthly reports, so the limited precision of this data prohibited any quantitative comparison with the continuous compaction record. Instead, analysis was focused on the effects of the withdrawals from these aquifers - their changing water levels - which were generally available at better resolution and quantified with much greater confidence.

Hydraulic Head Data

A thorough analysis of the water-level decline in the vicinities of Franklin and Suffolk was a key component of this study. Interpretations of geologic and geophysical data indicated that seven aquifers are present at each location, suggesting that a large amount of hydraulic head data might be necessary to understand the vertical head distribution in this system. An ideal data set, in fact, would have included water-level records for each of the seven aquifers at each of the two locations from pre-development conditions through the present. The availability of actual data was less than ideal, but an effort was made to identify and compile all of the water-level records collected near the two sites over the history of measurement. A search of the USGS Ground Water Site Inventory (GWSI) database yielded six records for the Franklin area and eight records for the Suffolk area. These records are presented in figures 4-6 and 4-7.

The Franklin water levels presented in figure 4-6 include the following wells: (a) 55B62, screened in the Yorktown-Eastover aquifer; (b) 55B67, screened in the Piney Point aquifer; (c) 55B16, screened in the Middle Potomac aquifer; (d) 55B22, screened in the Middle Potomac aquifer; (e) 55B35, screened in the Middle Potomac aquifer; and (f) 55B36, screened in the Lower Potomac aquifer. The location of these wells is presented in figure 4-1, and basic data on well dimensions are presented in table 4-1. Each of the water level records is presented with a graph scaled to maximize visibility, so it should be noted that the time (x axis) and hydraulic head (y axis) scales do not match among the 6 graphs. There was no common scale on which these records could reasonably be presented.

Table 4-1: Wells near the Franklin extensometer

Well ID	Well Elevation (ft above mean sea level)	Screen Interval (ft below land surface)	Aquifer Designation	Lateral Distance from Extensometer (ft)
55B62	27	25 – 30	Yorktown-Eastover	0
55B67	34	130 – 140	Piney Point	8,484
55B16	25	285 – 305	Middle Potomac	218
55B22	21	335 – 354	Middle Potomac	3,262
55B35	32	430 – 623	Middle Potomac	1,527
55B36	37	720 – 860	Lower Potomac	2,430

Two problems with these records are immediately obvious: (1) none of the records span the complete history of water-level decline, and most do not span the same time periods, though there is considerable overlap; and (2) few of the wells are located directly adjacent to the extensometer, though most are located within about a mile of the instrument. These are important issues because it was important to understand the history of water-level decline across the entire period of groundwater development (essentially, the twentieth century), and these records were collected for comparison with measured aquifer-system compaction data from the Franklin extensometer. For the later purpose, close proximity to the extensometer is most desirable. Of course, not all of the aquifers present at this location are represented in these water-level records, which is another constraint on the understanding of the system. The most complete story of water-level changes at Franklin is available from well 55B22, which is located in close proximity to the Franklin extensometer and was measured from approximately 1940 to 1990. This record was very important to the analysis, but its utility was somewhat limited because it did not extend through the last decade of the century.

The Suffolk water levels presented in figure 4-7 include the following wells: (a) 58B268, screened in the unconfined Columbia aquifer; (b) 58B269, screened in the Piney Point aquifer; (c) 58B270, screened in the Upper Potomac aquifer; (d) 58B271, screened in the Upper Potomac aquifer; (e) 58C56, screened in the upper part of the Middle Potomac aquifer; (f) 58C53, screened in the Middle Potomac aquifer; (g) 58B235, screened in the Middle Potomac aquifer; and (h) 58B273, screened in the Middle Potomac aquifer. These wells can be located in figure 4-2, and basic data on their dimensions are given in table 4-2.

Table 4-2: Wells near the Suffolk extensometer

Well ID	Elevation (ft)	Screen Interval (ft)	Aquifer Designation	Distance from Extensometer (ft)
58B268	35	12 – 20	Columbia	16,124
58B269	35	350 – 360	Piney Point	16,124
58B270	35	490 – 500	Upper Potomac	16,124
58B271	29	501 – 710	Upper Potomac	16,124
58C56	10	557 – 567	upper Middle Potomac	0
58C53	10	881 – 896	Middle Potomac	0
58B235	53	530 – 562	Middle Potomac	10,389
58B273	26	541 – 640	Middle Potomac	15,108

The Suffolk water-level records are much less useful than the Franklin records, because they are generally located much farther from the extensometer installation and generally span much shorter time periods. In particular, none of the water-records available for Suffolk show head decline across the early to middle parts of the twentieth century, though a few early data points for the Middle Potomac aquifer are available from 58B235 (figure 4-7g). Consequently, these water-level records by themselves provided little constraint on historical declines in water levels at Suffolk.

In order to address the problems with limited time periods for the water level records at both locations, early reports on groundwater conditions in Virginia were examined in detail. Sanford (1913) and Cederstrom (1945) provided both early water-level measurements at nearby locations and qualitative reports of approximate groundwater levels in the early part of the century. These reports were used to supplement the available records to generate a comprehensive understanding of the history of head decline at the two locations. Data from these reports were ultimately used in the generation of synthetic water-level records for input files for the compaction model.

From the inspection of water-level records at both sites, some important general observations can be made. First, declines in water levels have been most dramatic in the Potomac aquifers, and particularly in the Middle Potomac aquifer, which has experienced a head decline of over 60 meters from pre-development conditions (figure 4-6d). The largest head declines in the Potomac aquifers are apparent at the Franklin location, where the lowest hydraulic heads in the Coastal Plain system have historically been measured (Cosner, 1975). On the other hand, heads in the Potomac aquifers at Franklin have generally leveled off in recent years, with some minor fluctuations.

It is clear that water-level declines in the Potomac aquifers have been less severe at Suffolk, but the available records indicate that water levels are still declining significantly at Suffolk. It is unclear whether these declines result from the expansion of the regional cone of depression outward from the Franklin pumping center or from the increase in local pumping rates, but it is possible that both factors are responsible. An examination of withdrawal data and water-level records revealed that these drawdowns appear to represent the superposition of local and regional effects.

There is a strong apparent correlation between the water levels in the various aquifers at each site, particularly for the Potomac aquifers of Cretaceous age. It is not known whether this correlation is controlled by concurrent pumping from all of the Potomac aquifers or from leakage through the confining units, because many large wells are screened across two or three of the Potomac aquifers. However, the apparent correlation between the Potomac water levels and the water levels of the Piney Point aquifer (which is not pumped significantly at these locations) indicates that significant downward leakage is occurring through the system. Comparison of hydraulic heads indicates that flow at both sites likely has a strong vertical component (both downward and upward) toward the Middle Potomac aquifer, likely driven by pumping.

Perhaps the most significant observation resulting from the water-level analysis is the apparent effect of the confining unit(s) above the Potomac system, which has maintained a very large head gradient between the Piney Point aquifer above and the Upper Potomac aquifer below. At Franklin, this gradient is especially large, with almost 50 meters of head differential between two wells, separated by less than 60 vertical meters between the two aquifers. It is also notable that the Yorktown-Eastover aquifer and the unconfined Columbia aquifer show no significant effects from pumping, and appear to be totally isolated from the aquifers below.

In addition to analyzing water levels in the vicinity of Franklin and Suffolk, a larger effort was undertaken to understand the current head conditions in the Potomac system across the Coastal Plain of Virginia by generating a current map of the piezometric surface for the Middle

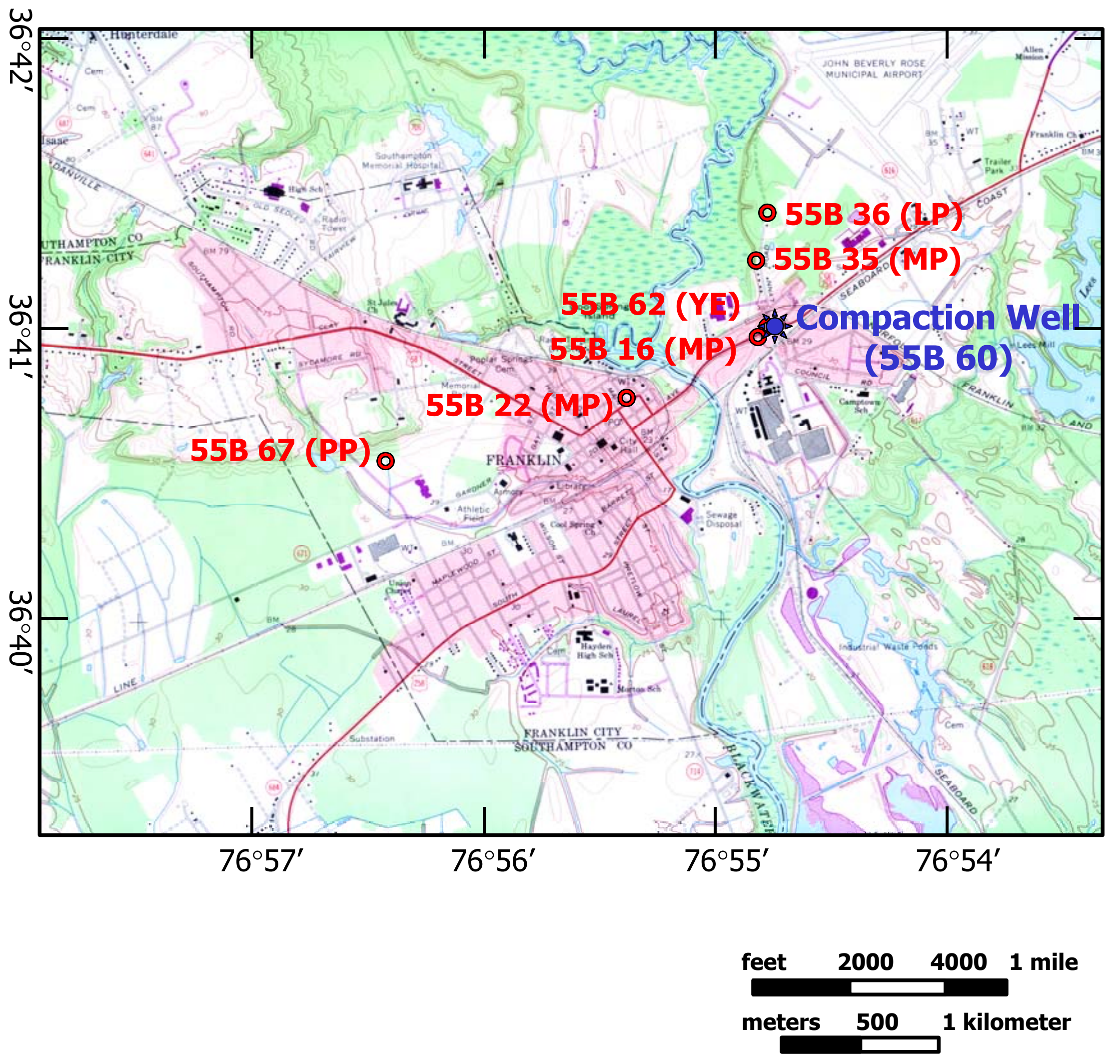


Figure 4-1: Detailed map of the Franklin, Virginia area. Wells are designated by USGS local identification numbers, with aquifer designations in parentheses. (Base map from the USGS 7.5' Quadrangle.

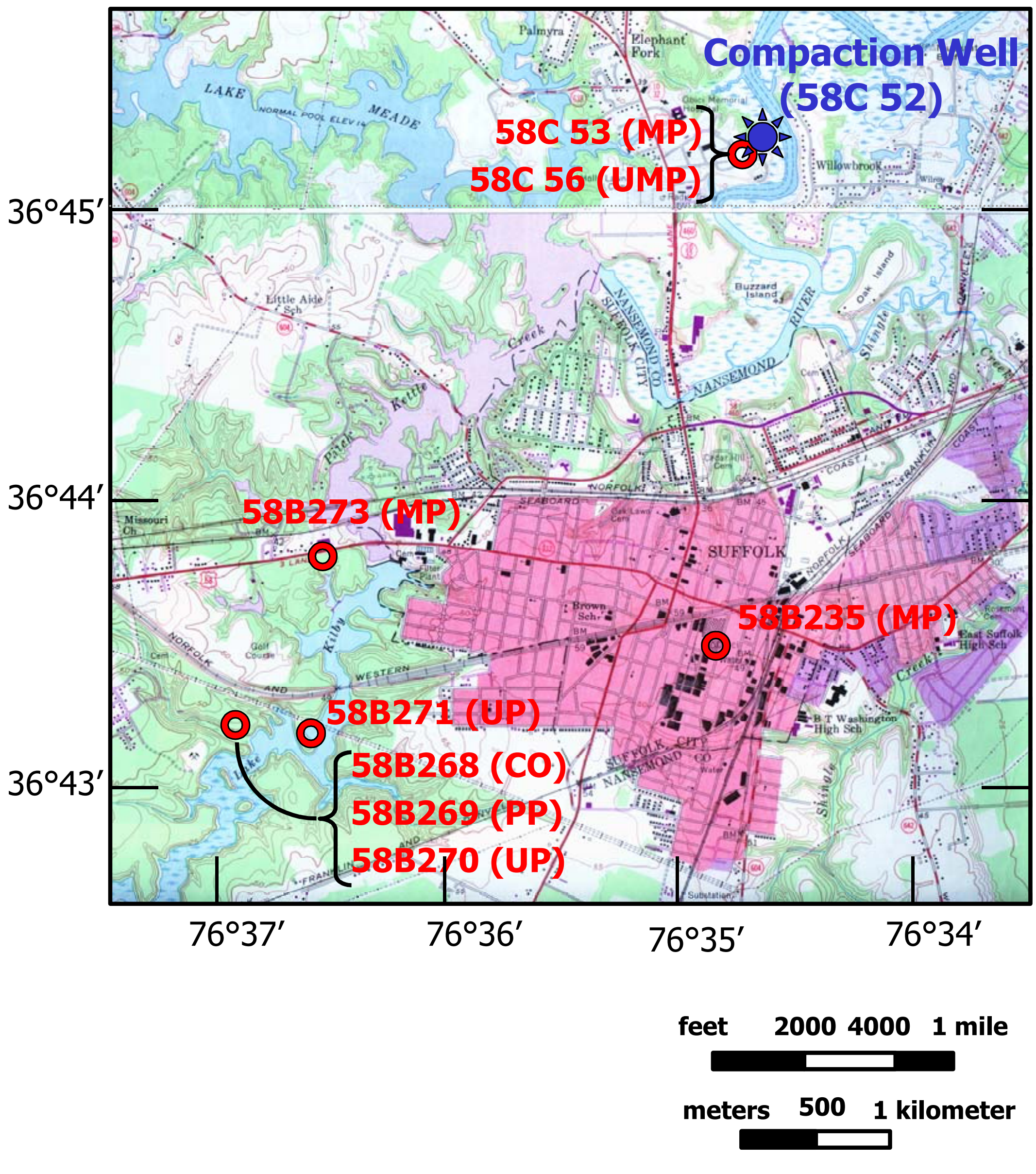


Figure 4-2: Detailed map of the Suffolk, Virginia area. Wells are designated by USGS local identification numbers, with aquifer designations in parentheses. (Base map from the USGS 7.5' Quadrangle.

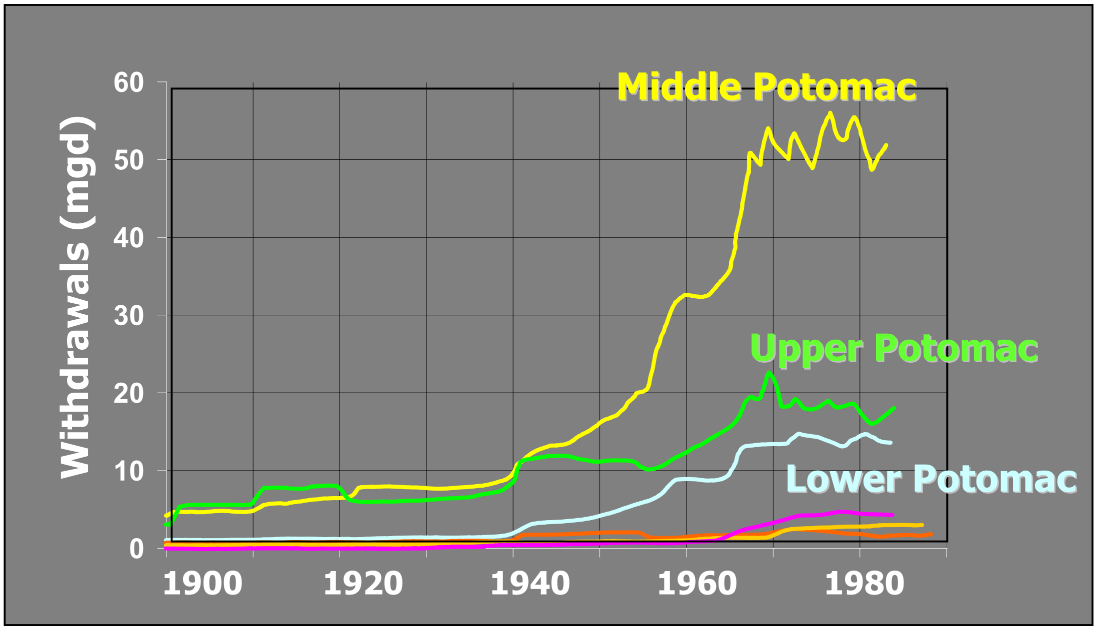


Figure 4-3: Historical record of yearly withdrawals (in millions of gallons per day) from the confined aquifers of the Virginia Costal Plain.

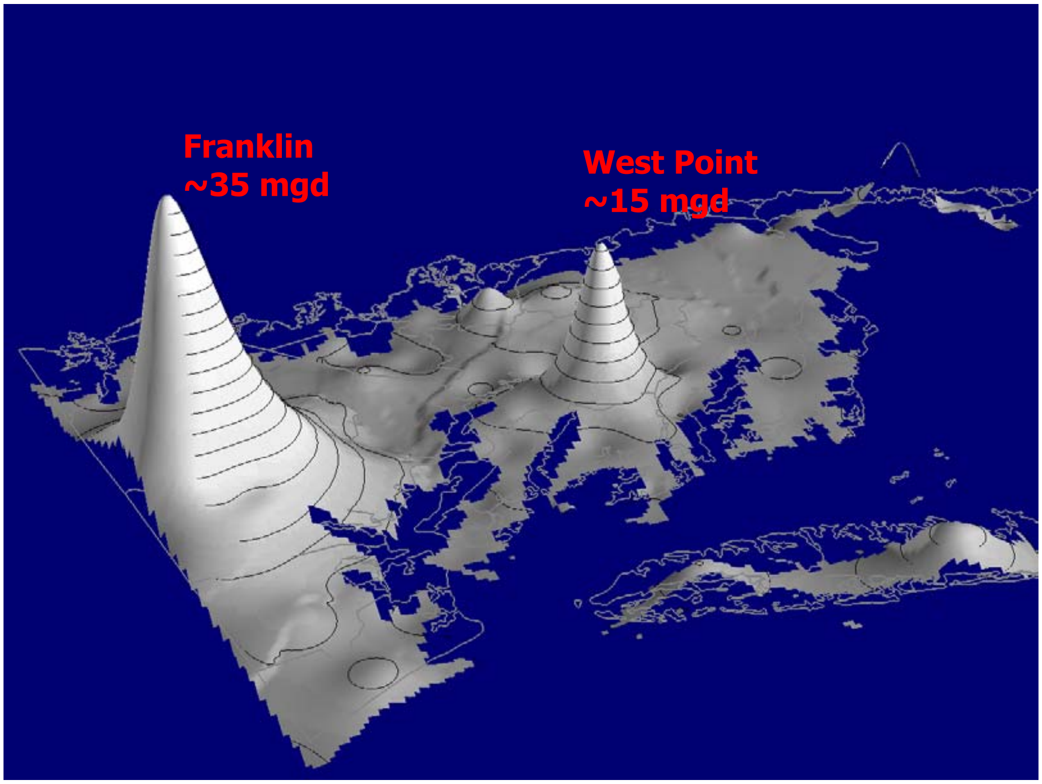


Figure 4-4: Diagram showing the magnitude and distribution of groundwater withdrawals from the confined aquifers in the Virginia Coastal Plain. Contours are in increments of million gallons per day (mgd). Large peaks are due to industrial withdrawals at Franklin to the south (left) and West Point to the north (right).

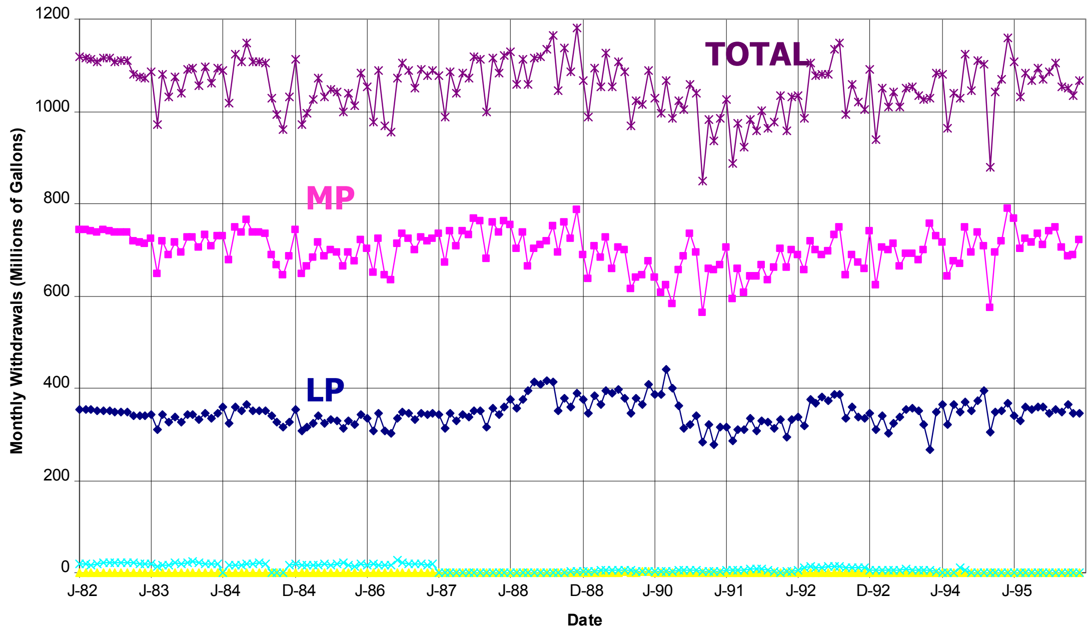


Figure 4-5: Monthly groundwater withdrawals (in millions of gallons) at Franklin from 1982 through 1995 for the Lower Potomac (LP), Middle Potomac (MP), Upper Potomac (UP) and Aquia (AQ) aquifers.

Fig. 4-6a

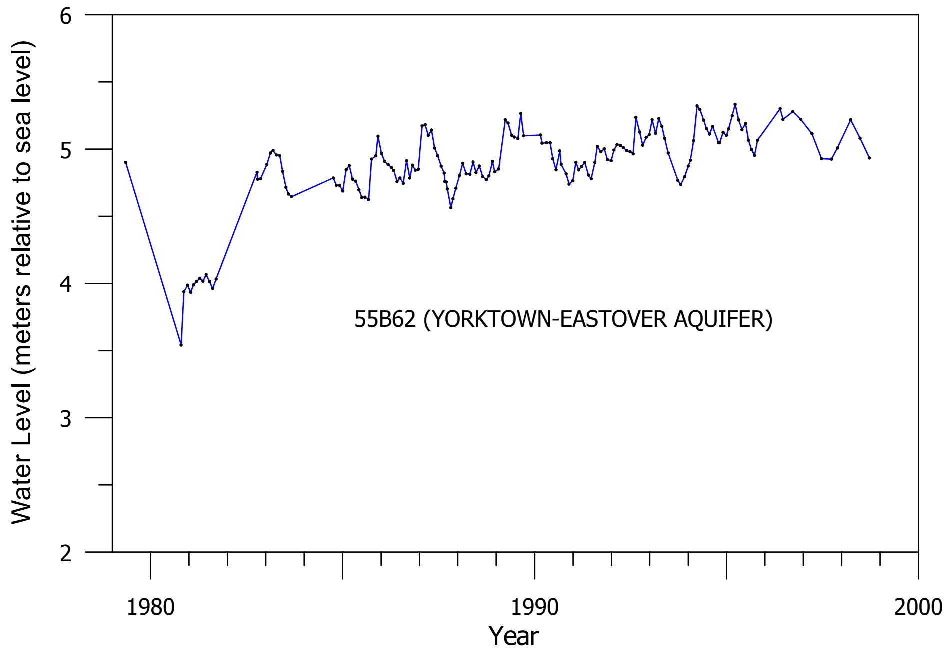


Fig. 4-6b

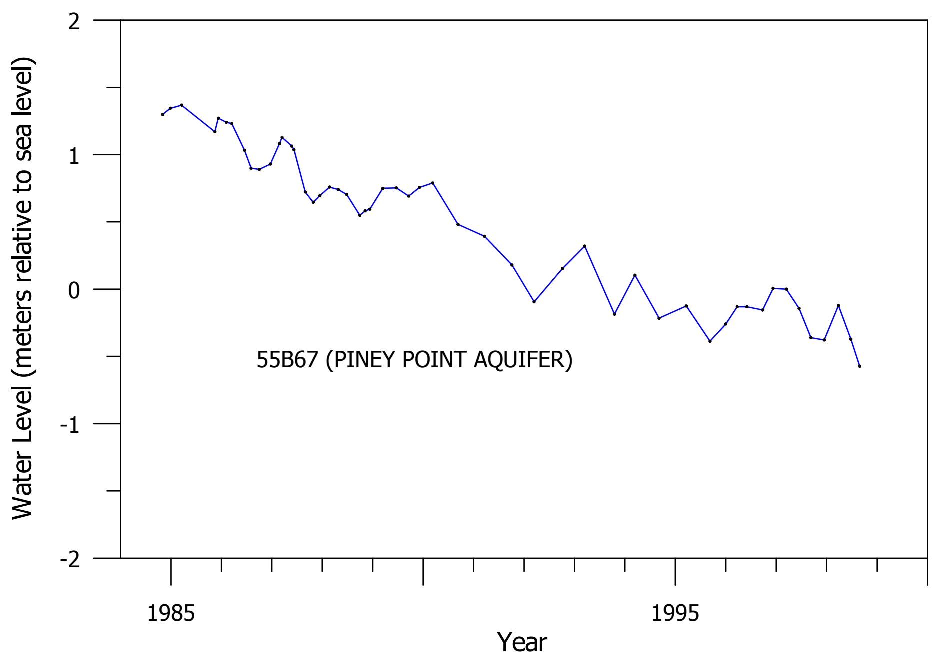


Figure 4-6 a-b: Water levels from available monitoring wells in the Franklin, Virginia area, in meters relative to the sea-level datum. Records are designated by USGS local identification numbers.

Fig. 4-6c

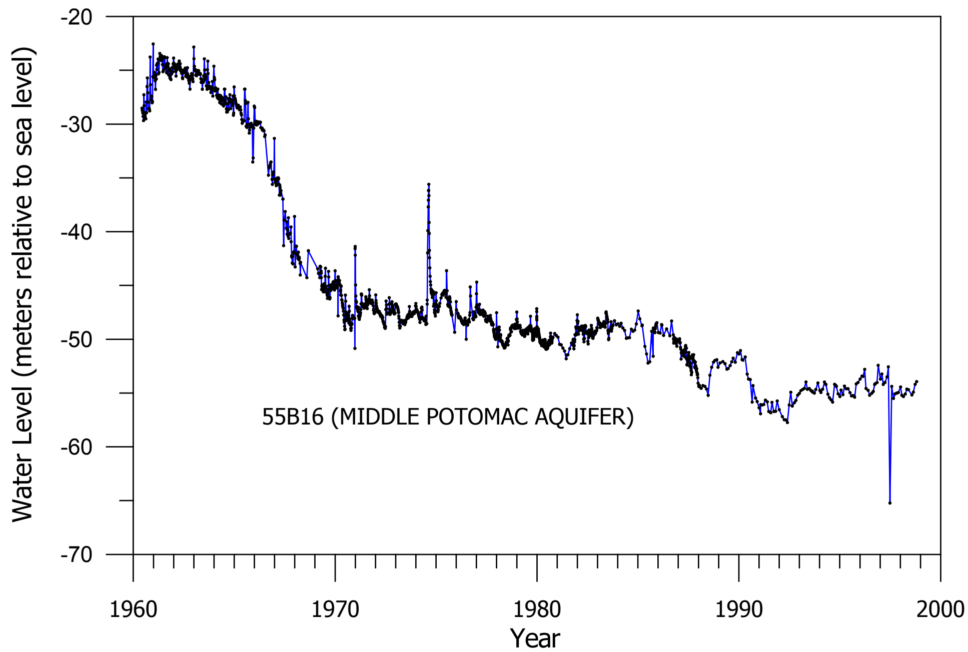


Fig. 4-6d

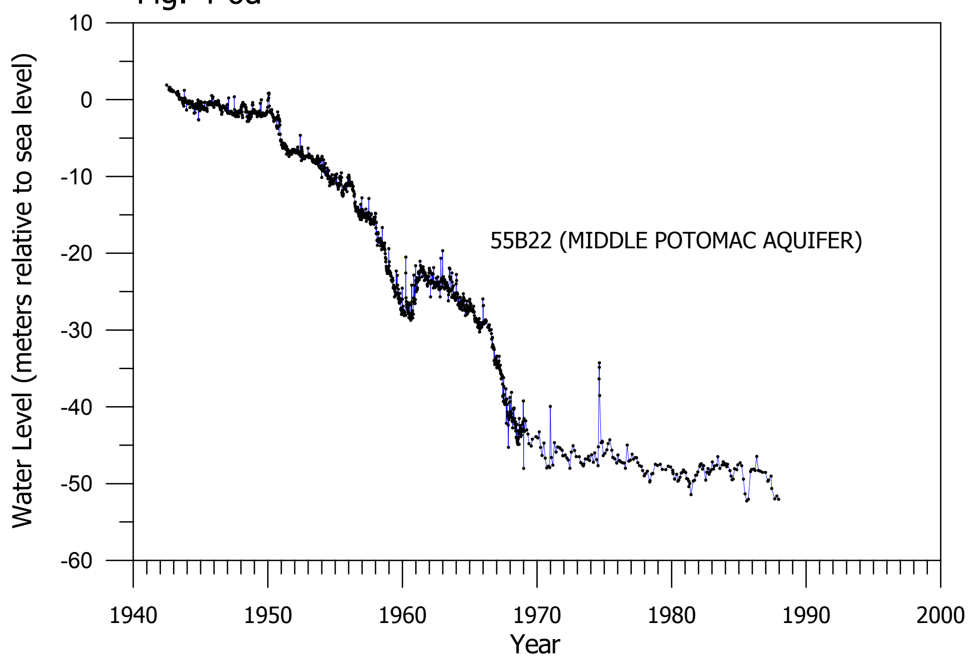


Figure 4-6 c-d: Water levels from available monitoring wells in the Franklin, Virginia area, in meters relative to the sea-level datum. Records are designated by USGS local identification numbers.

Fig. 4-6e

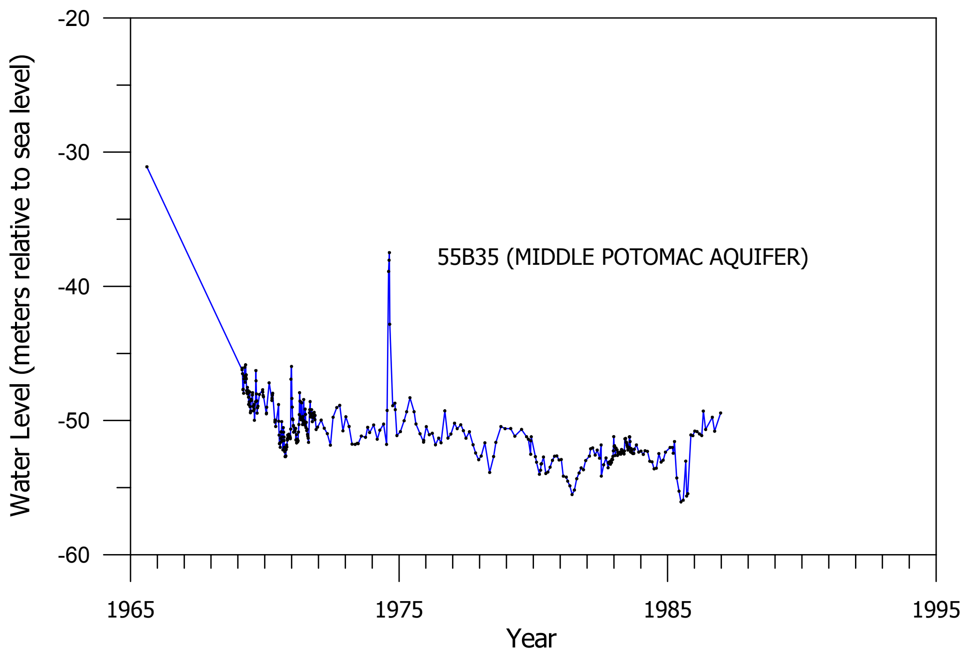


Fig. 4-6f

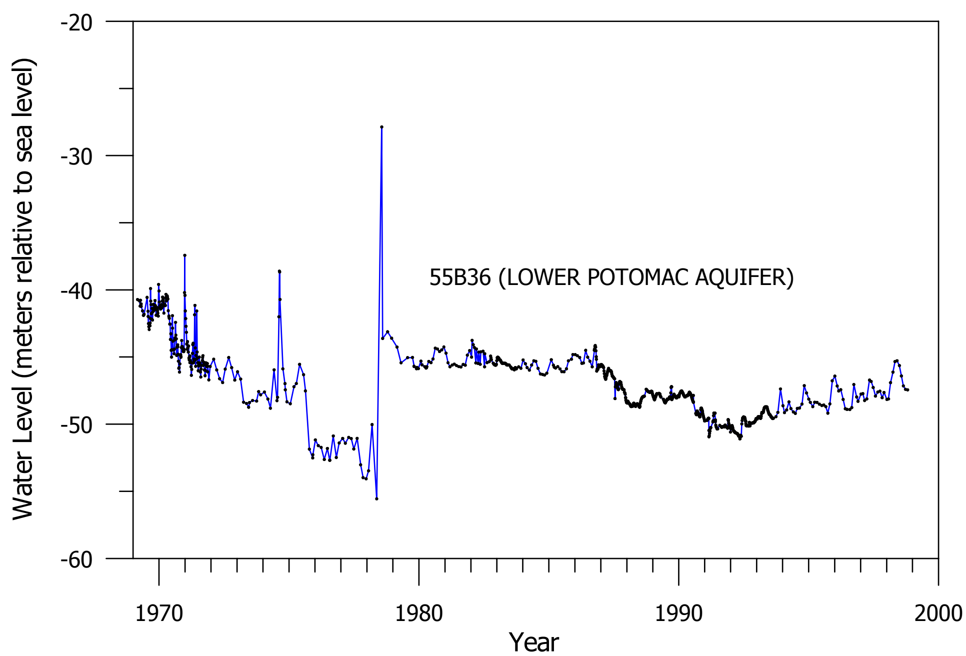


Figure 4-6 e-f: Water levels from available monitoring wells in the Franklin, Virginia area, in meters relative to the sea-level datum. Records are designated by USGS local identification numbers.

Fig. 4-7a

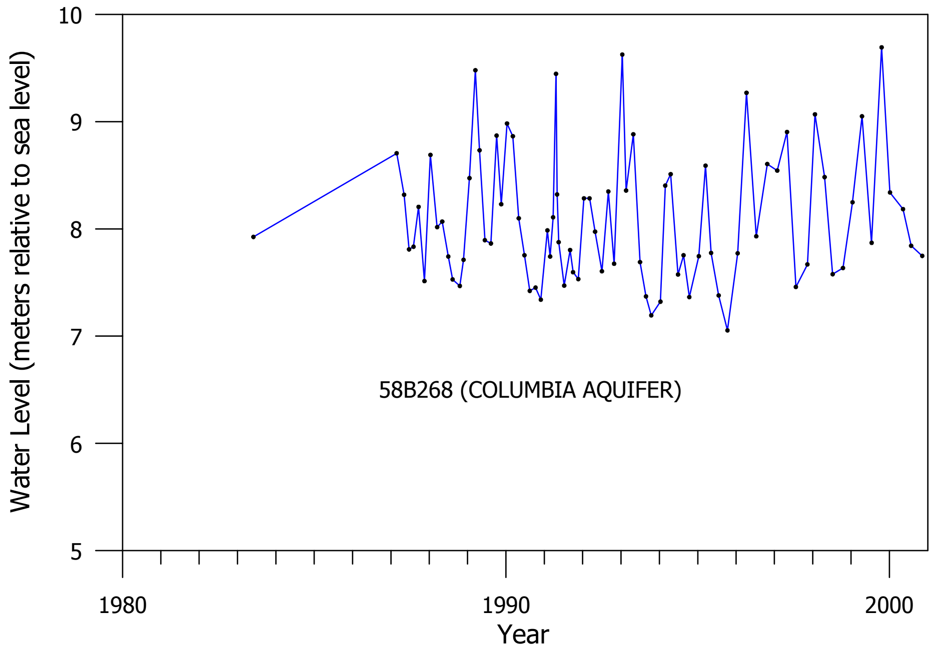


Fig. 4-7b

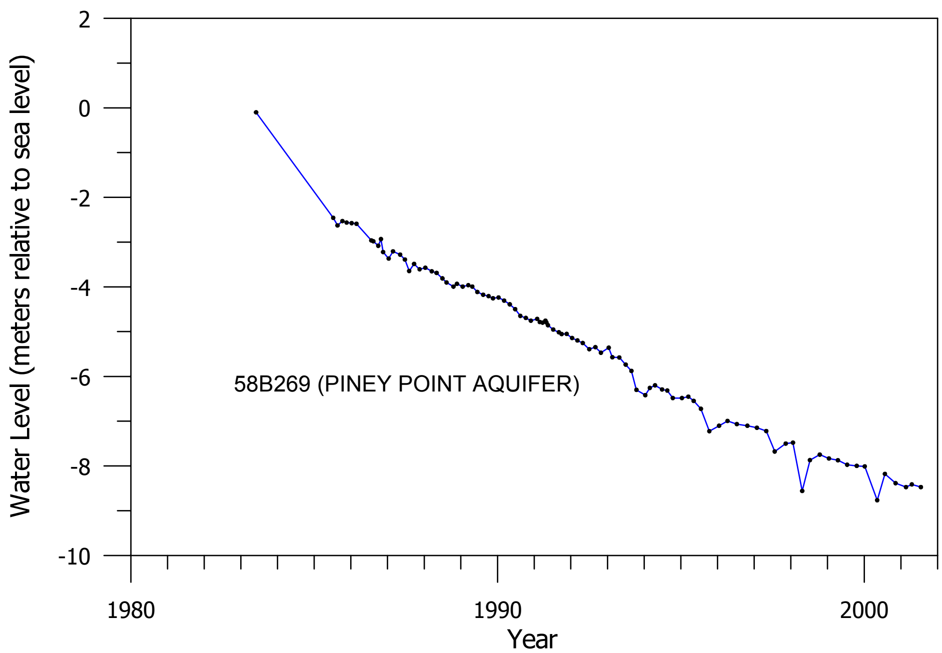


Figure 4-7 a-b: Water levels from available monitoring wells in the Suffolk, Virginia area, in meters relative to the sea-level datum. Records are designated by USGS local identification numbers.

Fig. 4-7c

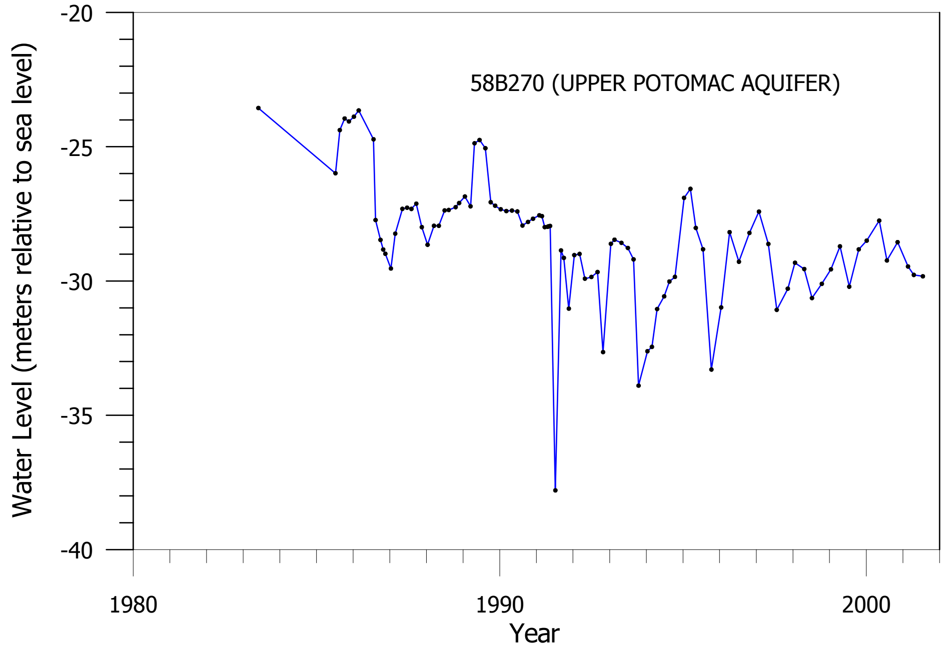


Fig. 4-7d

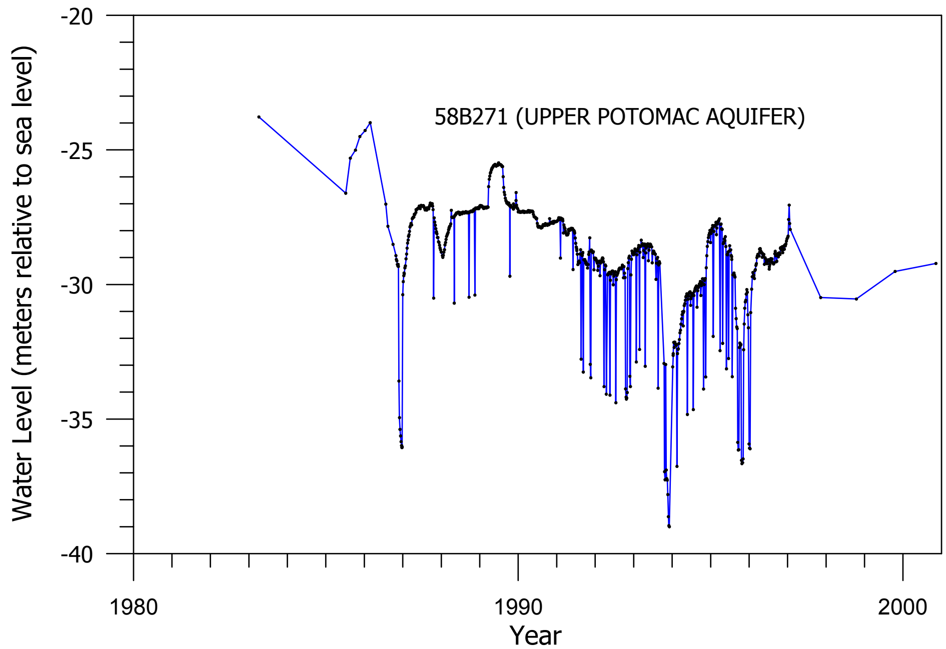


Figure 4-7 c-d: Water levels from available monitoring wells in the Suffolk, Virginia area, in meters relative to the sea-level datum. Records are designated by USGS local identification numbers.

Fig. 4-7e

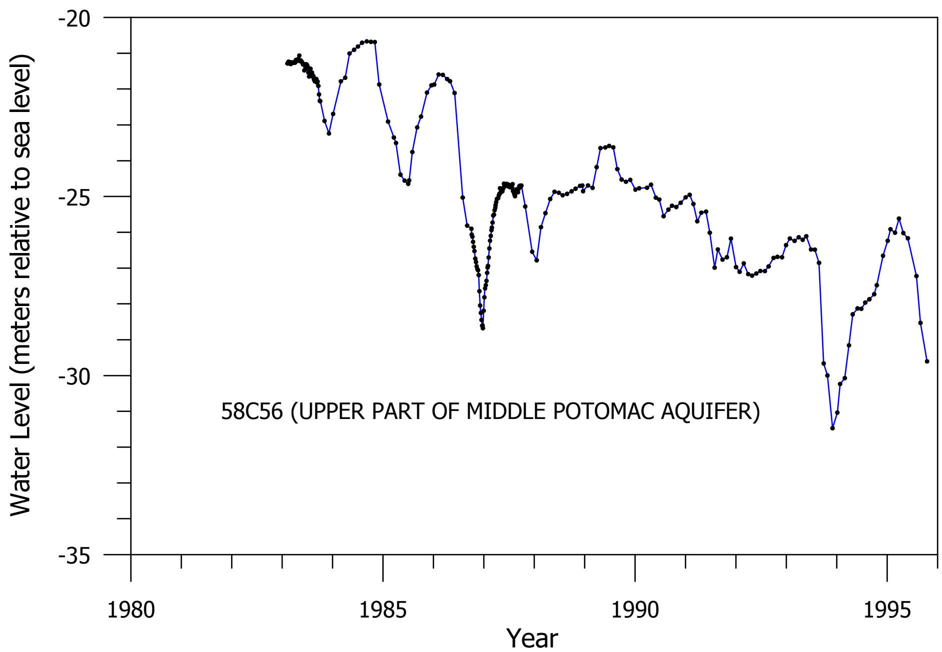


Fig. 4-7f

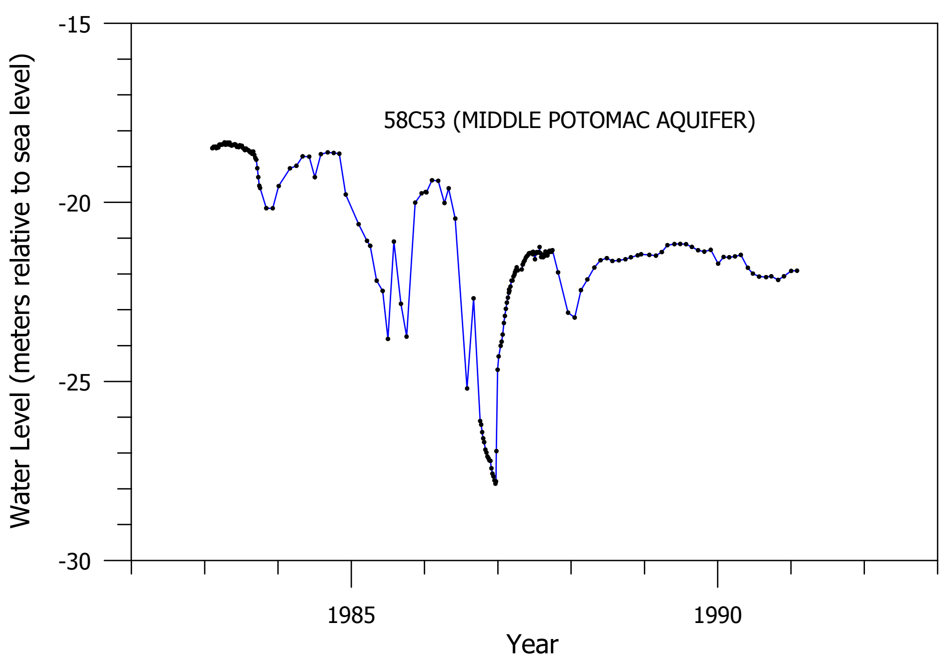
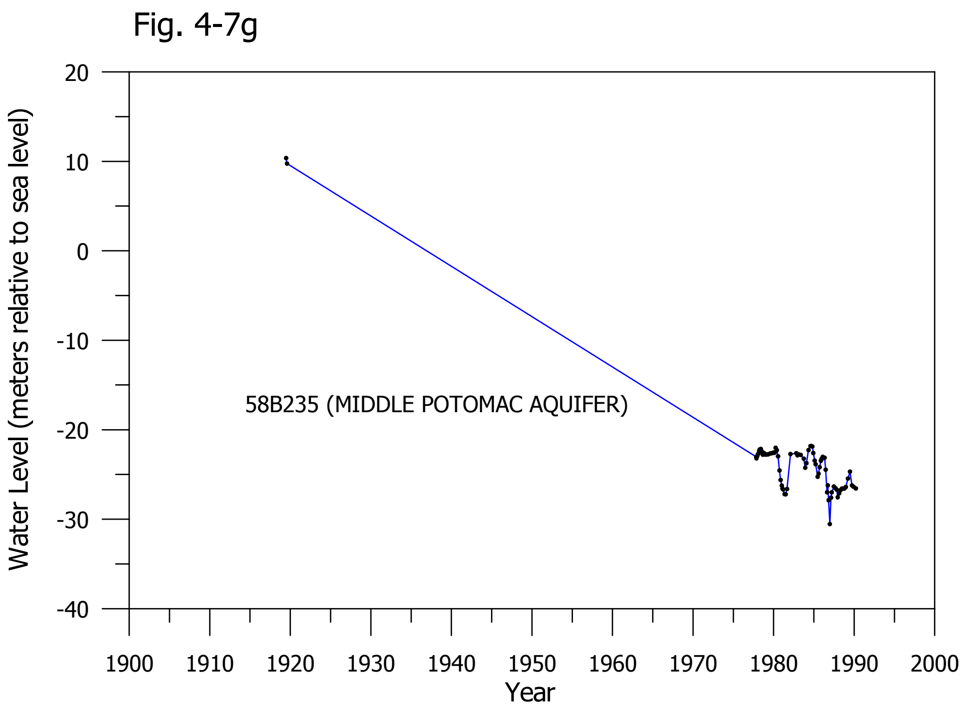


Figure 4-7 e-f: Water levels from available monitoring wells in the Suffolk, Virginia area, in meters relative to the sea-level datum. Records are designated by USGS local identification numbers.



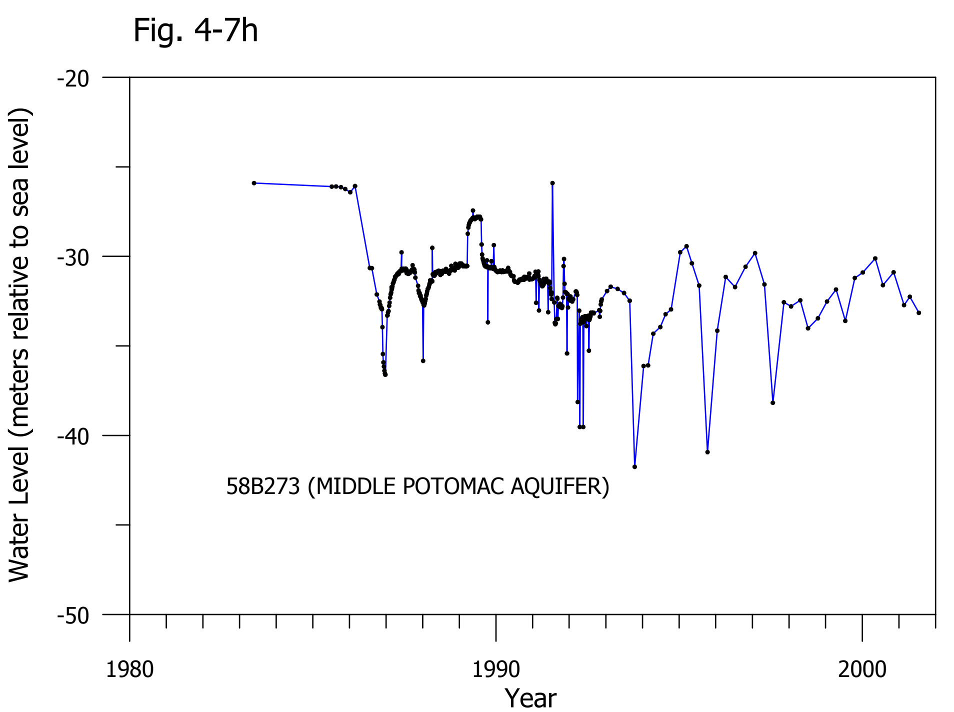


Figure 4-7 g-h: Water levels from available monitoring wells in the Suffolk, Virginia area, in meters relative to the sea-level datum. Records are designated by USGS local identification numbers.

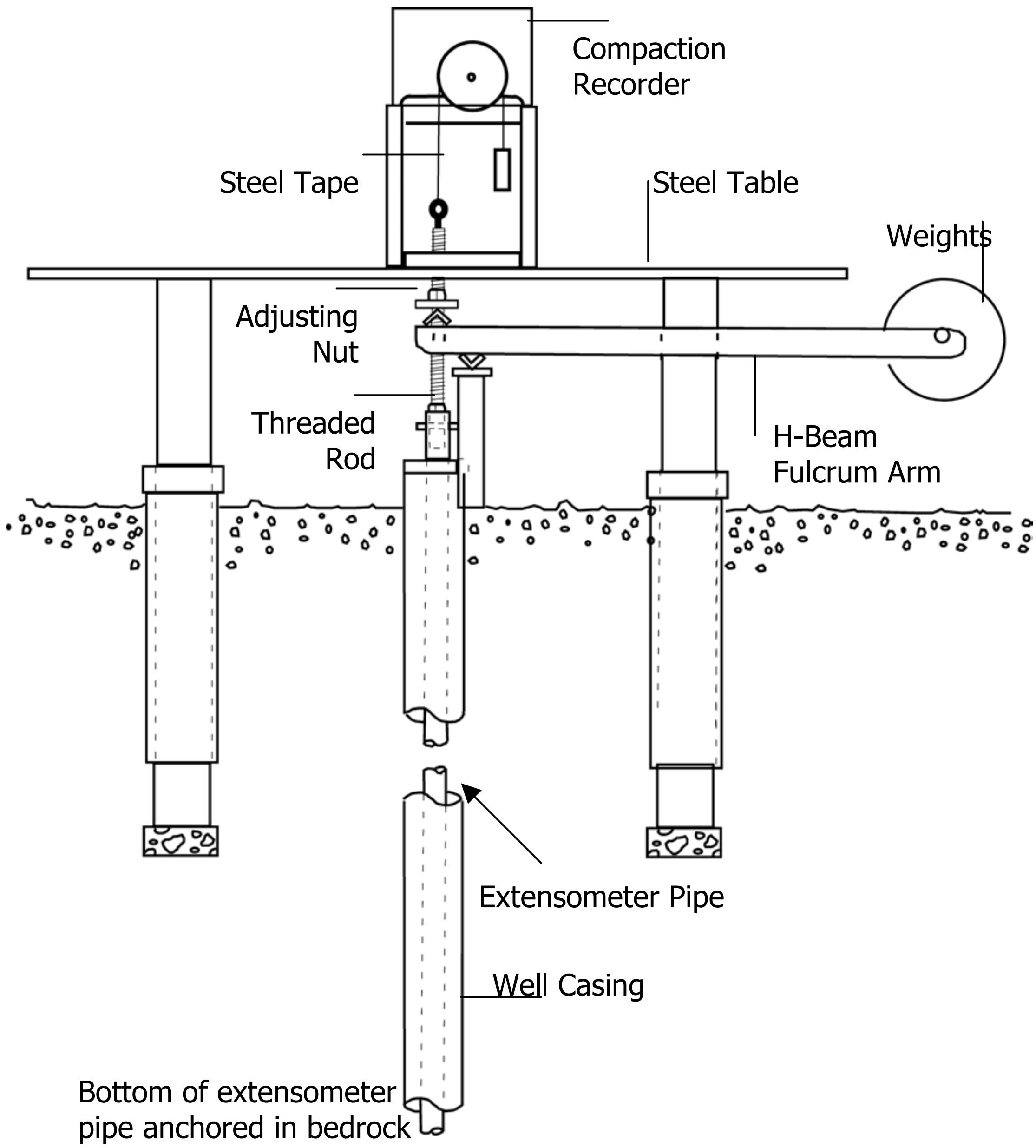


Figure 4-8: Schematic diagram of extensometer instrument.

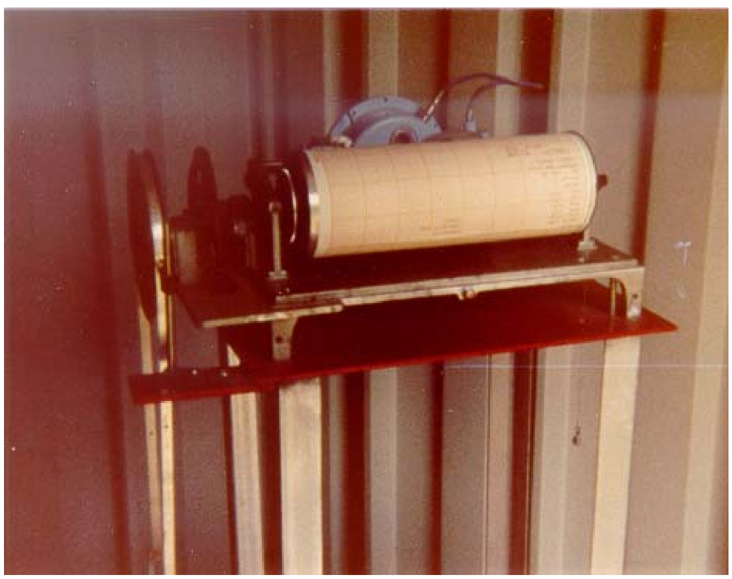


Figure 4-9: Photographs of extensometer.gauge and recorder.

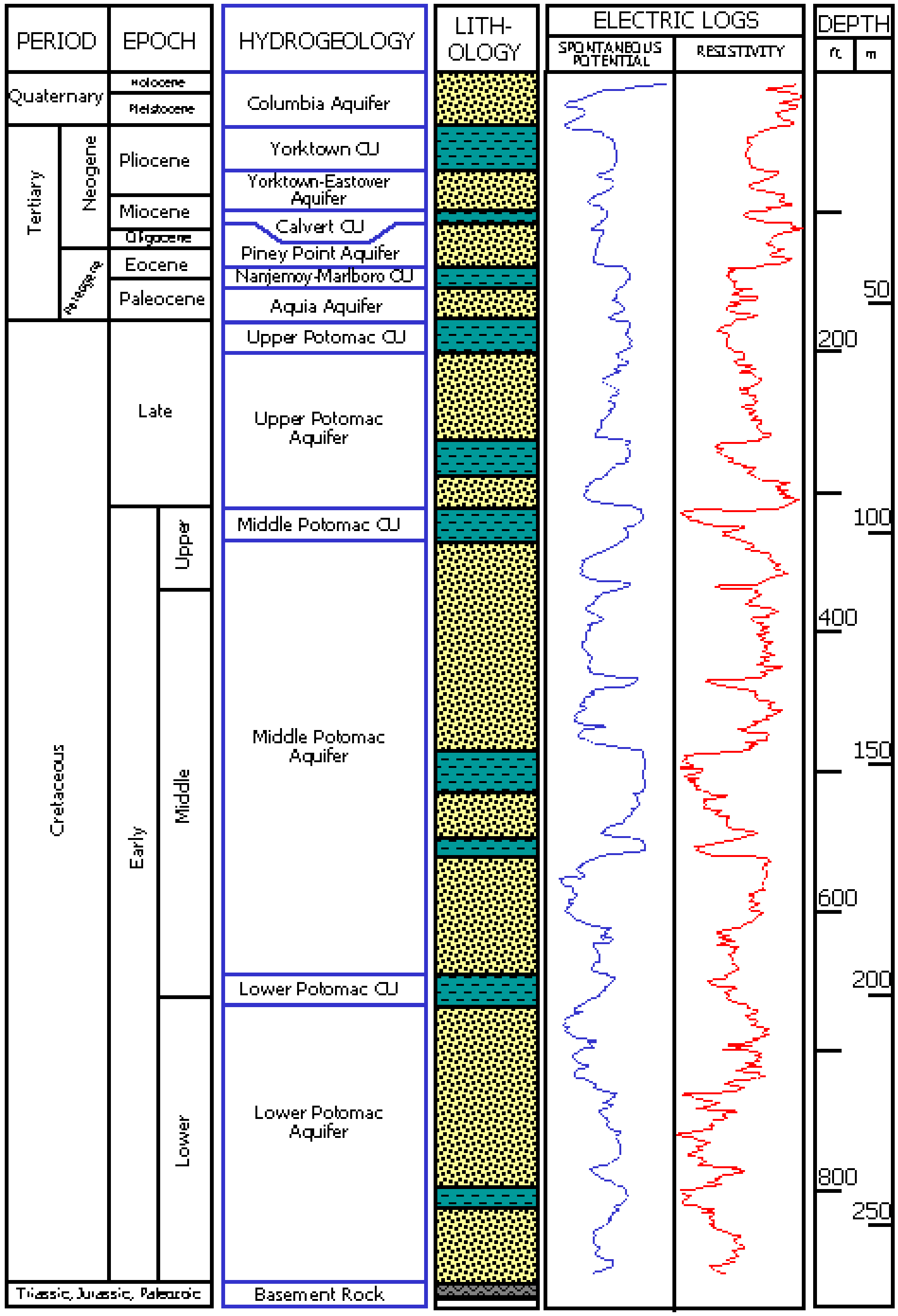


Figure 4-10: Borehole lithologic and geophysical log from the Franklin extensometer site, with hydrogeologic interpretations.

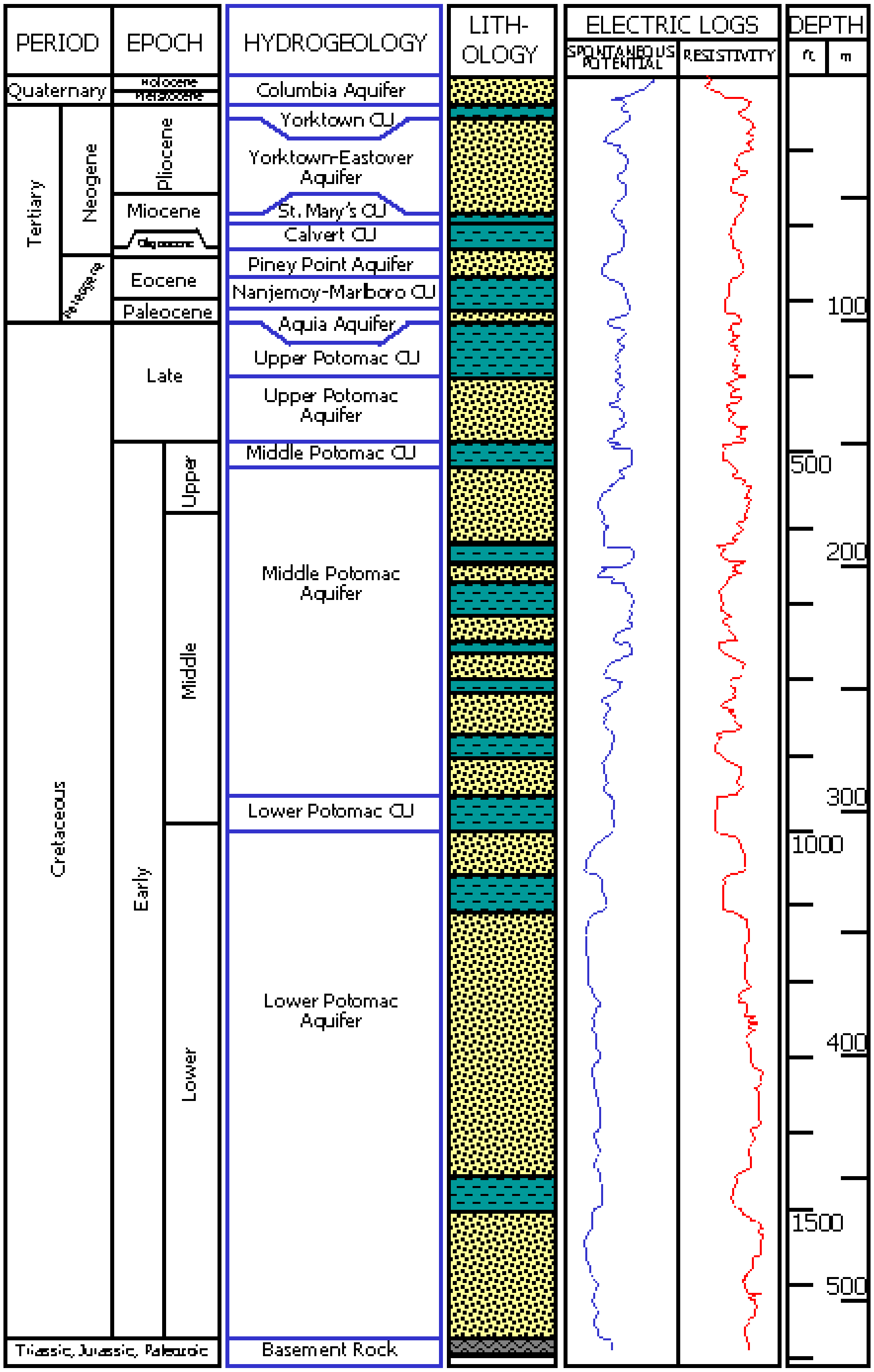


Figure 4-11: Borehole lithologic and geophysical log from the Suffolk extensometer site, with hydrogeologic interpretations.

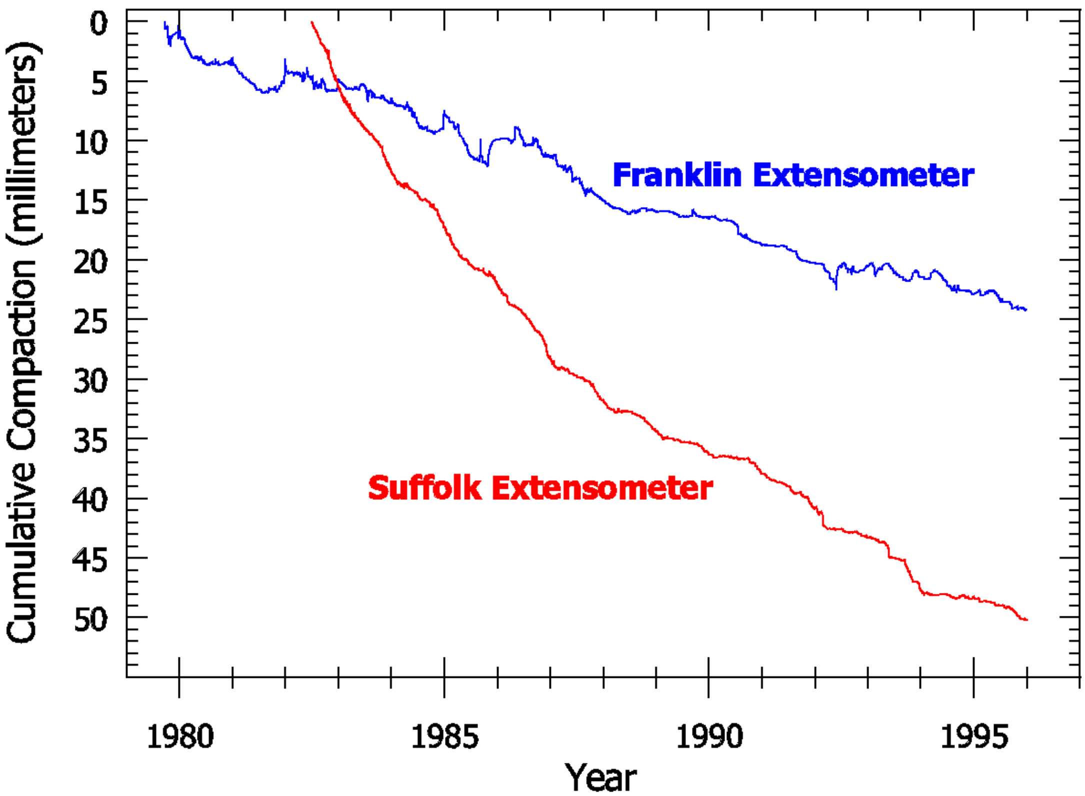


Figure 4-12: Extensometer records of measured aquifer-system compaction from Franklin (blue) and Suffolk (red).

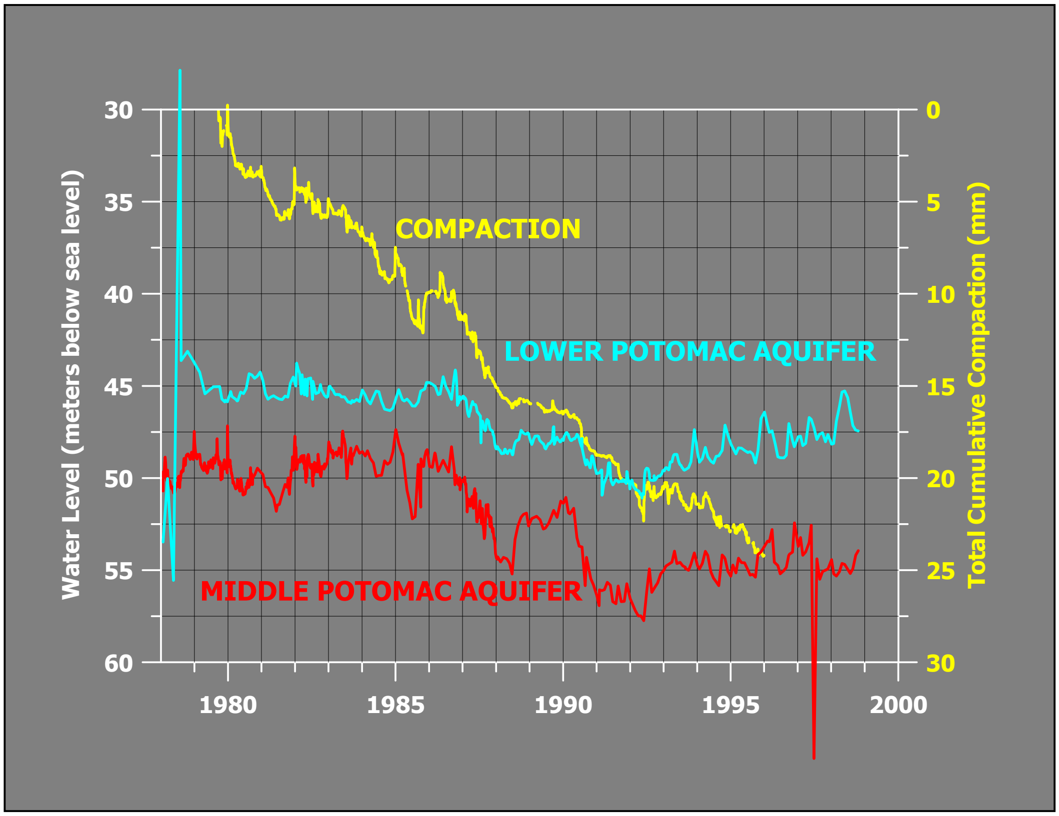


Figure 4-13: Comparison of Franklin compaction record with water levels in the Lower and Middle Potomac aquifers. Water levels units are meters below sea-level datum. Compaction units are millimeters of cumulative compaction since beginning of measured record (1979).

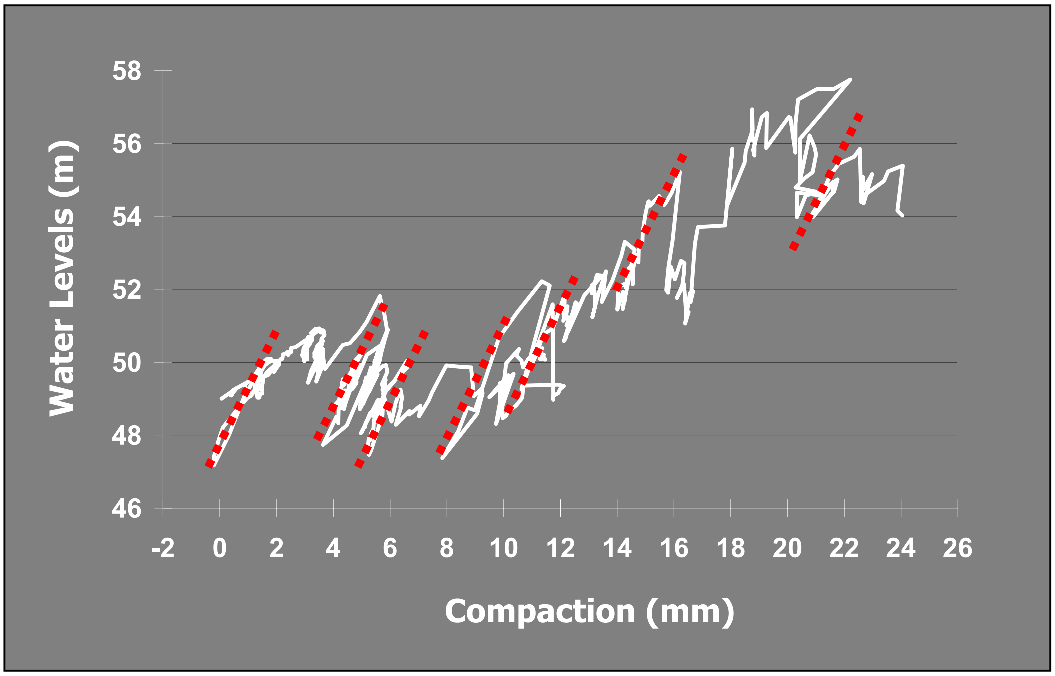


Figure 4-14: Graph of strain (meters of water) versus stress (mm of compaction) at Franklin area. Dashed lines show slopes of recoverable compaction loops.

V: ONE-DIMENSIONAL SIMULATION OF COMPACTION IN SOUTHEASTERN VIRGINIA

Because of the complexity of the Coastal Plain aquifer system, a numerical model was needed to account for vertical variations in applied stress across numerous hydrogeologic units, characterize the resulting time-dependent vertical variations in strain, and discern the contributions of individual hydrogeologic units to the total observed compaction. The one-dimensional finite-difference numerical model (COMPAC) developed by Helm (1975, 1976b) was chosen because it has been well documented and has been sucessfully used to simulate compaction in systems similar to that presented here (Epstein, 1987; Hanson, 1989). The COMPAC model uses the diffusion equation to calculate the time distribution of compaction in a hydrogeologic unit from the history of applied stress on the upper and lower boundary of the unit, as well as the unit's properties of vertical hydraulic conductivity, specific storage (recoverable and nonrecoverable), and thickness. Each Coastal Plain hydrogeologic unit was simulated separtately, but evaluation of the parameters of each unit was achieved by matching the sum of the simulated compaction in all units to the measured cumulative compaction of the complete system from extensometer data.

The relative simplicity of the COMPAC model is partly the result of the rather limited computer resources available at the time of model development, but simplicity was an advantage in this application because of the limited input and calibration data available. Despite its apparent limitations, a one-dimensional model is appropriate for simulation of compaction because of aquitard anisotropy (Helm, 1975). As Helm (1975) points out, vertical hydraulic conductivity is usually much smaller than horizontal hydraulic conductivity, which results in stresses acting in a direction that is "essentially vertical." This one-dimensional approach to simulation of aquifer-system compaction due to groundwater withdrawals has been shown to produce very accurate results in a number of cases (Epstein, 1987; Hanson, 1989).

Model Computation of Aquifer-System Compaction

The COMPAC model computes the transient vertical distribution of effective stress within an idealized hydrogeologic unit from its known stress (water-level) history and the hydraulic parameters K', S'_{skv}, and S'_{ske}. From the effective stress distribution, the model computes the resulting cumulative compaction and expansion with time at discrete horizontal planes within the unit (Helm, 1975). The development of the computational approach used in the model and outlined here were described in detail by Helm (1975, 1976b), and additional insightful comments on Helm's work were provided by Hanson (1989).

Helm (1975) pointed out that two relations are needed to compute compaction of an aquitard or confining unit from known changes in applied stress at an aquitard or confining unit boundary (i.e. in the aquifer): (1) the relation of internal effective stress to known boundary stress, and (2) the relation between the internal stress distribution and compaction. For an aquifer, it is assumed that changes in stress are instantaneously distributed throughout the thickness of the unit, so only the second relation is needed.

The vertical distribution of effective stress in a homogeneous compacting layer can be given by a linear partial differential equation (Scott, 1963):

$$\frac{\partial p}{\partial t} - \frac{\partial \sigma_T}{\partial t} = \left(\frac{K'}{S'_s} \right) \frac{\partial^2 p}{\partial z^2} \qquad (5\text{-}1)$$

where

p is the pore water stress [N/m^2],

σ_T is the total stress due to the weight of overlying materials [N/m^2]

K' is the vertical hydraulic conductivity of the compacting unit [m/d],

S_s' is the specific storage of the compacting unit [m^{-1}]

z is the vertical coordinate [m]

t is time

The rate of applied stress on the boundary of the vertically compacting layer is given by the second term on the left side of the above equation. If it is assumed that the boundary (overburden) stress is applied instantaneously at time zero and does not change, this term can be simplified to Terzaghi's formulation of the diffusion equation (Scott, 1963):

$$\frac{\partial p}{\partial t} = \left(\frac{K'}{S'_s} \right) \frac{\partial^2 p}{\partial z^2} \qquad (5\text{-}2)$$

This equation represents the distribution of pore pressure within the compacting unit. It can also be written in terms of effective stress, σ_e, because if σ_T does not change, the change in effective stress is equal but opposite to the change in the pore water stress (Helm, 1974):

$$\frac{\partial \sigma_e}{\partial t} = \left(\frac{K'}{S'_s} \right) \frac{\partial^2 \sigma_e}{\partial z^2} \qquad (5\text{-}3)$$

Because the change in p is equal and opposite to the change in σ_e, the difference between them is due only to a sign convention, and these two terms can be used interchangably for most situations involving compaction due to water-level decline. Due to the assumption of no change in σ_T, the formulations of Terzaghi and Helm presented above are less appropriate where the change in effective stress results from factors other than the reduction of head in a confined system (Hanson, 1989).

It should be mentioned that the compressibility term Ss' in the above equations has two components: one due to the compressibility of the granular skeleton (S'_{sk}) and the other due to the compressibility of water (S'_{sw}), (Helm, 1975). Because it has been shown that water is typically much less compressible than the skeleton (Scott, 1963), Helm chose to disregard the S'_{sw} component in his development of the model (Helm, 1975). However, in systems where S'_{sk} is small (such as the Coastal Plain of Virginia), the effect of water compressibility should not be disregarded altogether.

As discussed in previous sections, the term for the skeletal compressibility, S'_{sk}, has two components as well, depending on the whether the current effective stress in an aquitard or confining unit is above or below the past maximum (preconsolidation) level (Helm, 1975). When σ_e is less than the past maximum, the skeleton responds in a recoverable (pseudo-elastic) manner, and S'_{ske} is the more appropriate term. When σ_e is greater than the past maximum, the skeleton responds in a non-recoverable (inelastic) manner, and S'_{skv} is the appropriate term. A coarse-grained aquifer unit is assumed by the model to compact and expand elastically under all stress conditions, so the appropriate aquifer storage/compressibility term is S_{ske} (Hanson, 1989).

The COMPAC model uses a time-centered finite-difference numerical approximation (Crank and Nicholson, 1947) of Helms' effective stress equation (5-3) and an adaptation of the Thomas algorithm, a forward and backward substitution method (Remson and others, 1971), to calculate the transient stress distribution within an aquitard or confining unit and solve the set of equations for the effective stress at each node for each time step (Helm, 1975). Compaction is then computed with the following expression at each time step as the change in thickness ($\Delta b'$) of a bed with total thickness (b_0') (Helm, 1975):

$$\Delta b' = (0.5)\,\Delta z\left[S'_{skv}\sum_{j=1}^{J-1}\left(\Delta p^{n}_{vj} + \Delta p^{n}_{vj+1}\right) - S'_{ske}\sum_{j=1}^{J-1}\left(\Delta p^{n}_{ej} + \Delta p^{n}_{ej+1}\right)\right] \qquad (5\text{-}4)$$

where

$\Delta p^{n}_{vj} = p'\max^{n}_{j} - p'\max^{0}_{j}$,

$\Delta p^{n}_{ej} = p'\max^{n}_{j} - p'^{n}_{j}$,

Δb' is the change in aquitard or confining unit thickness,

j is the total number of nodes in the unit,

j–1 is the total number of elements in the unit, with uniform mesh spacing Δz

n is the time-step number,

p'^{n}_{j} is the effective stress at node j and time step n,

$p'\max^{n}_{j}$ is the past maximum effective stress at node j and time step n.

At any node, when p' is equal to p'max, the second (elastic) term in the brackets drops out, and compaction occurs at its nonrecoverable (virgin) rate. Otherwise, the first (inelastic) term drops out, and compaction occurs at its recoverable rate (Helm, 1975).

The above discussion, of course, applies only to a fine-grained aquitard or confining unit, but total compaction within an aquifer system (Δb*) in a given time step is the sum of aquitard and aquifer compaction: Δb* = Δb' + Δb (Hanson, 1989). The compaction of an aquifer unit is assumed to be an instantaneous (no time lag) response to a change in stress in the aquifer and is assumed to be fully recoverable (elastic) (Hanson, 1989):

$$\Delta b = S_{ske}\left(\Sigma b\right)\Delta p^{n}_{vj} \qquad (5\text{-}5)$$

The equations above present aquifer-system compaction in terms of changes in effective stress, σ_e, or in terms of the equivalent change in pore-fluid pressure, p. It is even more convenient to express this stress change in terms of equivalent hydraulic (pressure) head, h:

$$\Delta h = \Delta p/\left(\rho g\right) \qquad (5\text{-}6)$$

where

ρ is the fluid density of water, and

g is gravitational acceleration.

If it is assumed that gravity is constant, the fluid is uniform, and changes in fluid density related to the compressibility of water are negligible, then changes in head are directly proportional to fluid pressure (Sneed and Galloway, 2000). As a result, hydraulic head and pore-water pressure can be used interchangeably. In fact, the COMPAC model accomodates the use of hydraulic heads for the input stress file.

It is known that hydraulic parameters such as K', S'_{ske}, and S'_{skv} are not constants but may change under changing conditions of stress (Helm, 1976b). The COMPAC model is designed to accommodate stress-dependent parameters, and Helm (1976b) has shown that modeling with stress-dependent parameters produces more accurate simulation results under certain conditions (in particular, large changes in strain). Stress-dependent parameters were not used for the current study because no information was available to constrain the parameter ranges and very little was known about preconsolidation parameter values. Helm (1976a) showed that simulations with very small amounts of error could be produced using constant parameters, and other studies have yielded similar results, particularly for simulations over small ranges of strain (Epstein, 1987, Hanson, 1989).

The computations described in the proceeding paragraphs refer to the simulation of a single model unit. With the COMPAC model, a hydrogeologic unit can be simulated either as a separator (confining unit) between two aquifers, or as an aquifer containing one or more doubly-draining fine-grained interbeds (aquitards). A separator model unit simulates the response of a confining unit only, as heads change independently in the aquifers above and below the confining unit, while an aquifer/interbed model unit simulates the response of both the aquifer and its doubly-draining interbeds to stress (head) changes in the aquifer. Each of these approaches has its advantages and disadvantages, but together they provide the necessary tools for modeling all parts of a complex aquifer system like that found in the Coastal Plain of Virginia, and both were used in this modeling effort.

The aquifer/interbed model approach assumes that the hydraulic head is the same throughout the aquifer, and therefore acts the same on both top and bottom of all fine-grained interbeds. It is therefore most appropriate in aquifers with leaky or discontinuous interbeds that do not significantly impede vertical flow through the aquifer. For this case, where all interbeds behave in the same manner, Helm (1975) described the use of an equivalent (weighted) thickness of a representative idealized aquitard "whose calculated compaction would be consistently proportional to that obtained by summing the compaction," in a number of actual beds. Initially developed to save computational time, this approach was useful in the current study for summing the compaction response of a number of interbeds in the thick Potomac aquifers, where the lack of head data required the necessary assumptions, and where limited hydrogeologic data prevented different parameter values from being discerned for individual interbeds. The equivalent thickness formula proposed by Helm (1975) is given below:

$$b'_{0\,equiv} = 2\left(\left[\sum_{j=1}^{N}\left(b'_{0j}/2\right)^2\right]\Big/N\right)^{0.5} \qquad (5\text{-}7)$$

where

b_{0j}' is the initial thickness of the jth interbed, and

N is the number of interbeds.

As Helm (1975) pointed out, the development of this formula is based on the maximum distance for water movement out of a representative doubly-draining interbed or aquitard.

Model Approach

The Coastal Plain aquifer system was divided vertically into a number of discrete model units for both the Franklin and Suffolk sites, and each of the compacting units was simulated separately with COMPAC. Total compaction was computed for each site as the sum of the compaction in each of the model units, enabling comparison of simulated compaction records with measured compaction from the actual extensometer records. While the accurate simulation of total compaction was a goal of this exercise, identifying the contributions of individual hydrogeologic units to total compaction in the system was considered to be just as important for understanding the behavior of the system.

Aquifer-system compaction was simulated for the period 1900 to 1998 in order to span the entire history of compaction induced by groundwater withdrawals. This approach was important because the long time constants of confining units in many compacting environments results in significant delays between applied stresses (water-level declines) and the resultant strains (compaction and subsidence). Other modeling studies have reported errors resulting from stresses occurring before the start of the simulation period (Epstein, 1987). To avoid this problem and model the entire history of compaction in the Virginia Coastal Plain, these simulations were begun under the hydrogeolgic conditions present prior to significant development of the aquifers in the region.

The discretization of model units for both the Franklin and Suffolk sites was influenced both by the availability of hydraulic head (boundary stress) data for the various aquifers and by the initial interpretation of hydrostratigraphic units. These model units are presented in figures 5-1 and 5-2 and discussed later in more detail. In general, hydrogeologic units identified as aquifers (with some fine-grained interbeds) were simulated as aquifer/interbed model units, and hydrogeologic units identified as confining units were simulated as separators.

Some adjustments to this scheme were necessary due to limited data. The assumption for an aquifer/interbed model unit is that the head change throughout the aquifer is uniform and instantaneous, and the assumption for a separator model unit is that the unit acts as a distinct confining unit. Both of these assumptions are known to be more valid in the upper (marine) part of the aquifer system and less valid in the lower (fluvial-deltaic) Potomac section. As discussed in the section on the hydrogeologic setting, the fine-grained marine units form areally extensive confining units, while the thick clays of the fluvial-deltaic units are quite discontinuous. Hydraulic head data support this approach, revealing a high degree of connection between the 'aquifers' of the fluvial-deltaic Potomac system and suggesting that the concept of distinct aquifers and separators in this lower part of the system may be invalid or at least arbitrary. As a result, aquifer definitions for the model were constrained somewhat by judgements regarding which intervals were experiencing (approximately) the head changes represented in the available piezometers. Thus, the model discretization was constrained somewhat by the available geologic and hydrologic data, and water-level data in particular.

Model Input

The similarities in the hydrogeologic framework data and water-level data available at Franklin and Suffolk led to a very similar approach with regard to the discretization of model units at the two sites, depicted in figures 5-1 and 5-2. Interpretations of lithologic and geophysical logs indicated that the same hydrogeologic units are present at both locations, though most of the units are of course significantly thicker at the Suffolk site.

At both sites, model units 3, 5, 7, and 9 were simulated as aquifer units with symmetrically draining interbeds. For those units with multiple interbeds, simulations used

Helm's equivalent thickness approach, which was discussed previously. Though model unit 3 does not contain any fine-grained interbeds, compaction was simulated in this unit for both the Franklin and Suffolk sites in order to calculate any elastic compaction or expansion of the aquifer material. Model unit 1 was initially simulated as the uppermost unit at both sites, but early simulations revealed its contribution to compaction to be negligible, and it was dropped from the modeling effort thereafter.

Model units 2, 4, 6, and 8 were simulated as separators in the COMPAC model. These separators were intervals typically interpreted as fine-grained confining units between two distinct aquifers. As previously discussed, this separator/confining unit concept is more accurate for the marine section of the system, consisting of model units 2 and 4. There is no evidence that model units 6 and 8 are regionally extensive hydraulic separators; in fact, water levels indicate that the fluvial-deltaic Potomac units are somewhat interconnected. Thus, the geologic and hydrologic characteristics of the Potomac system make the identification of distinct separators somewhat arbitrary. As a result, fine-grained units were chosen as separators because they were thicker than some of the other clay units of the Potomac system and because they were located vertically at approximately the midpoints between the various available water-level measurement locations.

The simulation of unit 4 requires additional comment. Unit 4 was modeled as a separator, even though log interpretations indicate that it may contain a relatively thin (5-7 meters) sand unit, identified as the Aquia aquifer, sandwiched between the Nanjemoy-Marlboro confining unit above and the Upper Potomac confining unit below. This interval is dominated by fine-grained sediments of the two confining units, which span the transition from fluvial-deltaic to marine deposition. Furthermore, it is reported that the Aquia is not considered a viable aquifer in southeastern Virginia (McFarland, oral communication, 2001), so it is not generally exploited as a source of water, and water levels are generally not available. Given these conditions, the treatment of this unit as a separator is considered valid, especially because the Nanjemoy-Marlboro confining unit is known to contain a distinctive and extensive clay layer known as the Marlboro clay. The fine-grained units dominate the thickness of unit 4, and they would also be expected to dominate the vertical hydraulic conductivity and storage characteristics of the unit.

Input parameters for each of the model units included interval depth and thickness, initial preconsolidation (past-maximum) stress distribution, recoverable and nonrecoverable specific storage, and vertical hydraulic conductivity. The dimensions of each model unit are given in table 5-1 for Franklin and 5-2 for Suffolk. The hydraulic parameters associated with each unit are discussed in the calibration section. Each of the separator units required input files of the water-level records for the aquifers above and below the separator, and each of the aquifer/interbed units required an input file of the water-level record in the aquifer.

Table 5-1: Model Units for Franklin Simulation

Model Unit	Unit Type	Total Unit Thickness (m)	Thickness of Fine-grained Sediments (m)	Thickness of Aquifer Sediments (m)	Depth to Unit Midplane (m)	Stress Files
02	Separator	2.82	2.82	-----	31.52	YE, PP
03	Aquifer/Interbed	9.88	0.00	9.88	37.87	PP
04	Separator	17.87	17.87	-----	51.75	PP, UP
05	Aquifer/Interbed	19.88	0.00	19.88	70.32	UP
06	Separator	8.00	8.00	-----	83.96	UP, MP
07	Aquifer/Interbed	59.74	7.06	52.68	117.83	MP
08	Separator	9.41	9.41	-----	152.41	MP, LP
09	Aquifer/Interbed	106.29	14.76	91.53	210.55	LP

Table 5-2: Model Units for Suffolk Simulation

Model Unit	Unit Type	Total Unit Thickness (m)	Thickness of Fine-grained Sediments (m)	Thickness of Aquifer Sediments (m)	Depth to Unit Midplane (m)	Stress Files
02	Separator	13.45	13.45	-----	63.20	YE*, PP
03	Aquifer/Interbed	11.65	0.00	11.65	75.75	PP
04	Separator	39.44	39.44	-----	101.29	PP, UMP
05	Aquifer/Interbed	82.46	24.20	58.26	162.24	UMP
06	Separator	12.55	12.55	-----	209.75	UMP, MP
07	Aquifer/Interbed	104.90	34.95	69.95	371.08	MP
08	Separator	14.34	14.34	-----	328.06	MP, LP
09	Aquifer/Interbed	172.07	14.34	157.73	421.27	LP

* indicates constant value used for the Yorktown-Eastover water level

The hydraulic head records used as stress files for the model were constructed from both measured data and historical reports of water levels in an effort to simulate historical variations in head from predevelopment conditions to the present. Available water-level data (figures 4-6 and 4-7) from wells near the extensometers were used to construct water-level records for recent years for the aquifers of interest. However, the longest continuous water-level record near either extensometer location began in 1960, and several of the most useful records began in the early 1980s. To supplement the known records, historical measurements and qualitative descriptions from older reports were used to estimate water levels in the areas of interest. Comprehensive reports by Sanford (1913) and Cederstrom (1945) on groundwater conditions in eastern Virginia were particularly useful for this purpose. Measured data and historical estimates were combined, and linear interpolation was used to fill gaps and produce complete records of approximately monthly (30-day) values from 1900 to 1998.

At both the Franklin and Suffolk locations, additional water-level records were necessary for better vertical discretization of the model. At Franklin, a water-level record was needed for the Upper Potomac aquifer (UP), while a record for the Lower Potomac (LP) aquifer was needed at Suffolk. These water levels were estimated and interpolated using data from nearby wells in adjacent aquifers, as well as relations developed from more distant wells in the aquifers of interest. This process resulted in one 'estimated' water-level record for each of the two locations, but it should be noted that much more uncertainty is associated with these two estimated records.

The constructed water-level records for Franklin and Suffolk are presented in figures 5-3 and 5-4, respectively. The five water-level records constructed for Franklin include the Yorktown-Eastover (YE), Piney Point (PP), Upper Potomac (UP), Middle Potomac (MP), and Lower Potomac (LP) aquifers. The four water-level records constructed for Suffolk include the Piney Point aquifer (PP), the upper part of the Middle Potomac aquifer (UMP), the Middle Potomac aquifer (MP), and the Lower Potomac aquifer (LP). A water-level record for the Yorktown-Eastover aquifer was not available at Suffolk, so the value was estimated to be a constant 5 meters above sea level, approximating the Franklin conditions for the same aquifer. The water-level record(s) at each location associated with each of the model units is given in tables 5-1 and 5-2. These water levels are adjusted to reflect the head in meters above or below the sea-level datum (as opposed to land surface), and they also reflect a reversal of the sign convention typically used for hydraulic heads. The model input required negative values for heads above the datum and positive values for heads below the datum, and the data reflect this change. In these graphs, heads decrease in the downward (positive) direction and increase in the upward (negative) direction. The circles in figures 5-1 and 5-2 show the vertical locations of these water level measurements at each of the two sites.

Initial values of hydraulic conductivity and specific storage were developed from available aquifer tests and from calibration of a regional groundwater flow model (Harsh and Laczniak, 1990). However, aquifer tests typically do not reflect the effects of delayed release of water from confining units, so these values were not expected to be accurate for the fine-grained model units. The regional flow model produced by Harsh and Laczniak was only a quasi-three-dimensional model and did not explicitly represent confining units or consider the effects of confining-unit storage, so the model-derived values were only useful as starting points for the compaction model.

An initial preconsolidation stress value was needed for each model unit in order to control the stress condition at which the compaction of an interbed or confining unit transitions from recoverable (elastic) to non-recoverable (inelastic) behavior. Typically, this type of information is obtained from laboratory measurements of core samples and from estimates derived from field observations of system behavior. In this case, no laboratory measurements were available, and only meager information was available from the field. It was noted by Davis (1987) that significant land subsidence had only been observed in areas of the Atlantic Coastal Plain experiencing more than approximately 20 meters of head decline from predevelopment conditions. This observation provided an intitial estimate of the preconsolidation stress for the system as a whole, but the value of this parameter would be expected to vary somewhat with depth. Ultimately, no information was available to better constrain values of this parameter, and a value equivalent to 20 meters of drawdown was used for all model units.

Other input parameters important to the numerical model included the time interval (dt), the maximum computational time step (deldt), and the vertical grid spacing (dz). A time interval of 30 days was chosen to correspond to the approximate resolution available in the input stress (water-level) files. The computational time step was adjusted until the best stability and performance of the numerical simulation was achieved, and the optimal value was found to be 3. Not surprisingly, this value corresponds to Helm's suggestion in the model documentation that the best deldt value should be approximately one-tenth of the dt value. A vertical grid spacing of 0.01 meters was chosen to provide good spatial resolution, but it was discovered with some experimentation that variations in dz across a range of reasonable values did not significantly affect the computed compaction for a model unit.

Model Calibration

Each model unit was simulated separately, and the compaction records for all model units were summed to produce a total compaction record for the aquifer system at each site. The simulated record was compared to the measured extensometer record to check the validity of the simulation. Using an organized trial-and-error approach, values for hydraulic conductivity, recoverable and nonrecoverable specific storage, and preconsolidation stress in each of the units were adjusted to calibrate each model to (1) the measured extensometer record and (2) the estimates of total cumulative compaction presented in table 4-3.

Initial calibration of the model was focused on the Franklin location for several reasons. First, vertical discretization of the model was considered to be more accurate at Franklin due to greater confidence in field data. In addition, much more complete water-level data were available for Franklin. A larger number of useful monitoring wells were located near the Franklin extensometer installation, and the water-level records spanned much longer periods of time. Perhaps most importantly, the sensitivity of the Franklin extensometer was much better than that of the Suffolk instrument, providing a more accurate compaction record with which to match the simulated record. For these reasons, the calibration effort was first applied to the Franklin model, and the parameter values developed for the units at Franklin were used in the Suffolk model for verification.

In the calibration process, values of nonrecoverable specific storage, recoverable specific storage, vertical hydraulic conductivity, and preconsolidation stress were adjusted for each model unit in an attempt to match the total measured record. An understanding of the sensitivity of the model behavior to each of these parameters was useful in this process. The maginitude of S'_{skv} in the confining units and interbeds was most important in controlling the magnitude of total compaction observed, while the values of S'_{ske} were important in controlling the magnitude of aquifer-system expansion during periods of water-level rise. On the other hand, the rate and pattern of compaction was controlled by the K_v values of the fine-grained units, which essentially controlled the speed of the compaction response to aquifer water-level changes. For constant S'_{skv} and S'_{ske}, lower K_v values resulted in very slow drainage of the fine-grained layers and long time constants. Higher K_v values resulted in much more rapid compaction responses to changes in heads in the aquifers.

Because the model assumes that a change in head occurs immediately throughout an aquifer unit, the model does not use K_v values for coarse-grained intervals. However, values for S_{ske} in the aquifer units were needed, and they proved to be important to model calibration. The S_{ske} value for the aquifer units controlled the magnitude of the instantaneous compaction and expansion events associated with head changes. These instantaneous events were easy to pick out of the measured compaction record, so S_{ske} for the aquifers could be calibrated with a relatively high level of confidence. On the other hand, variations in S_{ske} among the aquifer units were not large enough to create significant changes in the instantaneous component of the compaction response, so identical values were ultimately used for all aquifer units.

The value of preconsolidation stress for a model unit would be expected to have a potentially large effect on calibration because it controls the point of transition from S'_{ske} to S'_{skv}, and S'_{skv} is typically much greater than S'_{ske}. However, using the preconsolidation stress value to calibrate this model was problematic because no data were available to describe the change in preconsolidation stress with depth. As a practical matter, the intitial level of preconsolidation stress was likely exceeded decades before accurate measurements of compaction were available, so little measured data were available to help with the calibration of this parameter across the transition between the elastic and inelastic stress regimes. However, the preconsolidation stress value affects the total magnitude of compaction more than the slope of the compaction record, so the accurate simulation of this parameter was considered to be less important. Instead, the values for the S'_{skv} parameter received the most attention for calibration of the total magnitude of compaction. Ultimately, calibration of the model was achieved using a preconsolidation stress value of approximately 20 meters of head for all model units, the value estimated by Davis from an analysis of subsidence in the Atlantic Coastal Plain (1987).

Changing values of thickness for model units, particularly confining units and interbeds, would obviously have a large effect on calibrated compaction results. However, model unit depths and thicknesses were known with greater certainty than any of the other model parameters, due to the availability of good lithologic and geophysical logs. Consequently, model unit dimensions were not adjusted in the calibration process.

Calibration of the total compaction model for the aquifer system ultimately involved adjusting the values of S'_{skv}, S'_{ske}, S_{ske} and K'_v for each of eight model units. With this many variables (4 parameters x 8 model units = 32 variables), it is intuitively obvious that many combinations of parameter values may produce a simulated compaction record matching the measured record. In fact, a number of researchers have addressed this problem of 'nonuniqueness,' though Helm (1976b) gave the problem the most thorough analysis.

Even if we consider only one discrete model unit, nonuniqueness is still a problem because of the relation between S'_{sk} and K'_v in the stress distribution equation (number?). A number of combinations of parameter values could conceiveably produce identical values for the term K'_v/S'_s, and consequently, for compaction versus time. This problem is given a quantitative treatement by Helm (1976b), who also outlines a quantitative solution to the problem (based on a numerical analysis of the K,S plane) in the unpublished report documenting the COMPAC model.

For the purposes of the current modeling effort, the problem of nonuniqueness was approached in a more qualitative, if somewhat less satisfying, manner. Values of K_v' and S_s' were constrained by a consideration of the geologic reasonability of their values for the various units. In terms of the calibration process, values for K'_v and S'_s were adjusted from their initial estimated values in a stepwise fashion in consecutive model runs. Parameter values considered or known to be hydrogeologically unreasonable (such as impossibly small K'_v) were discarded. This approach does not guarantee that there are no other reasonable combinations of parameter values that will produce calibration of the model, and it was almost certainly not the most efficient technique. Nonetheless, it was ultimately successful in generating a plausible and geologically reasonable calibration of the compaction model. Ultimately, enough different combinations of values were applied to the model to produce confidence in this approach.
Once a reasonable calibration was achieved for the Franklin model, the calibrated parameter values for each unit were applied to the Suffolk model for verification of the results. Little evidence exists to indicate that storage and conductivity parameters of the hydrogeologic units are significantly different at the two sites, so applying the Franklin parameters to the Suffolk model was considered the most valid approach, given the limitations of the Suffolk data described previously.

Discussion of Modeling Results and Sensitivity

The simulated historical compaction record at the Franklin site is presented in figure 5-5, along with the simulated compaction for each model unit and the measured compaction record from the extensometer. A comparison of simulated and measured compaction is shown in greater detail in figure 5-5a. Given the uncertainties in this modeling effort due to lack of data, the match between the simulated and observed record is good. The parameter values for each unit used to produce this calibrated simulation are presented in table 5-3.

Table 5-3: Calibrated Hydraulic Conductivity and Storage Parameters from Compaction Model

Model Unit	K'_v (m/yr)	S'_{skv} (m^{-1})	S'_{ske} (m^{-1})	S_{ske} (m^{-1})
02	2.0×10^{-3}	1.5×10^{-4}	6.0×10^{-6}	4.5×10^{-6}
03	2.0×10^{-3}	1.0×10^{-4}	6.0×10^{-6}	4.5×10^{-6}
04	2.0×10^{-3}	1.0×10^{-4}	6.0×10^{-6}	4.5×10^{-6}
05	5.0×10^{-4}	5.2×10^{-5}	6.0×10^{-6}	4.5×10^{-6}
06	6.6×10^{-4}	5.0×10^{-5}	4.5×10^{-6}	4.5×10^{-6}
07	6.1×10^{-4}	1.5×10^{-5}	4.5×10^{-6}	4.5×10^{-6}
08	4.4×10^{-4}	1.5×10^{-5}	4.5×10^{-6}	4.5×10^{-6}
09	4.4×10^{-4}	1.5×10^{-5}	4.5×10^{-6}	4.5×10^{-6}

The calibrated simulation of compaction at Franklin is satisfactory because the simulated record matches the measured compaction over the period 1979 through 1995 and because the total cumulative simulated compaction at the end of 1995 approximately matches the estimated total historical compaction (table 4-3). More important than the total record, however, is what this simulation reveals about the behavior of the aquifer system.

The most important result of this modeling effort is that ongoing compaction in recent years at Franklin has been driven not by compaction of the Cretaceous clays (model units 6, 7, 8 and 9) but instead by compaction of the marine clays of the confining unit overlying the Cretaceous system (model unit 4). The fine-grained sediments of unit 4 exhibit a significantly higher S'_{skv} value and somewhat lower K'_v than the Cretaceous clays, but it is more important to note that this is the only unit in the system that has experienced significant head declines in recent decades. The relatively thick unit 4 also has a low K'_v value, which means that changes in head above and below this separator take a long time to permeate through the unit. Thus, the unit is still internally equilibrating (and compacting) due to past head decline in the aquifer below (water-level record UP) even as it compacts due to ongoing head decline in the aquifer above (water-level record PP). As a result, this unit is responsible for approximately 40 percent of the total historical compaction and almost all of the compaction in the final decade of the modeled time period.

In contrast, the fine-grained interbeds and confining units of the Cretaceous system are relatively thin and have much shorter time constants, typically ranging from several months to several years. As a result, head changes in the Cretaceous aquifers (water-level records LP, MP, and UP) permeate through these units relatively quickly. From their simulated behavior, it appears that model units 6, 7, 8, and 9 have equilibrated from large water-level declines in the 1940s, 1950s and 1960s, and they generally experienced only minor water-level declines in the last few decades of the twentieth century. Though the interbeds and confining units of the Cretaceous system are collectively responsible for just under half of the total historical compaction, almost none of the ongoing compaction is occuring in these units. In fact, at least one of these units, unit 9, exhibited rebound or expansion in the last decade of the simulation. On the other hand, all of the Cretaceous model units (6, 7, 8, and 9) have clearly experienced head declines well past their preconsolidation stress levels, and the current stress in these units appears to be right on the threshold of the past maximum stress. This means that even small further declines in heads beyond current levels could cause increased rates of compaction in these units by moving the compressibilities into the non-recoverable regime.

Once the Franklin model was properly calibrated, the parameters used in that model were applied directly to the equivalent units in the Suffolk simulation for preliminary verification. This approach was taken because an approximation of the measured Suffolk record with parameter values obtained from the Franklin model should confirm the validity of the parameters. Thus, the Suffolk model was run with unit dimensions and water levels specific to the Suffolk site but with unit parameters that had been calibrated in the Franklin simulation.

The simulated total compaction record for Suffolk is presented in figure 5-6, along with the compaction records for the various model units and the measured extensometer record for comparison. The magnitude of the simulated compaction matches both the total estimated value and the 1982-1995 measured value remarkably well, though the pattern of the simulated compaction obviously deviates somewhat from the measured record. Some additional calibration of the Suffolk model by adjusting the parameter values was attempted unsuccessfully. Specifically, it was impossible to adjust the conductivity and storage values to prevent the expansion events exhibited in the simulated record while preserving the total magnitude of compaction. Ultimately, the conclusion was reached that much of the deviation of the simulated record from the observed was probably due to limitations in precision of the measured record due to the lack of precision of the Suffolk extensometer and recorder.

Explicit calibration of the Suffolk model was problematic due to the limitations in the measured data that had been first noticed during initial inspections of the extensomter record. Even at first glance, the measured extensometer record did not appear to respond closely to nearby water-level changes. In particular, short-term compaction and expansion that normally occur in the coarse-grained aquifer sediments were not apparent in the measured record. Thus, an apparent lack of resolution in the measured data constrained the appropriate comparison of the modeled and measured records.

It is also possible that at least a portion of the simulation error noted here is the result of conceptual problems with the model related to the model assumptions. Specifically, the assumption that changes in input heads occur instantaneously throughout the coarse-grained aquifer materials may be less correct for the Suffolk situation, for several reasons. First, the much greater unit thickness at Suffolk would require actual head changes to be equilibrated over much greater vertical distances within the aquifers. This is particularly problematic in light of the greater number and thickness of clay units in the Suffolk system, especially in the Middle Cretaceous part of the system. Heads measured at the screen intervals in this part of the system seem less likely to represent heads throughout the whole model unit, and this condition is an important modeling assumption. If heads measured at the indicated vertical intervals are not instantaneously transmitted through the real aquifer system, the elastic compaction and expansion of the system may be more subdued than the simulated record would indicate, even if unit parameters were correctly assigned. This condition could result in a measured compaction record similar to that seen at Suffolk (fig 4-12).

An important constraint on the calibration of the Suffolk record was the fact that model unit parameters would be expected to be very similar at the two locations. Though some minor differences could be expected, large departures in parameter values at these two locations would cast doubt on the validity of the modeling exercise. When a simulated compaction record could not be produced (with any hydrologically reasonable parameter values) to match the measured extensometer record, this was further evidence of problems in the measured Suffolk record. In the final analysis, it was considered most appropriate to model the Suffolk compaction with the Franklin-calibrated values and explain the reasons for the differences between the two records. These are the results presented in figure 5-6.

The modeling effort for Suffolk reveals some interesting things about the behavior of the system at that location. As in the Franklin situation, a large part (approximately 30 percent) of the total historical compaction at Suffolk is due to the compaction of model unit 4. Almost all of the rest of the total compaction record (65 percent) is due to compaction in the model units representing the Cretaceous system (units 5, 6, 7, 8, and 9). The differences between compaction at the Franklin and Suffolk locations is apparent when the ongoing record is examined. At Suffolk, simulation results indicate that the Cretaceous system (units 5, 6, 7, 8 and 9) was still undergoing compaction over the last decade of the modeled record. While water levels near the end of the simulation were still at least 25 meters higher at Suffolk than at Franklin, the Suffolk levels were still falling in the last two decades, while water levels at Franklin had leveled off.

The effect of continued declines in water levels at Suffolk helps to explain why measured (extensometer) compaction rates were much higher at Suffolk than at Franklin over the period of record, despite much lower water levels at Franklin. Of course, another important factor in explaining the differences at the two sites is the difference in total thickness and thickness of fine-grained units. Unit thicknesses at Suffolk are roughly double those at Franklin, which means that greater compaction rates could be driven by smaller declines in heads. Simulation results also indicate that the compaction of units 4, 5, 6, 7, 8, and 9 is occurring in the nonrecoverable regime and will continue to do so as long as water levels continue to decline.

By simulating the compaction of each unit separately, it is possible to divide the total compaction record for the aquifer system into its aquifer component (which is elastic and recoverable) and its confining unit/aquitard component (which has both recoverable and nonrecoverable components). Unfortunately, the model design does not allow the simulated compaction from confining units and aquitards to be divided into recoverable and nonrecoverable components as well. Both the Franklin and Suffolk simulated compaction records are shown in figures 5-7 and 5-8 along with their aquifer and fine-grained components.

Separating the records into these two components provides some interesting additional information about compaction in the system. Distinguishing the compaction in the aquifers from the confining units reveals the percentage of the total that would be recovered quickly if water level rose. At Franklin, approximately 35 percent (or approximately 47 mm) of the total historical compaction appears to be the result of elastic compaction within the aquifers. The portion of compaction due to the aquifer material at Suffolk is close to 35 percent as well. figures 5-7 and 5-8 also reveal very little elastic expansion of fine-grained units at Franklin or Suffolk over the simulation, though the rate of compaction clearly slowed in these units at Franklin near the end of the century. At Suffolk, both the fine-grained units and the aquifers continue to compact at a steady rate, with only short-term fluctuations in compaction of the aquifer units.

The simulated compaction records in 5-5 and 5-6 also reveal the inflection point where the compaction of fine-grained units first becomes nonrecoverable or inelastic and the compressibility term changes. This change occurs just before 1960 at Franklin and at approximately 1970 at Suffolk. At both locations, the component of compaction occurring in the fine-grained units begins to dominate the total record shortly after this transition, because the nonrecoverable component of compressibility (and storage) is much higher than the recoverable component.

Both the Franklin and Suffolk simulations reveal another important aspect of the behavior of this system related to the fine-grained sediments modeled with the unit 4 separator. This separator is undergoing a somewhat complex continued decline in head across its thickness. In addition, the simulated parameter values of this unit give it a time constant of several decades, which is quite long compared to the other units. As a result, compaction in this unit is continuing as the result of past head declines, and this compaction appears likely to continue into the future, even if water levels rise in the aquifers above and below. Though future scenarios described here were not simulated, increases in hydraulic heads in the adjacent aquifers would likely take many years to equilibrate through this unit and arrest compaction.

Model Conclusions

In general, results from simulation of compaction in the aquifer systems at both Franklin and Suffolk produced reasonable and geologically defensible parameter values for the hydrogeologic units of interest. These values are presented in table 5-3.

Specifically, S'_{skv} for all of the fine-grained units is small but significant, generally decreasing with depth and ranging from 1.0×10^{-4} to 5.2×10^{-5} m^{-1}. Values for S'_{skv} for the interbeds and confining units of marine origin (model units 2, 3, and 4) are approximately an order of magnitude higher than values for the continental deposits (model units 5 through 9), perhaps reflecting less preconsolidation of the younger and less deeply buried marine sediments.

Simulated values of S'_{ske} range from 4.5×10^{-6} m^{-1} to 6.0×10^{-6} m^{-1}, but greater uncertainty is associated with the calibration of S'_{ske} because heads have not recovered enough to cause substantial elastic expansion of confining units and allow accurate measurement of this component in the field. The storage values simulated here are relatively small compared to those measured in other environments experiencing larger rates of aquifer-system compaction, but they are one to two orders of magnitude larger than previously estimated for the Coastal Plain system.

Vertical hydraulic conductivity values are very low for all of the simulated confining units, ranging from 4.4×10^{-4} to 2×10^{-3} m/yr, but the simulated total compaction record is not extremely sensitive to changes in K'_v because most of the fine-grained units are relatively thin, particularly in the Potomac system.

The accuracy of the calibrated parameter values for all units younger than Tertiary age (model units 2 and 3) is uncertain, because water levels in those units have not declined substantially. These units do not appear to have contributed significantly to total compaction in the system, but even much larger storage values and smaller conductivity values would not produce significantly different compaction records for these units under the current stress conditions. As a result, parameter values for these units are assumed to be similar to those of the other marine units in model unit 4.

Simulation results indicate that the Potomac system of Cretaceous age, represented by model units 5, 6, 7, 8, and 9, has contributed about half (45 to 65 percent, depending on location) of the cumulative historical compaction. In recent years, compaction in the Cretaceous part of the aquifer system at Franklin has ceased and some expansion has been observed as heads in these units have either leveled off or risen. Compaction in the Cretaceous part of the system at Suffolk, however, has continued to increase due to continued head declines in this area.

Heads throughout the Potomac aquifers now approach or exceed the threshold of past maximum stress at both Franklin and Suffolk, indicating that even small declines in head beyond current levels could cause greater compaction of the Potomac system and larger rates of subsequent land subsidence.

Notably, the fine-grained marine sediments (model unit 4) overlying the Cretaceous Potomac system appear to be responsible for a large percentage (30 to 40 percent, depending on location) of the historical compaction and much of the ongoing compaction. Simulation results indicate that this unit is compacting slowly as withdrawals from the Potomac aquifers maintain a large head gradient, causing downward drainage of pore water through this interval. If correct, the simulation for model unit 4 indicates that small rates of residual compaction will continue for months or years even if hydraulic heads in the Potomac system do not decline beyond current levels.

Modeling efforts for both locations were reasonably successful in simulating the measured compaction records, though the Suffolk simulation was considerably less satisfying. The lack of agreement between the measured and simulated compaction records at Suffolk are thought to be likely due to (1) conceptual problems related to modeling assumptions regarding water levels, and (2) measurement errors associated with the operation and precision of the Suffolk extensometer. Based on observations outlined here and further efforts to produce an independent calibration for the Suffolk model, there is no evidence that unit parameter values of K'_v, S'_{skv}, S'_{ske}, and S_{ske} are significantly different at Franklin and Suffolk.

The success of the compaction simulations at Franklin and Suffolk reveals important information about the behavior of the Coastal Plain aquifer system in Virginia and suggests that this model could be used to predict the compaction response of the system to future changes in head. In general, compaction is occurring in the non-recoverable regime in both the Potomac system and in the confining unit made up of marine sediments of Tertiary age overlying the Potomac. Compaction in younger sediments appears to be insignificant, though it is not known whether this is due to lower compressibility and storage values for these sediments or the fact that hydraulic heads have not declined significantly in these units.

The simulation of compaction at Franklin and Suffolk indicates that parameter values for specific storage and hydraulic conductivity are probably very similar at the two locations, and differences in compaction rates are primarily the result of differences in the hydraulic head records and in unit thicknesses. The parameter values generated by this modeling effort are expected to be useful in future efforts to refine and improve the three-dimensional regional groundwater flow model of the Coastal Plain aquifer system in Virginia. The current model does not explicitly simulate confining units or the release of water from confining unit storage, and the results of this one-dimensional modeling suggest that the confining units may play a more significant role in the flow system than previously thought by releasing water slowly from storage.

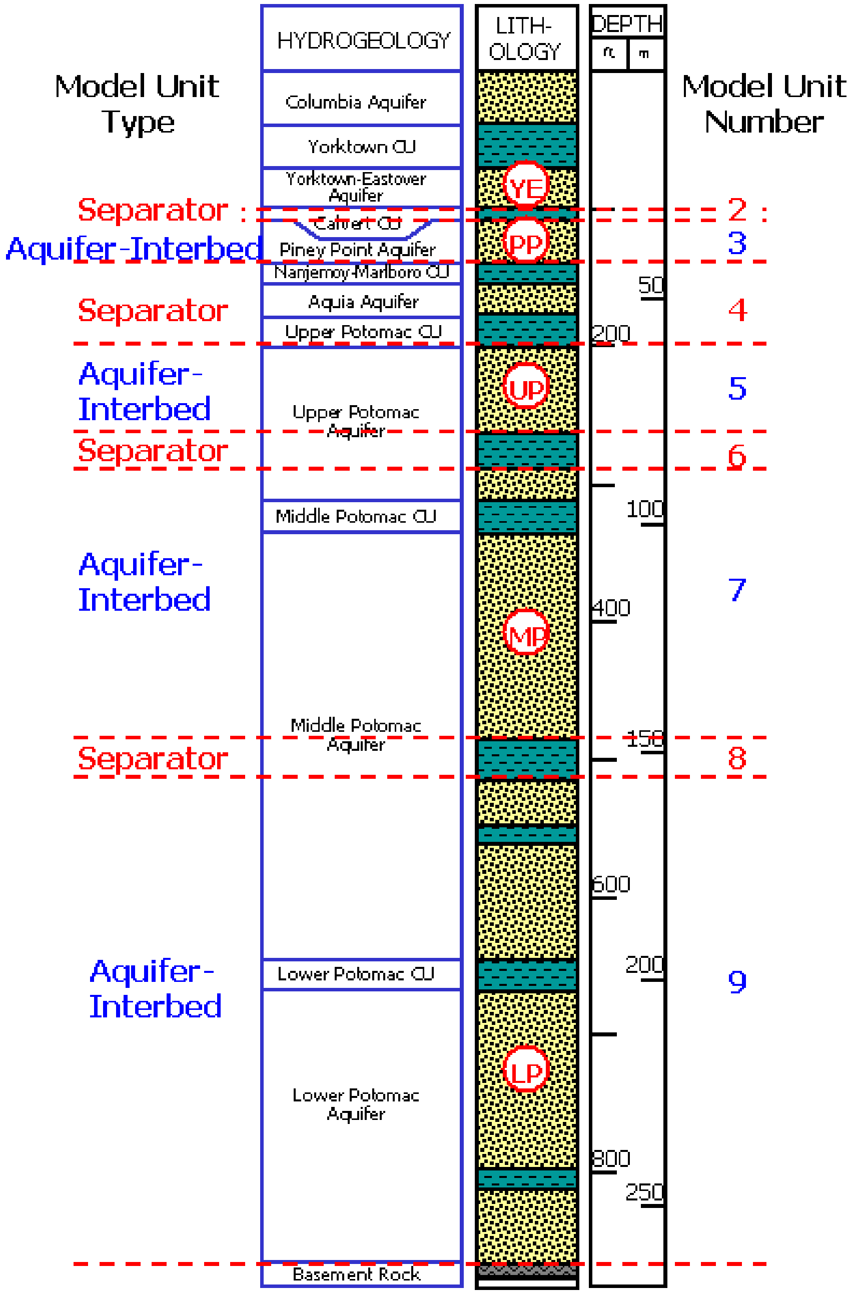


Figure 5-1: Discretization of the one-dimensional model for Franklin.

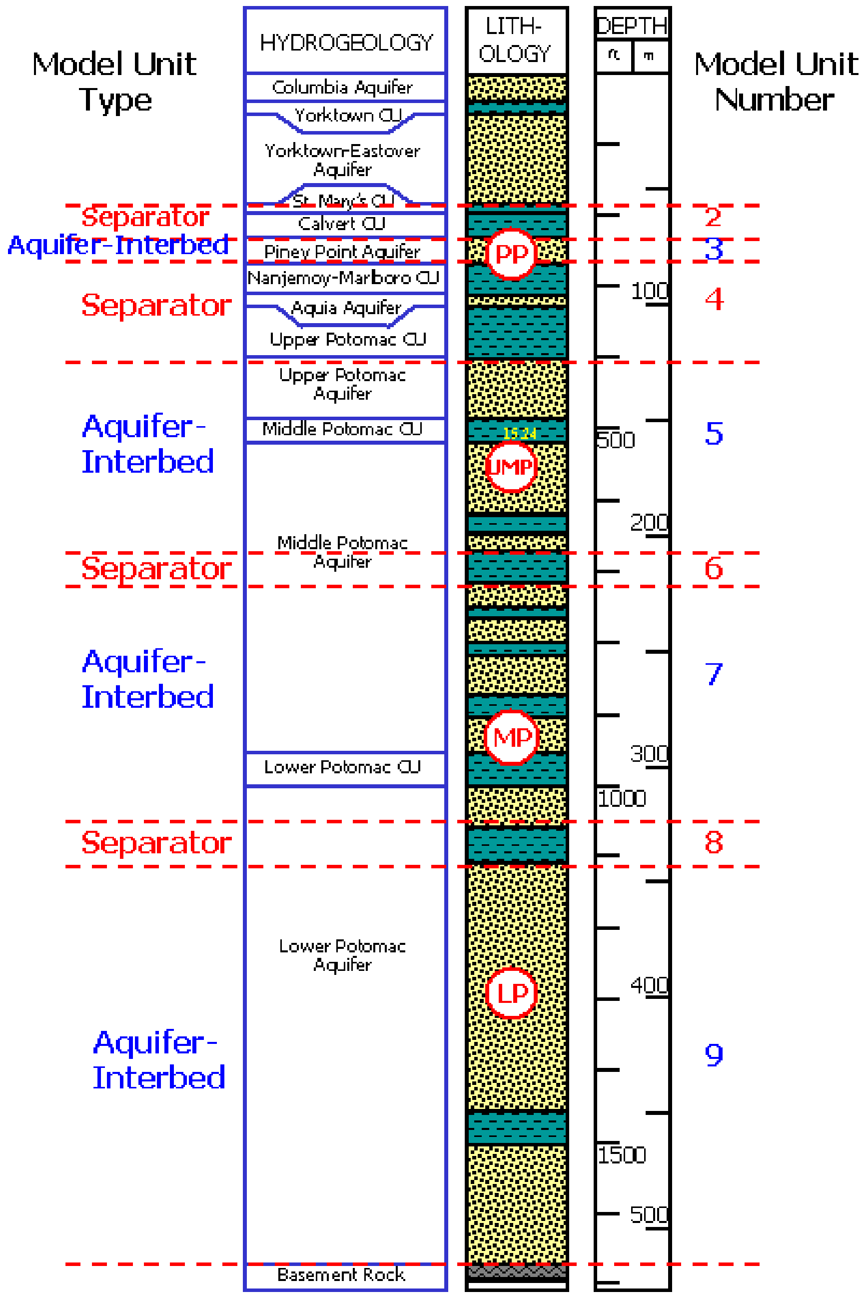


Figure 5-2: Discretization of the one-dimensional model for Suffolk.

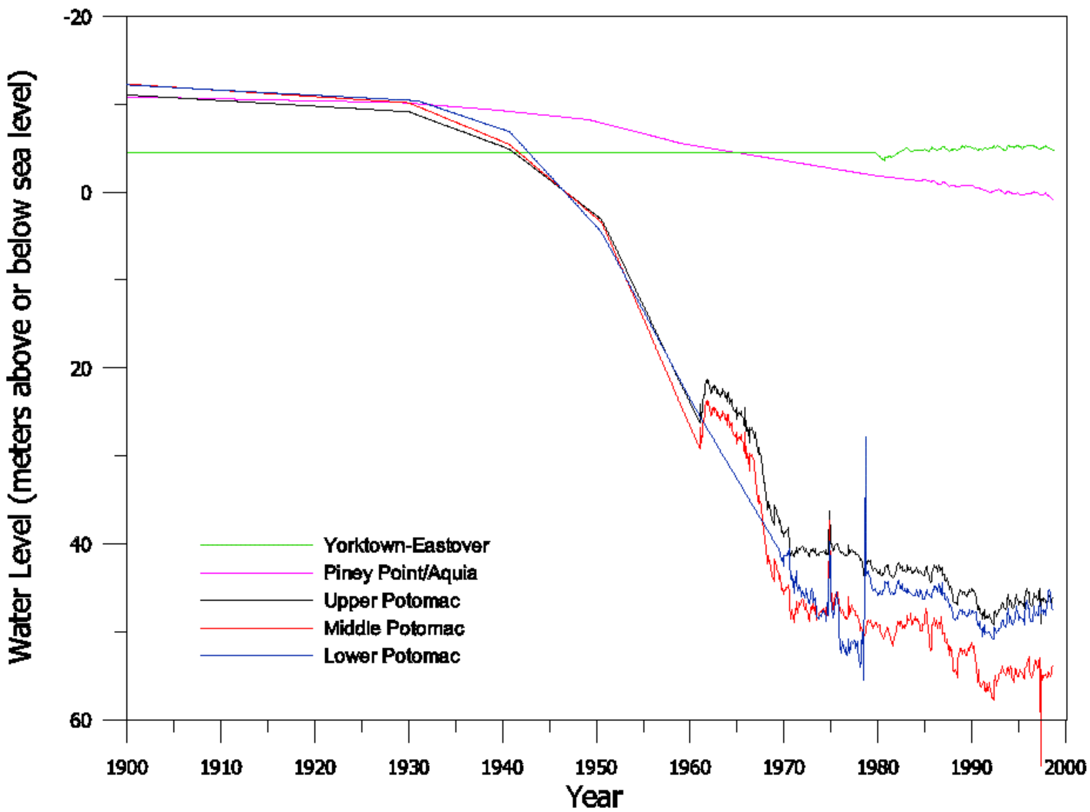


Figure 5-3: Synthetic water level records from Franklin area used as input stress files for simulation.

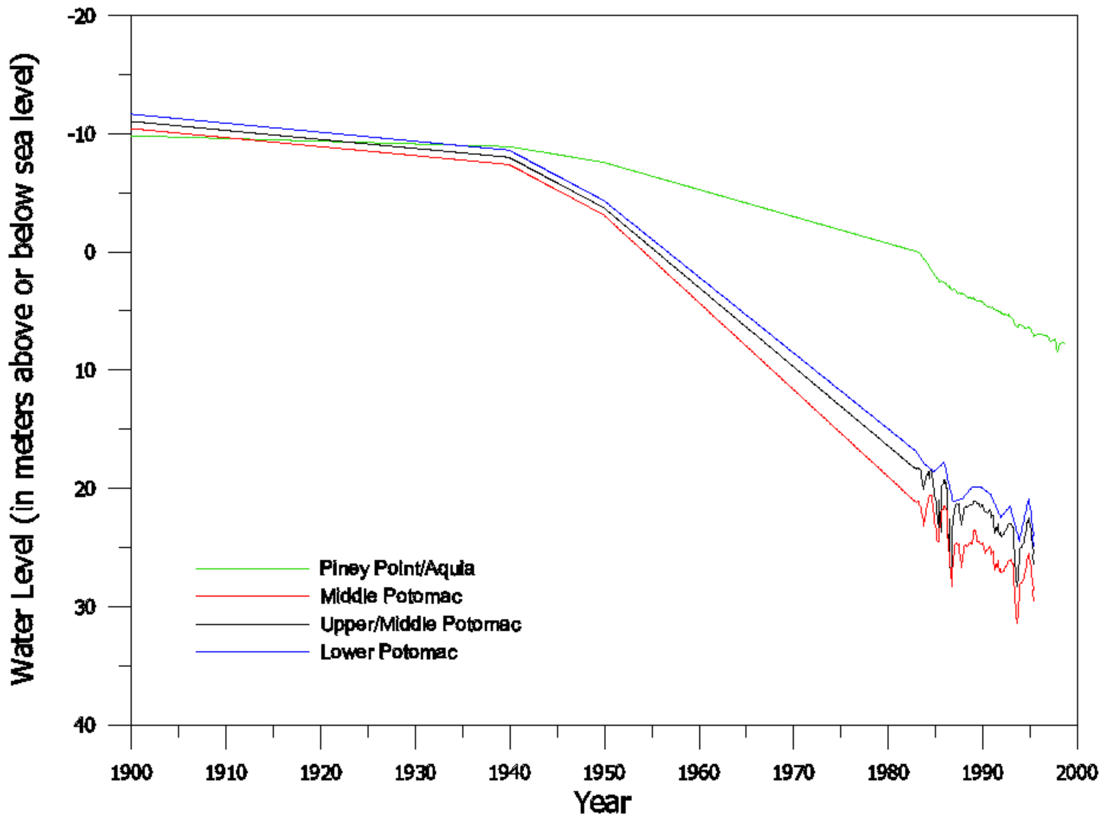
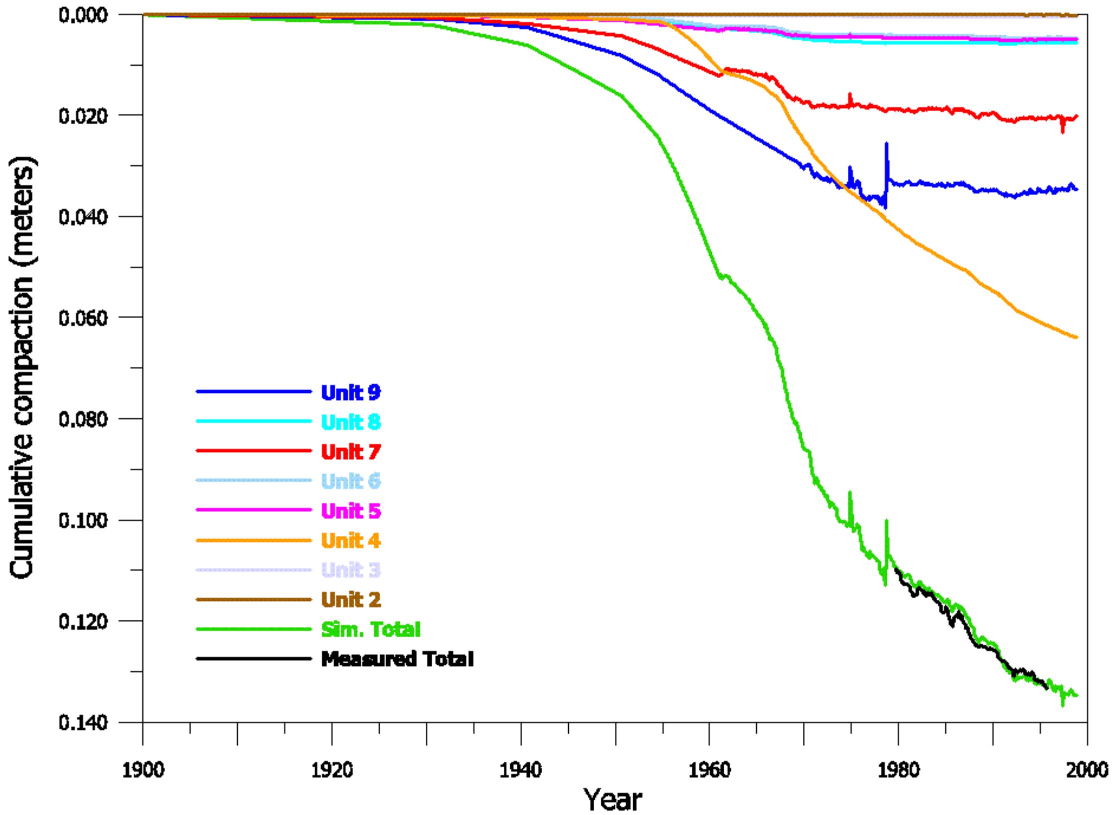


Figure 5-4: Synthetic water level records from Suffolk area used as input stress files for simulation.mulated and measured aquifer-system compaction at Franklin.



0.000
0.020
0.040
0.060
0.080
0.100
0.120
0.140
Cumulative compaction (meters)
1900
1920
1940
1960
1980
2000
Year
Unit 9
Unit 8
Unit 7
Unit 6
Unit 5
Unit 4
Unit 3
Unit 2
Sim. Total
Measured Total

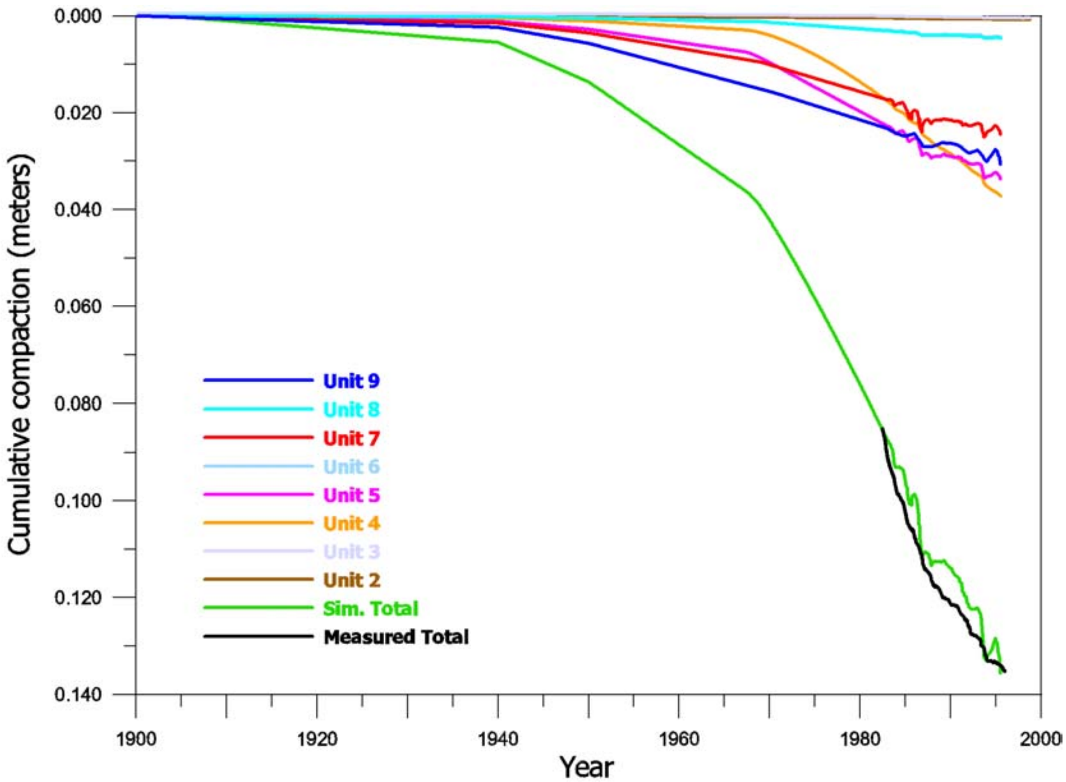


Figure 5-5: Simulated and measured aquifer-system compaction at Franklin.

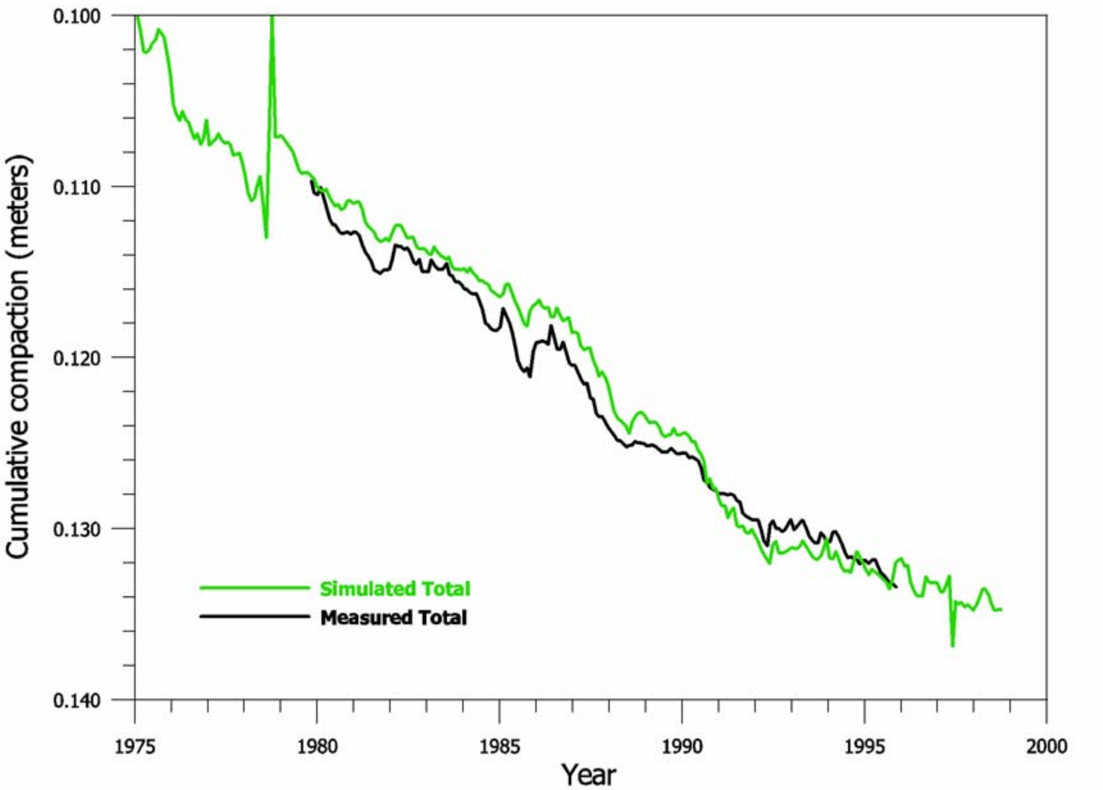


Figure 5-5a: Close-up of simulated and measured aquifer-system compaction at Franklin for period of measured extensometer record.

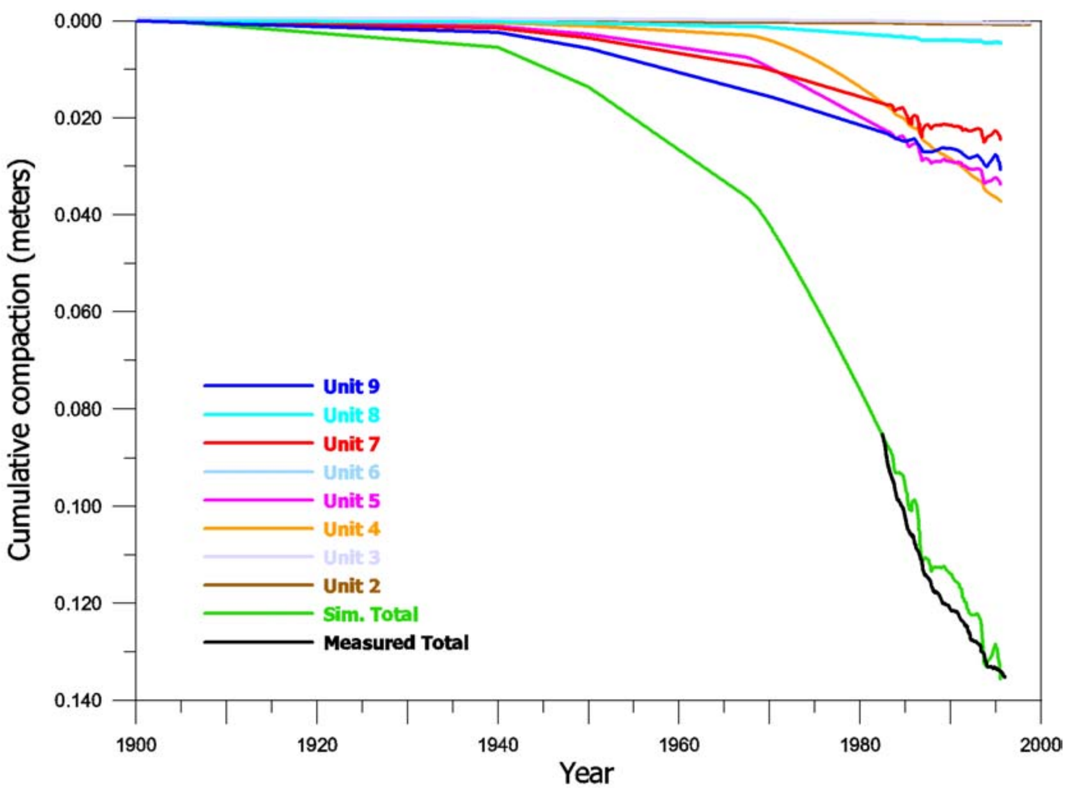


Figure 5-6: Simulated and measured aquifer-system compaction at Suffolk.

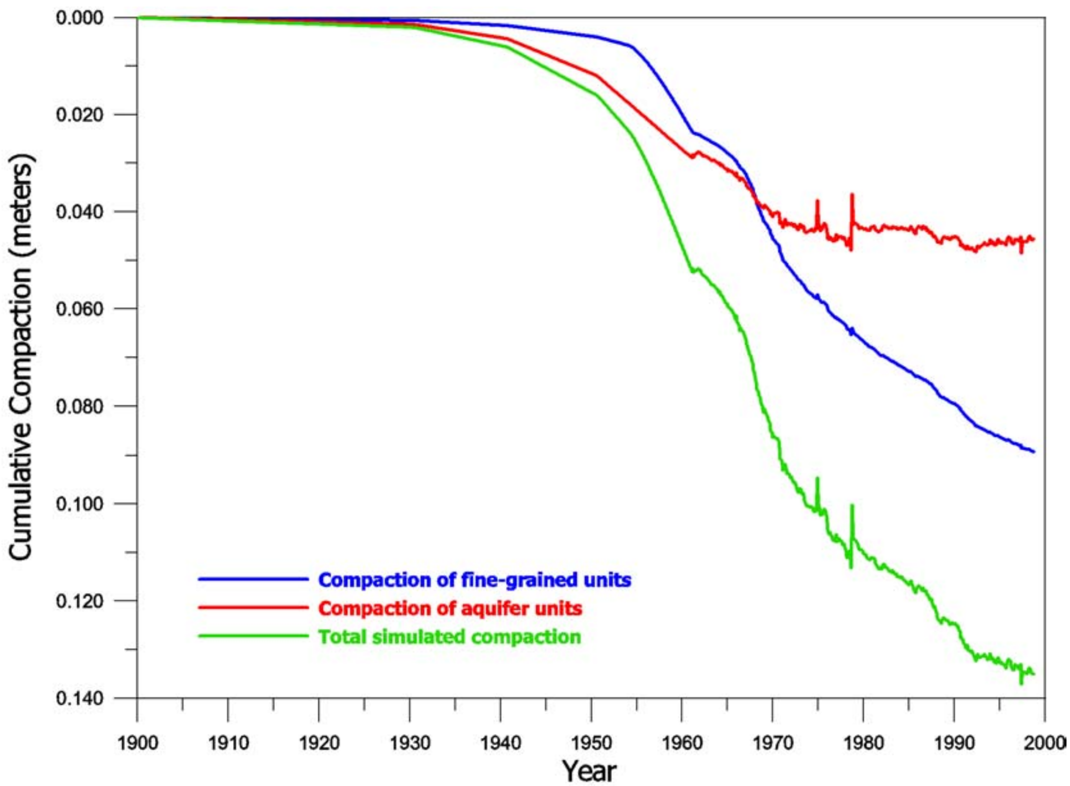


Figure 5-7: Comparison of simulated compaction at Franklin for fine-grained confining units and aquitards (blue) and coarse-grained aquifer units (red). Total simulated compaction record is green.

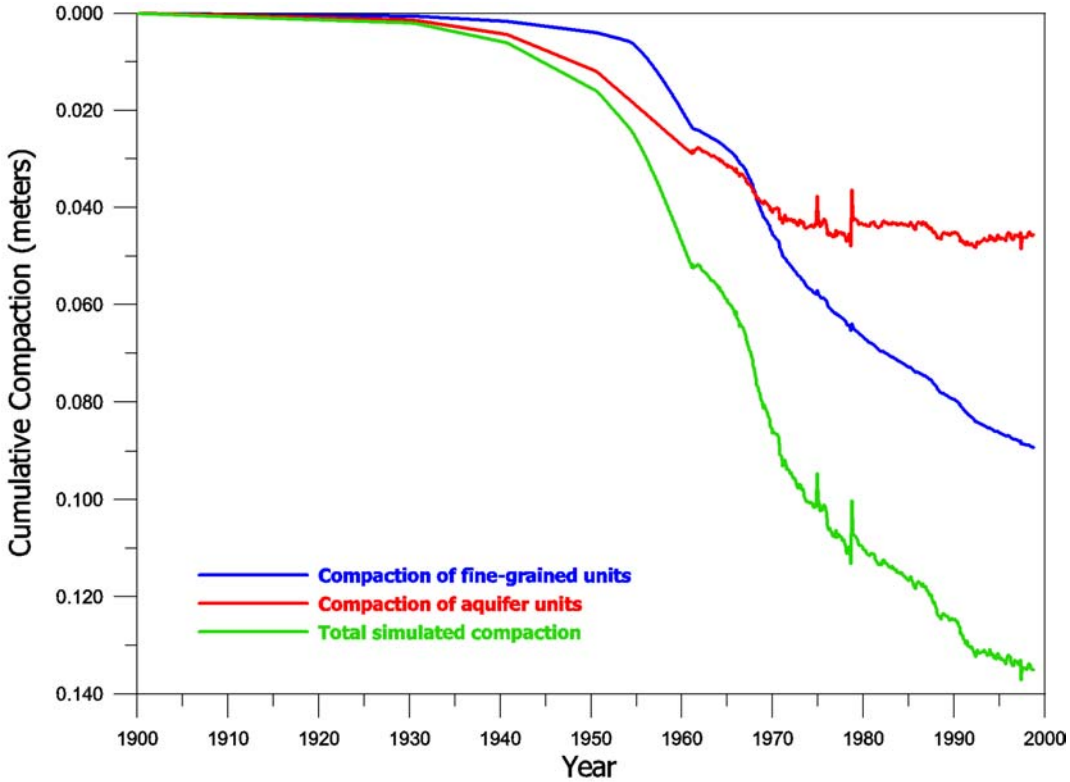


Figure 5-8: Comparison of simulated compaction at Suffolk for fine-grained confining units and aquitards (blue) and coarse-grained aquifer units (red). Total simulated compaction record is green.

VI. LAND SUBSIDENCE AND RELATIVE SEA-LEVEL RISE

The constraints of one-dimensional modeling limit the direct application of the model results to the locations explicitly simulated, in this case Franklin and Suffolk. The application of the parameters developed in this one-dimensional simulation to a three-dimensional groundwater flow model is necessary to accurately predict aquifer-system compaction and the resulting land subsidence across a large area like the Virginia Coastal Plain. Nonetheless, some general predictions and calculations can be made in the absence of such a model, given the relative consistency of the geologic units across the region. It should be noted, however, that the thickness of these units varies considerably across the Coastal Plain, and this change in thickness can significantly influence the computation of compaction and subsidence, as the results from Franklin and Suffolk revealed.

If it is assumed that the modeled value for preconsolidation stress is approximately correct, then substantial subsidence may have occurred at any location where head decline in the Potomac aquifer system has exceeded 20 meters from predevelopment conditions. Referring back to the map of the potentiometric surface for the Middle Potomac aquifer in 1998 (figure 1-2), we can estimate the extent of this stress condition with the aid of geographic information systems techniques. This relatively crude calculation reveals that the area of eastern Virginia affected by subsidence resulting from groundwater withdrawals extends from the North Carolina border to the middle of the Middle Peninsula and from the Chesapeake Bay almost to the Fall Line (figure 6-1).

Furthermore, observations of changes in the potentiometric surface over several decades indicate that the regional cone of depression depicted in figure 1-2 is expanding northward (as populations grow in the counties of Virginia's Middle Peninsula and Northern Neck) and eastward (as communities along the Chesapeake Bay begin to pump and treat more brackish water from the Potomac system in those locations). The result of these changes is that the area affected by land subsidence is likely to grow, and rates of subsidence might reasonably be expected to increase in some locations, particularly along the western shore of the Chesapeake Bay.

Given the small rates of land subsidence observed in eastern Virginia, small increases in those rates may have relatively insignificant consequences, but any consequences are likely to be most severe in low-lying coastal areas subject to the effects of rising sea levels. The analysis of sea level rise is extremely complicated and well beyond the scope of this document, but it is interesting to note that the rate of relative sea level rise (measured from long-term staff gauges) has been observed to be anomalously high in southeastern Virginia, just a few miles from Suffolk. In fact, rates of relative sea-level rise at several locations in southeastern Virginia have been observed to be much higher than the approximately 2 mm/year average for the Atlantic Coast and elsewhere in the Chesapeake Bay over the period 1950-1993 (http://co-ops.nos.noaa.gov/seatrnds.html), (figure 6-2). The highest rate of relative sea-level rise in the region over this period, 4.1 mm/year, was measured at Portsmouth, Virginia. This rate of relative sea-level rise at Portsmouth is approximately equal to the sum of the measured compaction rate near Portsmouth and the average rate of sea-level rise along the Atlantic coast.

Land subsidence has been observed to have the most significant cultural and environmental effects in coastal areas where it magnifies the effects of rising sea levels. Even at these relatively low rates, land subsidence can have serious effects. For example, reports from the Virginia Institute of Marine Science (VIMS) have indicated higher than normal rates of erosion of tidal marshes in this area, with associated habit loss (Carl Hershner, VIMS, oral communication, 2001). It may be too early to attribute these losses to land subsidence, but further studies are underway.

As part of the Chesapeake Bay program conducted by the National Oceanic and Atmospheric Administration (NOAA), a number of long-term global positioning system (GPS) stations have been established in recent years to monitor land-surface movements in the bay area (http://www.ngs.noaa.gov/GRD/GPS/Projects/CB/MAP/map.html). These GPS stations are currently collecting elevation data in order to establish long-term trends, though land-surface movements of just a few millimeters per year are currently below the resolution of the instruments. Nonetheless, GPS techniques provide perhaps the best way to monitor land-surface movements over this region. Concurrently, continued direct measurements of aquifer-system compaction at the Franklin and Suffolk extensometer sites will enable comparison to land-surface movements derived from remote sensing techniques, resulting in a better understanding of the environmental changes resulting from groundwater withdrawals in eastern Virginia, Maryland and North Carolina.

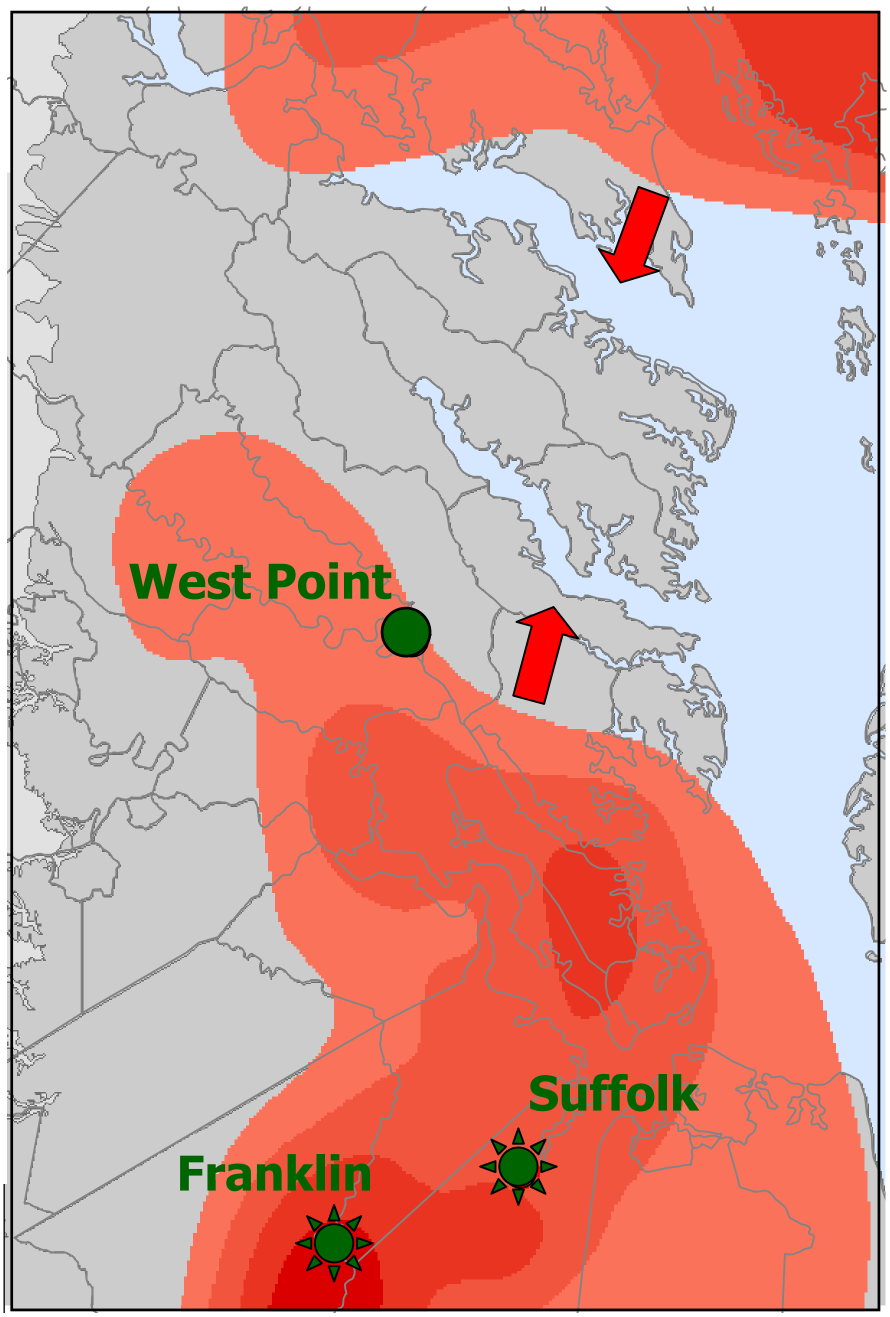


Figure 6-1: Map of estimated area affected by measurable land subsidence due to ground water withdrawals. Shaded are depicts region of greater than 20-meter head decline from predevelopment conditions, which should be associated with land subsidence.

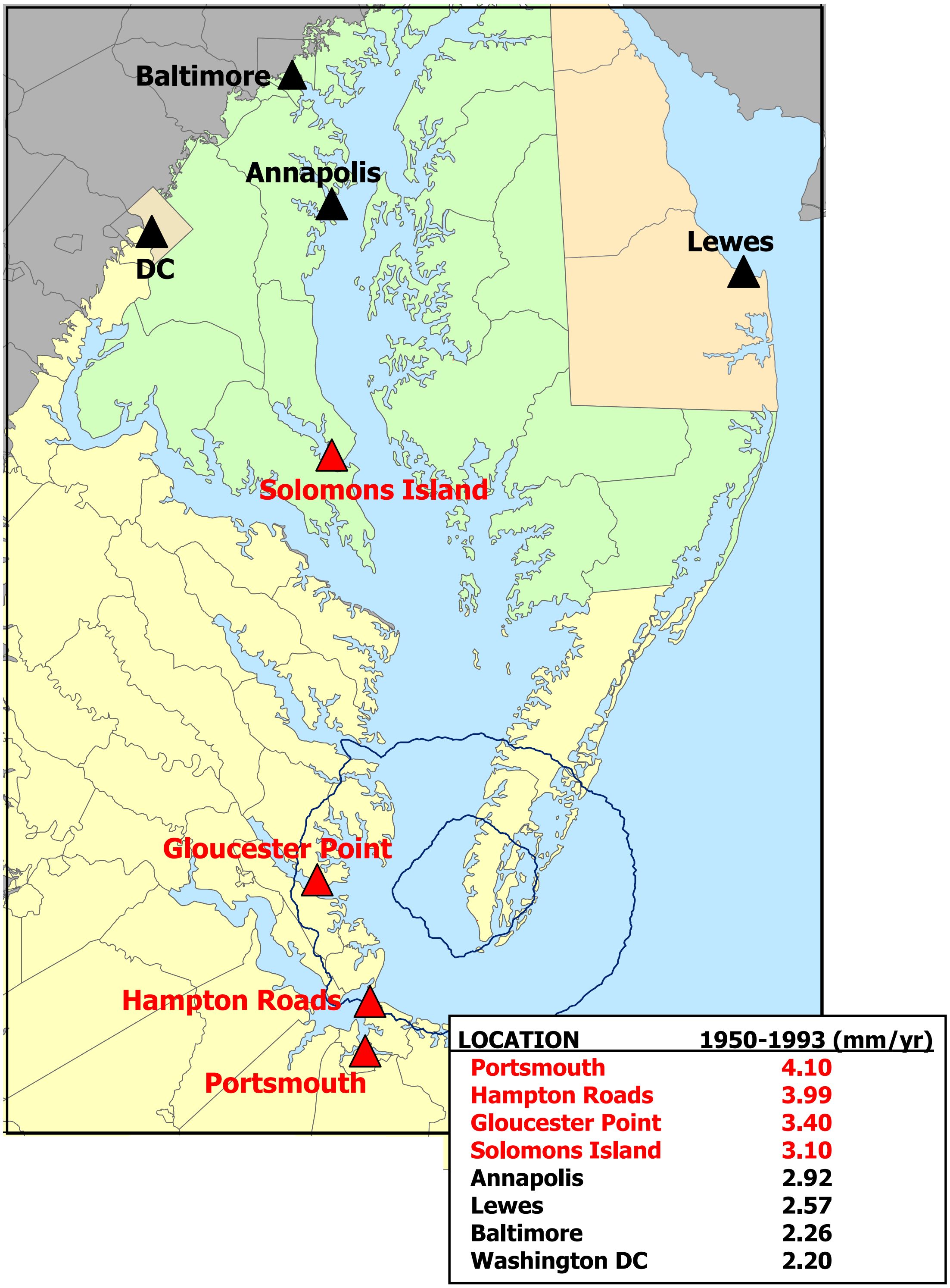


Figure 6-2: Mean rates of relative sea level rise (mm/year) in the Chesapeake Bay area from 1950 through 1993.

VII. SUMMARY

Over the past century, large increases in ground-water withdrawals from the confined aquifers of the Virginia Coastal Plain have resulted in large declines in hydraulic heads and measurable amounts of subsequent land subsidence. Measurement and simulation of aquifer-system compaction indicate that the magnitude and rate of compaction and the resulting land subsidence in this region are small when compared with many other areas experiencing subsidence. Nonetheless, analyses of extensometer measurements have confirmed that measured compaction is the result of ground-water withdrawals from the confined aquifers, and simulations have increased our understanding of the complex relation between measured hydraulic heads and compaction in this system. Simulation results explain the uneven spatial and temporal distribution of compaction in the aquifers and confining units of the Coastal Plain. In addition, simulations have revealed that storage values for the fine-grained units in the Coastal Plain are larger that previously thought, suggesting that confining-unit storage may substantially affect ground-water flow in this complex aquifer system. These results are expected to be useful in ongoing efforts to simulate ground-water flow. Previous flow models of the Virginia Coastal Plain have not considered the effects of confining-unit storage, which may be important for accurate calibration of hydraulic heads, particularly for transient simulations. Furthermore, measurements and simulations of compaction may be useful for predicting and managing land subsidence in coastal environments.

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