ESTIMATING THE COMPONENTS OF A WETLAND WATER BUDGET

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Key words: wetland hydrology, water budget, water balance, precipitation, evapotranspiration, runoff, stream flow, groundwater, water storage

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by

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(ABSTRACT)

The design of wetlands to replace those lost to development requires quantitative understanding of the wetland water budget in order to estimate the amount of water available to the wetland over time. Many methods exist to estimate each component of the wetland water budget. In this study, monthly values of the water budget components namely, precipitation, runoff, evapotranspiration, and groundwater seepage were calculated using a water budget model and compared to on-site field measurements for a wetland in Manassas, Prince William County, VA. The monthly precipitation estimated from a weather station 32.18 km from the site differed from the on-site values by as much as 2.9 times. Runoff estimates calculated by the Soil Conservation Service (SCS) method using antecedent moisture condition (AMC) II underpredicted runoff for every month by as high as 100 percent compared to the on-site measured runoff. The choice of AMC greatly affected the SCS runoff estimates. Runoff was the dominant water budget component at the Manassas wetland. The evapotranspiration (ET) estimates using the Thornthwaite method either over or underestimated ET when compared to ET calculated from diurnal cycles of the water table in the wetland. Groundwater seepage losses were calculated using Darcy's equation with an assumed hydraulic gradient of one, and with gradients measured with nested piezometers. Seepage losses at the Manassas wetland were negligible. Overall, the water budget model provided conservative estimates of the available water in the wetland during the 10-month period of observation.
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1. INTRODUCTION

Wetlands are land areas covered by shallow water or saturated within the root zone, with hydric soil conditions, and supportive of hydrophytic vegetation but not flooding-intolerant vegetation (Mitsch and Gosselink, 1993). They may have important and useful environmental functions. Some wetlands moderate floods, remove pollutants and sediments from inflows, and have aesthetic values. They include some of the most productive ecosystems on earth. Many valuable species of fish, fur animals, waterfowl, and trees are wetland dependent (Tiner, 1984; Mitsch and Gosselink, 1993).

Wetlands make up about 6% of the world's land surface (Bazilevich et al., 1971; Maltby and Turner, 1983). Around 1985, it was estimated that 53% of wetlands in the continental U.S. had been destroyed since the late eighteenth century (Dahl, 1990). The U.S. Fish and Wildlife Service estimated that by the mid-1980s, 103 million acres of wetlands remained in the U.S. (Frayer et al. 1983). There is a national consensus that these remaining wetlands need protection and their complex functions need study (Mitsch and Gosselink, 1993).

The national policy of no net loss of wetlands provides the framework for the preservation of the nation's remaining wetlands. Its purpose is to achieve no overall net loss of the nation's remaining wetlands, create and restore wetlands, and increase the quantity and quality of wetlands (National Wetlands Policy Forum, 1988). Wetland mitigation accomplishes this by compensating for adverse impacts to wetlands often caused by development such as highway construction, coastal drainage and filling, and commercial development. Many early mitigation wetlands failed because they did not replicate the wetland ecosystem form and functions. The major problems were associated with an incomplete understanding of wetland hydrology resulting in water levels that did not support wetland functions (Mitsch and Gosselink, 1993).

The hydrology of wetlands establishes and maintains specific types of wetlands and their associated processes (Mitsch and Gosselink, 1993; Kusler, 1987). Hydrology determines the origin, location, extent, duration and frequency of flooding, and the flora and fauna in a wetland. In turn, these factors determine wetland functions, sensitivity to environmental perturbation, and chemical and physical properties, such as nutrient status and availability,
salinity, redox potential, and pH of wetland soil sediments (Carter, 1986; Kusler, 1987; Mitsch and Gosselink, 1993). Therefore, successful wetland mitigation depends primarily on a quantitative understanding of the wetland hydrological budget over time. Without appropriate hydrology, other important factors, such as pH, organic matter, species selection, microtopography, soil compaction, and subsoil conditions on their own or in combination are not enough to sustain wetlands. Studying the wetland hydrology requires an accurate estimate of each component of the wetland water budget. However, there are no standard practices or methods for calculating the components of a water budget for wetland mitigation. The approval of a wetland mitigation project plan depends primarily on state and federal environmental regulators. Therefore, state guidelines for wetland mitigation differ not only from state to state but from area to area within a state.

The Army Corps of Engineers (ACOE) and the Environmental Protection Agency (EPA) have developed a memorandum of agreement (MOA) concerning wetland mitigation under the guidelines of the Clean Water Act, Section 404(b)(1). The MOA requires appropriate, practical, and feasible compensatory mitigation for unavoidable adverse impacts on wetlands after all other appropriate and practicable avoidance and minimization measures have been taken. Compensatory mitigation implies either the restoration of adversely impacted wetlands or the compensatory creation of wetlands off-site. However, the MOA does not specify a method to calculate the water budget for wetland mitigation sites. The water budget is essential to help predict future hydrologic conditions and to ensure that reconstructed soil wetness conditions are appropriate for the intended vegetation community. Private consultants therefore use their own proprietary methods in wetland mitigation, and typically do not make them public for competitive reasons. Individual ACOE districts are trying to establish methods for calculating a wetland water budget to establish minimum standards for permit review and to allow some uniformity in comparing designs.

The Norfolk ACOE district has prepared an internal document that explains how to estimate a water budget (Westbrook, 1994). This document closely follows the guidelines and procedures recommended by Pierce (1993) for calculating a wetland water budget. Since 1986, Dr. Gary Pierce has been an instructor for the ACOE Waterways Experiment Station
teaching "Wetland Development and Restoration" workshops across the country. Pierce's guidelines are the basis of the water budget model recommended for use within the ACOE. However, the Norfolk ACOE water budget is less detailed than Pierce's (1993) approach. For the water balance comparison in this report, the Westbrook (1994) method will be referred to as the modified Pierce model, but it is important to note that it is not specifically endorsed by the ACOE for national use, and is really a set of recommended procedures for use within the Norfolk District.

In wetland construction, it is essential to estimate the amount of water available to the wetland and the fluctuations of the wetland water levels over time. The modified Pierce model accomplishes this by calculating the potential storage of water as the difference between the volumes of inputs and any outputs of water not controlled by an outflow structure. Inputs include precipitation, surface runoff, and groundwater inflow. However, to underestimate the total inflow, groundwater inflow is set to zero. Outputs consist of evapotranspiration, downward groundwater seepage, and surface water outflow. However, surface water outflow can be controlled by an outflow structure and is therefore not included in the water budget. The modified Pierce model assumes that an outflow structure, such as a spillway, will be used to control a maximum water level specified in the wetland design. The actual storage of water in the wetland is then expressed as the levels of the free water surface in the wetland over time. Following Pierce (1993), precipitation data is obtained from a nearby weather station, potential evapotranspiration is calculated using the Thornthwaite (1948) method, surface runoff values are determined using the SCS runoff method, and groundwater seepage is calculated using Darcy's equation. The ACOE simplifies Pierce's (1993) model by setting the hydraulic gradient equal to one in groundwater estimates using Darcy's equation.

Although accuracy associated with the modified Pierce model for estimating wetland water budgets is unknown, it is commonly used for wetland designs which could cost millions of dollars once installed. A comparison between the modified Pierce model and actual field measured wetland water budget components is needed to estimate the magnitude of error associated with each water budget component, and to evaluate the accuracy of the overall
model approach to determining wetland water balances.

The purpose of this study was to make this comparison. The specific objectives of this study were:

1. To calculate water budget components using the methods described by the modified Pierce water budget model.

2. To compare the modified Pierce water budget components and potential storage with measured field data.
2. LITERATURE REVIEW
2.1 Characteristics and Functions of Wetlands

The term wetland is new to many people, although they may be familiar with swamps, bogs, marshes, mires, bottomlands, prairie potholes, or wet meadows all of which are different types of wetlands. Wetlands are often between well-drained upland and permanently flooded deep aquatic systems (Mitsch and Gosselink, 1993; Tiner, 1984). They are also found in depressions or along rivers, lakes, and coastal waters, and exist there because of periodic flooding (Tiner, 1984). Characteristics of some wetlands are standing water, saturated soils, anoxic conditions, and high biodiversity with species from both aquatic and upland systems (Mitsch and Gosselink, 1993). The most common feature that most wetlands share is soil or substrate that is at a minimum periodically saturated with or covered by water. The depth, length, and timing of water saturation determines the soil development and the types of plant and animal communities (Cowardin et al., 1979), and is commonly referred to as the wetland hydroperiod.

Defining wetlands and associated boundaries has proven to be a difficult task for wetland scientists and regulators because of wetland diversity. It is difficult to impose an artificial boundary on an area that lies along a continuum between dry and wet environments (Cowardin et al., 1979). In 1956, the U.S. Fish and Wildlife Service published the earliest definition of a wetland in a document referred to as Circular 39 (Shaw and Fredine, 1956). In 1979, they published another document called Classification of Wetlands and Deepwater Habitats of the United States (Cowardin et al., 1979). Today, wetland regulators follow the U.S. government regulatory definition of a wetland which is governed by the U.S. Army Corps of Engineers under the implementation of a dredge-and-fill permit system required by Section 404 of the 1977 Clean Water Act Amendments. Although the two definitions vary, they all recognize hydrology, wetland vegetation, and hydric soils as the three key attributes for defining wetlands (Shaw and Fredine, 1956; Cowardin et al., 1979; Tiner, 1984; Mitsch and Gosselink, 1993).

In general, wetlands have many values and functions for fish and wildlife, the environment, and society. Some wetlands maintain the fish and shellfish populations both inland and on the
coast. The same is true for waterfowl and bird habitats. Wetlands can serve as year-round habitats for resident birds and breeding grounds, overwintering areas, and feeding grounds for migratory birds. Furbearers and other wildlife, such as rabbits, deer, turtles, reptiles and amphibians are also wetland dependent (Tiner, 1984).

Environmental quality can be improved through aquatic productivity and water quality. Many wetland types are some of the most productive ecosystems on Earth, supplying food which supports aquatic and terrestrial life. Water quality is improved because wetlands remove and recycle nutrients, filter pollution, remove sediments, produce oxygen, and absorb chemicals. Wetlands temporarily store flood waters and help to slow the velocity of water thereby lowering peak flow heights and reducing soil erosion. They sometimes recharge aquifers that are sources of drinking water. These are all indirect benefits to society; others are more direct. Wetlands also provide an income for many people through the production of timber, peat, cranberries, blueberries, rice and grasses for livestock (Tiner, 1984). Many cultures, such as the Babylonians and Egyptians have benefited economically from wetlands since early civilization (Mitsch and Gosselink, 1993). People can also enjoy the recreational and aesthetic benefits of wetlands (Tiner, 1984).

Since the late eighteenth century wetlands have been drained and filled destroying the values and functions previously discussed (Dahl, 1990). Studying wetland hydrology is an important first step in understanding how wetlands function and to help us protect, restore, and recreate these important ecosystems for ourselves and future generations.

2.2 Components of a Wetland Water Budget

Wetland hydrology is best quantified by means of a water budget. A wetland water budget follows the principle of conservation of mass and represents a budget of inputs, outputs, and storage of water in the wetland. Quantifying these components provides the basis for effective wetland management in water resources planning and in agricultural and engineering applications (Gilman, 1994). A wetland water budget for a specified period can be expressed as: \( P + SWI + GWI = ET + SWO + GWO \pm \Delta S \), in which \( P \) is net precipitation, \( SWI \) is surface water inflow, \( GWI \) is groundwater inflow, \( ET \) is evapotranspiration, \( SWO \) is
surface water outflow, GWO is groundwater outflow, and ΔS is change in storage (positive if increase and negative if decrease). Each component can be expressed as a depth per unit area or as a volume (Carter, 1986; Kusler, 1987; Mitsch and Gosselink, 1993). In some wetland studies it is meaningful to calculate just one component of the water budget, such as the case when Ewel and Smith (1992) measured the ET rates of pond cypress trees (Taxodium distichum var. nutans) in three swamps by measuring the diurnal changes in water levels in central Florida. In other studies, it is necessary to calculate all water budget components (Carter, 1986). Rushton (1996) measured rates of evapotranspiration and net seepage (ΔGW=groundwater inflow minus outflow) from a freshwater marsh in Florida and compared the results to a mass water budget as described above, i.e., ΔGW+ET=P+SWI-SWO±ΔS.

The relative magnitude of water budget components varies depending on wetland function, location, climate, and water source. These components can either be measured in the field or estimated. Field measurements can be expensive and time consuming, depending on which component is being measured. Rainfall and surface water levels are easy and inexpensive to measure compared to groundwater seepage and evapotranspiration which are more difficult to measure directly (Carter, 1986; Rushton, 1996). It is often more feasible to estimate such components (instead of field measurements) because of the simple procedures and low cost associated with these estimates.

The accuracy of the overall wetland water budget depends on the magnitude of the errors associated with the measurements or estimates of the individual components (Carter 1986). Winter (1981) studied the uncertainties in estimating the water budget of lakes and found that associated errors are commonly not analyzed fully or are overlooked. It is a common practice to measure or estimate all but one component, and to estimate the unmeasured component as the residual of the water budget equation. When groundwater or evapotranspiration is used as the residual, all errors associated with each of the other components of the water budget are included in this residual component (Carter, 1986; Rushton, 1996). Without an idea of the errors associated with each component of the water budget, the residual term has no meaning (Winter, 1981). Another factor associated with the accuracy of the overall water budget is the length of the period over which it is calculated.
Winter (1981) concluded that the time factor influences the error of both estimated and measured water budget components. Errors tend to decrease with increasing time intervals because there is more time for the errors to balance. Rushton (1996) found the water budget for a freshwater marsh in Florida to balance with an error of 2 percent on a yearly basis, with errors less than 7 percent on a seasonal basis, and as great as 30 percent on a monthly basis. Percentages in error also need to be kept in perspective because a small percent error of a large volume of water can be a greater source of error in the overall water budget than a large percent error of a small volume of water (Winter, 1981). Each component of a wetland water budget is discussed below.

2.2.1 Precipitation

The precipitation component of the wetland water budget can be measured on site using a rain gage or estimated from a nearby weather station. Types of gages are non-recording, such as dip stick gages, and recording, such as tipping bucket gages. Errors associated with the dip stick method occur from misreading the scale, water creeping up the stick, and the water displaced by the dip stick. Tipping bucket gages have errors due to mechanical devices and inadequate calibration of instruments (Winter, 1981). For on-site recording, the study area size and gage density have a large impact on the errors associated with values for precipitation. Monthly precipitation errors can range from 10 to 20 percent decreasing to 5 percent for seasonal estimates using a gage density of one gage per 647.5 km\(^2\) (250 mi\(^2\)). Daily mean precipitation can be in error by 4 percent at a gage density of one gage per 2.59 km\(^2\) (1 mi\(^2\)) (Winter, 1981). Additional errors associated with the type and positioning of the gages are collectively termed as instrumentation error. Winter (1981) summarized that the instrumentation error for long term data ranged from 1 to 5 percent and increased to 15 percent depending on the gage height above the land surface. Studies by Rodda (1968) and Neff (1977) showed that gages above ground caught 5 - 15 percent less rain than ground level gages. Rain gages with wind shields record 20 percent more rain than gages without wind shields (Winter, 1981).

The amount of precipitation caught in rain gages set in the open is less than the amount
that reaches the ground surface under the canopy. Total precipitation is made up of interception, throughfall, and stemflow. Interception is the amount of precipitation that is retained by the vegetation and throughfall is the amount of water that reaches the ground below. Stemflow is the water that passes down the stems of the vegetation and can be minor depending on the vegetation (Mitsch and Gosselink, 1993). Studies show that interception ranges from 8 to 35 percent of the total precipitation for forests and 10 to 35 percent for grasslands or croplands. Interception by emergent macrophytes is probably similar to grasslands or croplands. Interception is expected to be lower annually for herbaceous plants because of their decrease in biomass during winter months (Mitsch and Gosselink, 1993). Although interception decreases precipitation reaching the ground, it has been hypothesized that this loss is offset in the overall water budget since the intercepted water evaporates from the plant surfaces and reduces the energy available for transpiration and therefore the amount of water transpired during the rain event (Dunne and Leopold, 1978).

When using precipitation records obtained from the National Weather Service for on-site estimates, errors increase with the distance from the weather station. These errors are highest for single high intensity localized storm events and short term averages (Winter, 1981).

2.2.2 Surface Water Flow

Surface water flow occurs as channelized flow in streams and overland flow (Carter, 1986). Flow rates in stream channels, termed discharge, can be calculated by multiplying the area of the channel at a given location by the average stream velocity at a given time (Fetter, 1994). The wetted perimeter and cross sectional area changes with the height of water in the stream above a specified reference level, termed as the stage. If these measurements are made for a range of stages then a stage discharge curve can be developed relating the discharge of the stream to recorded stage. Natural phenomena, such as changes in the river bed slope and cross sectional area and backwater effects caused by a change or lack of hydraulic gradient, can alter stream conditions during rating curve measurements and create inconsistencies between measurements. However, a stage discharge curve is a regression type curve and the discharge determined from a good rating curve should be more accurate than any of the
individual measurements (Winter, 1981).

Stream flow at a given stream cross section, represents the accumulation of overland flow, interflow and base flow within the watershed contributing to the stream. Overland flow is the precipitation which does not infiltrate into the soil and flows overland into the stream. Interflow is the water which has infiltrated into the soil and discharges into a stream before reaching the groundwater table. Base flow is the water that enters the stream from groundwater (Chow, 1964; Gray, 1973). To simplify the system, stream flow can be divided into surface runoff and base flow. Surface runoff consists of overland flow and some interflow, and base flow contains groundwater and some interflow (Chow, 1964). This separation of interflow is unavoidable because both surface runoff and base flow have unknown amounts of interflow which cannot be isolated and measured. During rainfall events, the stream flow consists of both surface runoff and base flow. When the storm subsides, surface runoff decreases at a steady rate until it diminishes to zero leaving base flow as the only component of stream flow (Chow, 1964). The two components can therefore be separated if the stream flow is measured over time before and after rainfall events.

The curve obtained from plotting stream discharge versus time is called a hydrograph and the area under the curve is the total volume of stream flow. The separation of runoff and base flow components from stream flow is known as hydrograph separation. In practical hydrograph analysis, the hydrograph separation is usually made in an arbitrary manner. Although more sophisticated methods are available for hydrograph separation, the most common method extends a straight line from the point on the hydrograph where surface runoff begins (point of rise) to the end point of runoff where the baseflow recession curve begins (Wisler and Brater, 1959; Chow, 1964; Gray, 1973; Raudkivi, 1979). The straight line approach to baseflow separation is simple. The area under this line is the volume of base flow in the stream. The volume of runoff is determined by subtracting the volume of base flow from the total stream flow volume.

There are two common methods available to directly estimate runoff into streams. They are the Rational Method and the Soil Conservation Service (SCS) runoff method. The Rational method is designed to calculate maximum rates of discharge for sizing culverts and
storm sewers and does not determine the total surface runoff from an area which is needed for wetland water budgets (Walesh, 1989; Pierce, 1993).

The SCS runoff method was developed by the United States Department of Agriculture (USDA) for small watersheds (Mockus, 1972; USDA, 1973). It provides a relationship between accumulated runoff and accumulated rainfall. The SCS method requires daily rainfall and watershed data (Mockus, 1972; USDA, 1973) and the equation relates total runoff (Q) to total precipitation (P), potential maximum soil retention (S), and initial abstraction (Ia). Runoff from an area will usually begin some time after the rain has started and the retention of water before runoff starts is called the initial abstraction (Ia). The depth of rainfall after runoff begins is called the effective rainfall (Pe) and is equal to precipitation minus initial abstraction, Pe = P-Ia. The portion of the effective rainfall that does not contribute to runoff (Q) is called the retention of water in the soil (F). If the storm continues indefinitely, the runoff and effective rainfall rates become equal indicating that the soil has retained all the water it can hold meaning that F is equal to some potential maximum soil retention S. As F approaches S, and Q approaches Pe, it is assumed that F / S = Q / (P-Ia), at any time before Q = Pe = P-Ia. Since F = P-Ia-Q this becomes (P-Ia-Q) / S = Q / (P-Ia) which can be solved for Q to obtain Q = (P-Ia)^2 / (P-Ia+S) (Ogrosky, 1964; Mockus, 1972; USDA, 1973). From very sparse data from small watersheds, the empirical relation Ia = 0.2S (unit inches) was estimated from the regression of Ia plotted against S and therefore Q = (P-0.2S)^2 / (P+0.8S). Plots of Q against S would depend on the parameter S and the curves would depend on its value. The curve number (CN) was therefore defined from the relation, S = 1,000 / (CN-10), so that CN = 100 when S = 0, for CN > 0. Therefore, if S = 0, then Q = P, and there is 100% runoff. The CN is a single parameter encapsulating the land use, hydrologic soil group, and vegetative cover characteristics of the watershed. The CN for various hydrologic soil-cover complexes are provided in tables (Chow, 1964; Ogrosky, 1964; Mockus, 1972; USDA, 1973). The CN value depends in part on the antecedent soil moisture condition (AMC) which would influence the soil maximum retention (S). Three AMC conditions were defined; I for dry periods, II for average moisture periods, and III for wet periods depending on the cumulative rainfall 5 days before the storm event (Musgrave and Holtan, 1964; Mokus, 1972). The SCS runoff
method is detailed in a document referred to as TR55 (USDA, 1986) which contains tables of CN for different cover complexes based on an AMC of II. The National Engineering Handbook, Section 4, Hydrology, provides a table to convert CN for AMC II to CN for AMC conditions I and III (Mokus, 1972).

Rawls et al. (1980) showed that the accuracy of CN estimates are significantly affected by the land use classification system, but not by the hydrologic soil group and vegetative cover data. The effects of CN variation were shown to decrease as the rainfall depth increased and were highest near the beginning point of runoff (Bondelid et al., 1982). In general, the SCS method works well in practice. Steenhuis et al. (1995) found that plots of the observed runoff against effective precipitation for two watersheds in Australia and three in the northeastern United States were described by the SCS curve-number equation.

2.2.3 Evapotranspiration

Evapotranspiration (ET) is the combination of water loss by surface evaporation (Eo) and plant transpiration. The rate of ET is affected by solar radiation, wind speed and turbulence, relative humidity, and available soil moisture in addition to plant species, age, and rooting depth (Carter, 1986). In wetlands, the presence of aquatic plants can greatly increase the loss of water compared to surface evaporation alone depending on the plant transpiration rates (Winter, 1981). On the other hand, it has been shown that plants can reduce the amount of surface evaporation compared to open non-vegetated water bodies. Open water marshes showed a 30 to 80 percent decrease in surface evaporation due to the increased shade, humidity, and reduction in wind speed provided by the plants. This does not indicate that ET also decreases due to the vegetation because the transpiration rates of plants are usually more than the reduction of surface evaporation (Rushton, 1996). Depending on the plant transpiration rate, which varies from season to season, the ET to open surface evaporation ratio (ET/Eo) may be greater or less than one. The ET/Eo ratio was 2.5 for cattail, 1.2 for pickerel weed, and as low as 0.85 for duck weed and bog mat (Rushton, 1996).

Numerous algorithms have been developed to estimate ET using readily available climatological data. One simple and well-known algorithm for ET estimates is the
Thornthwaite method (Thornthwaite, 1948; Veihmeyer, 1964; Mitsch and Gosselink, 1993). Thornthwaite's method appears to be a good approach to estimating ET for wetlands because it assumes that soil moisture is not limiting as is the case for most wetlands (Pierce, 1993).

Thornthwaite developed his empirical equation using easily obtainable climatological data for mean monthly temperature to calculate potential evapotranspiration (PET). PET is the water loss under conditions of continuously adequate soil moisture as opposed to actual evapotranspiration (ET) which is the actual water loss to the atmosphere. Thus the ratio of ET to PET is usually $\leq 1$ (Mather, 1961). After examining monthly water-loss data from land areas throughout the U.S., Thornthwaite concluded that there was no simple relationship between monthly evapotranspiration and monthly temperature. In 1948, Thornthwaite developed a heat factor: $I = \sum (T/5)^{0.514}$ where $T$ is the mean monthly temperature for each of 12 months in degrees Celsius. This is used in the general equation: $PET = 1.6 (10T / I)^a$.

The exponent $a$ is expressed by the relation: $a = 0.49239 + 0.01792 (I) - 7.71EE-5 (I^2) + 6.75EE-7 (I^3)$. The value of $a$ varies from 0 to 4.25 and $I$ between 10 and 150 (Thornthwaite et al., 1944; Thornthwaite, 1948; Veihmeyer, 1964). The PET rate obtained using the above algorithm is referred to as the unadjusted PET rate because it does not take into account the variation in the number of hours in the day with season and latitude and the variation in the number of days in a month which varies from 28 to 31. The adjusted PET rates can be calculated by either reducing or increasing the unadjusted PET rate by tabulated adjustment factors that vary according to latitude and month (Thornthwaite et al., 1944; Veihmeyer, 1964).

Thornthwaite's equations were based on fits to actual data on ET under non-limiting conditions. PET values calculated with the Thornthwaite algorithm using weather bureau temperature data compared well with annual observed consumptive use data in irrigated valleys in the western U.S. In this case, the term consumptive use was considered equivalent to PET. The percent difference between PET and consumptive use was very close ranging from 0.2 to 3.7 percent except for one site in Tule Lake, California which was 14.5 percent different (Thornthwaite et al., 1944).

To check his formula further, Thornthwaite needed additional data for comparisons.
Without additional consumptive-use data available, he used annual observed runoff maps prepared by the National Resources Board (NRB). He estimated water available for runoff for comparison to the NRB runoff maps. Thornthwaite computed runoff estimates by comparing monthly estimates of potential evapotranspiration (PET) with observed monthly precipitation (P). He then used a soil water storage factor to estimate values of the actual evapotranspiration (AET). When the water storage factor was $> 0$ and $\leq 10$ cm, the $AET = PET$. If the water storage factor was $\leq 0$ then the $AET = P$ when $P < PET$ and the $AET = PET$ when $P > PET$. The water available for runoff was calculated as the $AET$ subtracted from $P$. Runoff maps made using these values compared very well with the runoff maps prepared by the NRB (Thornthwaite et al., 1944; Thornthwaite, 1948) even though seepage and other factors were not considered.

Thornthwaite compared his computed runoff with actual observed runoff for individual watersheds for long series of years. For one test, Thornthwaite computed runoff for the James River Basin in Virginia and compared his values to observed runoff for the Great Valley of Virginia. Although his computed runoff was not for the exact location where the actual runoff values were observed (due to the unavailability of temperature and precipitation data), there was close agreement between the two sets of results. Similar computations for other watersheds were performed with acceptable results (Thornthwaite et al., 1944). The Thornthwaite equation predicts observed PET values well without directly taking into account other primary controlling factors of PET, namely, wind, humidity, solar radiation, and plants. Thornthwaite explained that these important influences on PET vary together with temperature (Thornthwaite and Mather, 1955).

Direct measurement of ET with field instrumentation is difficult and requires long term measurements. In addition, field instrumentation must be properly installed and maintained (Carter, 1986). There are several methods that directly measure ET in the field using evaporimeters such as evaporation tanks and pans, and lysimeters. Indirectly, it is possible to estimate ET from measurements of the energy budget in the plant canopy over uniform areas (Hsu et al., 1972), soil moisture sampling, inflow-outflow measurements of water or measuring the water vapor increase in air flowing through vegetation chambers (Veihmeyer,
A method described by White (1932) directly measures ET in the field based on the diurnal cycles of groundwater levels in observation wells (Todd, 1964; Mitsch and Gosselink, 1993). It has been used successfully for measuring the ET of wetlands (Dolan et al., 1984; Farrington et al., 1990; Gilman, 1994). The method was simple and used relatively inexpensive instrumentation (Smithers et al., 1995). White (1932) developed this method in his study to estimate groundwater use by plants (transpiration) and evaporation from soil. This study was performed in Escalante Valley, Utah where the water table was continuously recharged by lateral groundwater seepage. Seventy-five observation wells were installed in fields with hydrophytic vegetation exploiting the shallow groundwater. The vegetation consisted mainly of greasewood (*Sarcobatus vermiculatus*), shadscale (*Atriplex confertifolia*), salt grass (*Distichlis spicata*), tussock grass (*Sporobolus airoides*), pickleweed (*Allenrolfea occidentalis*), and seep weed (*Dondia moquinii*). Groundwater levels were measured with a steel tape or with an automatic water stage recorder. Stage recorders were used on most wells for varying lengths of time. Almost all wells were installed in areas where the mean depth to the groundwater during the growing season was less than 3.0 meters (10 feet). The capillary fringe in these areas ranged from less than 2.5 cm in clean gravel to 2.4 m (8 feet) or more in silt or clay.

During the growing season the continuous records of water levels in these wells were observed to fluctuate diurnally in a regular fashion, starting to decline between 9 to 11 a.m., reaching a low between 6 to 7 p.m., then starting to rise again between 7 to 9 p.m. until 7 to 9 a.m. These diurnal cycles were superimposed on the annual seasonal trend, therefore the drawdown during the day can be greater than or less than the night time recovery depending whether the seasonal trend was decreasing or increasing on that day. The cycles began in the spring with the appearance of foliage and persisted until the fall frost when the vegetation died back or became dormant. The diurnal cycles did not occur in non-vegetated areas, such as in plowed fields and cleared lands, in areas where the water table was far below the surface, and in areas where the plants were cut. However, the seasonal trend was still observed in the non-vegetated areas. The magnitude of the daily decline increased with increasing temperature,
wind, and sunlight intensity and decreasing humidity and varied with the stage and vigor of plant growth. The daily decline decreased on cloudy days and remained constant or rose on rainy days indicating that ET was minimal or stopped during these times. The conclusion from these observations was that ET was the direct cause of the observed diurnal cycles (White, 1932).

White (1932) also measured the amount of water removed by ET and the diurnal cycles of the water table in experimental vegetated tanks layered with gravel then filled with soil to simulate the hydrologic and soil and plant conditions in the field. The tanks contained alfalfa (*Medicago sativa*), greasewood (*Sarcobatus vermiculatus*), and salt grass (*Distichlis spicata*) transplanted from the field. A pipe was inserted vertically through the soil and into the gravel bed at the bottom of each tank. Water was supplied from a reservoir through this pipe to continuously replenish water lost by ET. The level in the pipe was maintained at a constant height to duplicate the slightly positive pressure conditions in the field. The water fed to the pipe flowed into the gravel bed then upward into the soil due to the slight positive pressure. The water table levels in these experimental tanks showed similar diurnal cycles as those observed in the field monitoring wells. Evidently, the drawdown observed during the day was caused by plants removing water directly from the capillary fringe and from soil surface evaporation if the capillary fringe was near the soil surface. The water depleted from the capillary fringe was replaced by upward groundwater movement through capillary action. However, during the day, the depletion occurred faster than the continuous recharge of the water table, and this resulted in a lowering of the water table which increased the hydraulic head between the water table in the tank and the fixed level in the pipe maintained by the reservoir of water. The height (*H*) of water in the soil lost during a 24-hour cycle of ET was equal to the observed daily water table drop (*Δd*) plus the simultaneous rise of the water table due to the positive water pressure (*dailyΔr*), i.e. $H = Δd + (dailyΔr)$. During the night when ET became negligible, the water table rose to restore hydrostatic equilibrium. Therefore, the height the water table rose during the night (*nightΔr*) was equal to the height the water table fell during the day (*Δd*), i.e., $Δd = (nightΔr)$. Substituting gave $H = (nightΔr) + (dailyΔr)$. This means that the 24-hour rise of the water table that resulted from the positive water pressure
was the same as the loss of water due to ET and Eo. White (1932) found that there was little movement of water to the capillary fringe from the water table in the tanks during the hours from midnight to dawn. He found that the mean rate of rise of the water table, which changed continuously depending on the hydraulic head, occurred between midnight to 4 a.m. during each 24-hour period of observation. The rate of rise increased as the water table dropped increasing the hydraulic head between the water table and the fixed elevation in the pipe. It decreased as the water table rose back to equilibrium decreasing the hydraulic head.

Therefore, when the water table was at some fixed mean level, the hydraulic head driving its rise was at a corresponding mean value during the 24 hours if there was no seasonal trend. White (1932) neglected any slight losses by ET between midnight to 4 a.m. and the differences in the pressure head at this time and used the hourly rate of recharge from midnight to 4 a.m. as the average rate for the 24 hour period. Thus, \( H = (24r) \), where \( r \) was the hourly rate of rise of the water table at the mean hydraulic head from midnight to 4 a.m.

The daily groundwater lost to ET by the plants in the vegetated tanks was therefore be computed as \( ET = y(24r) \). Here \( ET \) was taken as equal to the depth of groundwater withdrawn which was equal to the 24-hour change in the water table elevation \( (H) \) times the specific yield \( (y) \) of the soil. The specific yield, defined as the depth of water gained or lost per unit change in water table elevation, was determined in the field studies by observing the changes in the water table elevation in a column of undisturbed soil excavated near the observation wells after the addition or drainage of known depths of water.

In contrast to the constant recharge level maintained in the tanks, there was a cyclic seasonal trend for recharge in the field due to the regular precipitation cycles over the recharge areas. This seasonal trend would be included in the observed changes in the water table elevation during the 24 hour interval. Therefore, for the field observations, the height the water table rose during the night \( (nightΔr) \) was not equal to the height the water table fell during the day \( (Δd) \) as it was in the vegetated tanks, i.e., \( (Δd) \neq (nightΔr) \). The daily water table drop \( (Δd) \) was equal to the nightly water table rise \( (nightΔr) \) plus the net fall \( (+) \) or rise \( (-) \) of the water table \( (s) \) due to the seasonal trend during the 24-hour period from the previous 24-hour period, i.e., \( Δd = (nightΔr) \pm s \). Substituting this into \( H = Δd + (dailyΔr) \)
gave $H = (\text{night } \Delta r) \pm s + (\text{daily } \Delta r)$, which was equal to the 24 hour rise of the water table plus or minus the seasonal trend. The value of $s$ was added if the trend was decreasing and subtracted if the trend was increasing. White (1932) found that the difference between the average hydraulic head between midnight and 4 a.m. and any increase or decrease of hydraulic head due to the seasonal trend was generally not great and can be ignored. Therefore, the average rate of rise in the water table from midnight to 4 a.m. could again be used as the average rate for the 24 hour period. With these assumptions, $H = (24r \pm s)$. The ET was computed by the formula $ET = y(24r \pm s)$. Values for $r$ and $s$ were determined from the water table elevations in wells equipped with automatic water level recorders (White, 1932). From these elevations the hourly rate of rise ($r$) was calculated as the change in height that the water table rose from midnight to 4 a.m. divided by 4 hours. The seasonal trend ($s$) was calculated as the systematic rise or fall of the water table from midnight to midnight of consecutive days.

To verify that the ET calculated using the diurnal cycles in groundwater levels were reasonable, White (1932) directly measured ET as the amount of water supplied from the reservoir to the alfalfa tanks and compared it to the ET calculated from the diurnal water table cycles in the tanks. The ET was measured as 0.214 inches per day. Using the formula $ET = y(24r)$ he calculated 0.166 inch/day, which is 78 percent of 0.214. The discrepancy in the values was attributed to daily changes in tank temperature which may have occurred during the period when these measurements were obtained. A mulch of sawdust on top of the tank was only partly effective as a heat insulator because it was disturbed by the wind. In a series of experiments White (1932) showed that the water table declined or increased from 0.38 cm to 0.76 cm with a fall or rise, respectively, of 1 degree centigrade in temperature of the soil in the tanks. He found it impossible to determine a correction for temperature in the vegetation tanks (White, 1932).

The amount of water lost to ET per unit weight of dry plant matter per unit area in the tanks was also computed from the measured water supply. This coefficient was used to calculate the ET in the field from the dry weight of plant matter produced per unit area. This was compared to the ET estimates from the formula $ET = y(24r \pm s)$. The difference in ET between the two methods was 0.4 inches for alfalfa, 10.5, 1.8, and 1.7 inches for salt grass,
1.5 inches for a shad scale and greasewood combination, and a difference of 1.6 inches for greasewood. The ET estimates from the two methods were very close in agreement except for the 10.5 difference in one salt grass comparison which was partly due to an inaccurate specific yield measurement.

Measuring the diurnal cycles in groundwater levels requires that the capillary fringe be in the root zone. When the water table is at or near the soil surface the evaporation component of ET increases and vice-versa (Todd, 1964). When the water table is above ground surface, the soil's specific yield is equal to one so that the change in water table elevation above ground surface is equal to the loss of water to ET. Observation of diurnal cycles in groundwater levels therefore appears to be a possible method for estimating ET in wetlands. It measures ET directly unlike many other methods such as the pan evaporation, mass transfer, and energy balance methods which measure or estimate surface evaporation and estimate ET using empirical adjusting coefficients (Veihmeyer, 1964; Winter, 1981). Smithers et al. (1995) showed that the diurnal cycles in groundwater levels were sensitive enough to show the different rates of ET between the plant communities of sedges and reeds. He compared these ET rates to the evaporation from a Class A-pan. The Class A-pan evaporation was similar to the estimated ET from the sedge meadow, but was much larger than the ET from the reedbed (Smithers et al., 1995). Dolan et al. (1984) measured ET from the diurnal cycles in groundwater levels from a Florida freshwater wetland. The ET rates from his study compared favorably with literature values for a wide variety of freshwater and marine wetland communities. Dolan et al. (1984) also estimated average daily ET rates using the Thornthwaite algorithm for the same time period. He provided the average daily rate of ET for the same months for each method. The ratios between the two methods were 1.0, 0.8, 1.0, 0.9, 1.0, 1.4, and 1.2 for May, June, July, August, September, October, and November, 1977 respectively. The ratios were 1.8, 2.0, 1.4, 1.4, 1.4, and 1.1 for January, February, March, April, and May, 1978 (Dolan et al., 1984).

Measuring ET from diurnal cycles in groundwater level has its disadvantages. The cycles are measured during time intervals with no rain, therefore the monthly ET rates are probably overestimated because ET decreases on rainy days (Smithers et al, 1995). Another
disadvantage of the method is the sensitivity of the formula to the specific yield. White (1932) concluded that ET estimates from the diurnal cycles in groundwater levels were easily obtained from water level recorders, but the specific yield of the soils in which the cycles take place was exceedingly difficult to determine. The fundamental assumption in this method is that the average hourly rate of rise during recovery of the water table is equivalent to the average hourly rate of supply of water to the plant roots from the water table. The validity of this assumption has not been tested in wetland environments.

2.2.4 Groundwater Flow

Groundwater flow can be large or non-existent and is a difficult and time-consuming water budget component to measure (Carter, 1986; Novitzki, 1989). Groundwater moves in the direction of the hydraulic gradient and is described by Darcy's Law as \( Q = -K\frac{dh}{dl} \), where \( Q \) = flow (m\(^3\)/s), \( K \) = saturated hydraulic conductivity (m/s), \( A \) = cross sectional area perpendicular to flow (m\(^2\)), and \( \frac{dh}{dl} \)=hydraulic gradient (m/m) (Fetter, 1994). Vertical flows are often negligible for wetlands in soils with high clay content and very low vertical hydraulic conductivity or in soils over an impervious stratum. Hydraulic conductivity can be estimated from grain size analysis of the aquifer material or measured in laboratory permeameter tests or from pump tests in field wells. Grain size analyses provide only a rough approximation for \( K \). Ranges in cm/s are 1.0-0.01 for coarse sand, 0.01-0.001 for fine sand, 0.001-0.00001 for silty clay, and less than 0.000001 for clay (Das, 1994). Field tests are more accurate than laboratory tests because they are representative of a larger portion of the aquifer (Winter, 1981). Studies show that laboratory tests can differ from on-site tests by as much as 100 percent (Winter, 1981).

Groundwater flows can be estimated directly in the field using a network of monitoring wells or piezometers. The monitoring wells are fully screened over the water table and are used to measure the water table elevation for horizontal hydraulic gradient. Piezometers are open ended to measure the pressure head at a certain depth and are useful in determining the vertical hydraulic gradient. Water table elevations in monitoring wells are used to calculate point values of the hydraulic head. These values are then interpolated areally to obtain
hydraulic head contours which are then used to obtain magnitudes and direction of hydraulic gradients. Groundwater flows are then calculated using Darcy's equation and measured values of K from pumping tests. Well density required for accurate hydraulic gradient contours varies considerably (Winter, 1981). Studies show that hydraulic head contour maps based on sparse data points can have a large error (Winter, 1981). Hanson (1972) found that three contour maps of 1.6 km, 6.4 km, and 9.7 km well spacings had errors in contour lines that differed from an accurate map by 0.028 m, 0.35 m, and 0.40 m, respectively. Application of Darcy's Law assuming homogeneous aquifer conditions may not be appropriate when physical and hydrological properties of aquifer materials in wetlands are poorly understood or have been disturbed (Carter, 1986; Mitsch and Gosselink, 1993).

Groundwater flow in a wetland was measured by Rushton (1996) as the residual of a simplified water budget equation. For days having no precipitation or surface water flow the total water budget equation \( P + SWI + GWI = ET + SWO + GWO + \Delta S \) simplifies to \( \pm \Delta GW = ET \pm \Delta S \). And assuming no ET at night the equation simplifies to \( \pm \Delta GW \) (night) = \( \pm \Delta S \) (night) assuming all errors equal zero. The period with no ET was taken between 6 p.m. and 6 a.m. Measurements were made when the water table was above the ground surface. The results indicated that the net groundwater seepage can be calculated directly from the nightly water level cycle. These groundwater seepage estimates compared well with other estimates calculated as the residual of the total water budget with an error of 7 percent for three-month periods and 2 percent when compared on a yearly basis (Rushton, 1996).

2.2.5 Storage

The change in storage component (\( \Delta S \)) of the wetland water budget consists of changes in storage volume of surface water, soil moisture in the vadose zone, and groundwater (Fetter, 1994; Smithers et al., 1995). Change in storage of surface water over a specified time interval is usually expressed as the change in level of the free water surface (\( \Delta l \)) of the wetland times the areal extent of the ponded area of the wetland as a function of water level \( A(l) \), i.e. \( \Delta S = \Delta l\times A(l) \). The change in storage of the sub-surface layer outside of the ponded area can be calculated using the change in the water surface in conjunction with the specific yield (\( y \)) of
the soil, i.e. \( \Delta S = y (\Delta I \times A(l)) \) (Smithers et al., 1995). The ponded area, and therefore the volume of open water of the wetland, increases as the water level increases and vice-versa. Smithers et al. (1995) estimated the change in surface storage of the wetland by establishing a stage to surface storage relationship derived from topographic maps. Winter (1981) points out that the change in volume of lakes is seldom accounted for in water budgets and little has been mentioned about the errors involved in ignoring this change. Owen (1995) found that the error in estimating the change in subsurface storage was subject to error in measuring specific yield and water levels. In his study, specific yield was estimated by dividing the rainfall on-site by the water table rise in hydrographs assuming that the rise is attributed to the total rainfall and does not take runoff into account. He showed that the error associated with using such estimates of specific yield resulted in an error in seasonal ET of 43% to 60% compared to ET calculated using a mass balance approach (Owen, 1995).

### 2.3 Modified Pierce Water Budget Model

The modified Pierce water budget model recommended by the Norfolk district ACOE assumes that an outflow structure, such as a spillway or other engineered structure, will be designed to maintain a specified maximum free water level in the mitigation wetland. In order to predict inter-year variations for determining the maximum level, Pierce (1993) recommends calculating monthly values of each element of the water budget for the driest, wettest, and most typical year based on precipitation data for the past 30 years. This would provide a more realistic estimate of the range of the monthly storage and water levels in the wetland from year to year. This requires quantifying the primary sources and losses of water prior to wetland mitigation.

The modified Pierce model accomplishes this by estimating the potential storage of water in the wetland as the difference between the inflows and the uncontrolled outflows expressed as depth over the constructed site with reference to an arbitrary datum. The change in potential storage is reflected in the level change of the free water surface. Potential storage is calculated based on estimates of the inflows from precipitation and runoff and the outflows from ET and groundwater seepage. Potential storage does not include stream baseflow,
stream outflow and groundwater inflow.

Precipitation data are taken from a nearby weather station, potential ET is calculated using the Thornthwaite (1948) method, and runoff using the SCS runoff method (USDA, 1986). Groundwater seepage from the wetland can be calculated using Darcy's equation with an estimated hydraulic conductivity of the soil. In large areas with heterogeneous soils, Pierce (1993) suggests measuring the range of soil hydraulic conductivities and evaluating the groundwater seepage at the extreme values of these ranges. Monthly values of water budget components measured as volume are divided by the wetland area to obtain depth (Westbrook, 1994).

The potential storage is the maximum amount of water available for a given wetland design area. When the potential storage is greater than the maximum depth of water the wetland can hold as set by an outflow structure, the actual storage of the wetland is equal to the maximum depth. The wetland storage would remain at this maximum depth until the potential storage becomes a negative value meaning a net monthly loss of water. The actual storage of the wetland is then the maximum depth minus the net loss. The actual storage for each month is plotted to determine the drawdown regime in the constructed wetland relative to the maximum depth of storage (Westbrook, 1994).

Pierce (1993) has applied his water budget method to wetland designs and has implemented successful wetland construction projects for mitigation purposes and shoreline erosion control throughout the eastern seaboard and the Midwest. The success rate of the modified Pierce model has not been documented, although it closely follows Pierce's (1993) model.
3. MATERIALS AND METHODS

3.1 Introduction

One objective of the wetlands mitigation research program administered by the Virginia Transportation Research Council (TRC) is to develop a workable approach to estimating the annual water budget for proposed and constructed wetland mitigation sites. One such site at Manassas is an existing wetland currently targeted to be enhanced and expanded into a emergent, scrub-shrub, and forested wetland habitat. Researchers at Virginia Tech are involved in a cooperative effort with TRC, through parallel funding provided by the Virginia Department of Transportation (VDOT) Environmental Division, to assess the soil-hydrologic regime at the Manassas site. Mitigation activities at the site are coordinated by VHB, Inc., Williamsburg, VA as part of their statewide wetland services contract with the VDOT Environmental Division. VDOT needs such assessments at designated compensation sites before grading commences (or before the land is actually purchased) in order to predict what the soil wetness regime will be across the site after the development is completed. As part of this assessment, this study was designed to evaluate a currently recommended procedure for estimating the wetland water budget components at the site by comparing estimates for each component with those obtained from actual measurements at the wetland site. The Manassas site was recently acquired by VDOT with the intent of expanding and enhancing the existing wetland acreage to mitigate for wetland impacts of work on the nearby Route 234 Manassas Bypass corridor. VDOT has proposed mitigation plans to grade down marginal upland areas surrounding the existing wetland and place a water control structure to detain water at the current level of the naturally ponded zone at the site.

3.2 Wetland Site Description

The wetland site (Figure 1) is located in Manassas, Prince William County, VA. It is south and adjacent to the Manassas Airport which is located south of Route 28 on Wakeman Road. The site is located in the Culpepper Triassic Basin centered at 38° 43′ 15″ latitude and 77° 30′ 20″ longitude. The wetland was measured as approximately 8.5 ha. The mitigation site is currently a natural wetland site undergoing secondary succession from previous
agricultural use that ceased in 1982. The previous farming operation encompassed the entire wetland. The land was farmed with the aid of a drainage system. Drainage tiles have been found at the south end of the wetland and are no longer operational. All drainage tiles appeared dry upon inspection except for one where a negligible amount of water flowed.

The wetland is dominated by an open-water/emergent wetland system at its center that is enhanced by beaver activity on Cockrell Branch which flows through the site. The beavers established a dam at the south end of the wetland which increased the potential storage by raising the water level of the open water/emergent portion of the existing wetland. During 1996 at least two attempts were made to remove the beavers from the area and the dam was broken during March 1996. Removing the beavers proved to be unsuccessful because the beavers shortly rebuilt the dam. VDOT decided to leave the beaver dam and incorporate its elevation into the wetland mitigation design.

Cockrell Branch collects water from one main watershed (W1) which is 436.4 ha. The remaining watersheds contributing water to the wetland are smaller in size and have been combined as watershed 2 (W2) totaling 67.1 ha. Cockrell Branch forks into Broad Run at the southwest end of the wetland. The open-water/emergent wetland zone is surrounded by a transitional shrub-scrub/forested wetland that grades up in elevation into non-jurisdictional wetland areas. A small stream flows into this shrub/scrub area and is a primary water source maintaining the wetness regime of the higher areas (north) surrounding the central portion of the wetland. The wetland is adjacent to Broad Run, riparian to Cockrell Branch, and is underlain by red weakly cemented Triassic siltstones and shales. The bulk of the wetland area appears to be formed in relict alluvium from a past meander of Broad Run, but significant amounts of local slope colluvium and alluvium from Cockrell Branch are also found throughout the site.

The wetland boundary used in this study was delineated by VHB, Inc. based on the prescribed criteria for delineation of jurisdictional wetlands. Jurisdictional wetlands criteria specify characteristics of hydrophytic vegetation, hydric soils, and wetland hydrology as outlined in the federal manual for identifying and delineating jurisdictional wetlands (Federal Interagency Committee for Wetland Delineation, 1989).
was used as the area boundary for all water budget calculations and is shown in Figure 1.

According to the Prince William County Soil Survey (Elder, 1989), three soil series are predominantly found in the wetland area. Bermudian silt loam (Fluventic Dystrochrepts), 0 to 2 percent slope, is located on the western side of the wetland along Broad Run. Aden silt loam (Aeric Ochraqualfs), 0 to 2 percent slope, is distributed in the center of the wetland. Dulles silt loam (Aquultic Hapludalfs), 0 to 4 percent slope, is distributed around the eastern side of the wetland. In April, 1996, Mr. Bob Hodges (Virginia Tech soil scientist and surveyor) field checked the existing Prince William County Soil Survey and found the majority of soils mapped were representative of their series and/or mapping units. Where soils differed from the mapped unit, they did not differ significantly in drainage class. Observations from soil borings and pits indicated that the soils are laterally variable in depth to rock and/or their stratification due to the complexity of the fluvial and colluvial processes over time.

Wetland plant communities were documented from a detailed field reconnaissance conducted by VHB, Inc. The wetland plant communities were characterized as riparian scrub thicket and emergent marsh. The riparian scrub thicket occurs in areas where regular flooding occurs along small stream mouths and channels. Most prominent is the area at the Cockrell Branch delta. Characteristic woody plants are black willow (Salix nigra Marsh.), tag alder (Alnus sp.), elderberry (Sambucus sp.), silky dogwood (Cornus amomum Mill.) and sycamore (Plantanus sp.), maples (Acer spp.), and green ash (Fraxinus pennsylvanica L.). There is a very dense herb cover consisting of grasses (Poaceae family), mainly panic grasses (Panicum spp.), sedge (Carex spp.), rushes (Juncus spp.) and many emergent wildflowers. Typical herbs are spotted touch-me-not (Impatiens capensis Meerb.), tearthumb (Polygonum arifolium L.), water hemlock (Cicuta maculata L.), rice cutgrass (Leersia spp.), and leathery rush (Scirpus acutus Bigelow). The emergent marsh is an expansive complex that occupies the central portion of the site. Due to continuous beaver impoundment, much of the area is covered with a dense growth of soft rush species. Interspersed with this characteristic plant are rattleboxes (Ludwigia spp.), sedges (Cyperaceae), spikerushes (Eleocharis spp.), bedstraw (Galium spp.), water willow (Justicia americana L.), water plantain (Alisma sp.), and starwort (Callitriche sp.). In numerous places, there are larger patches of cattail (Typha sp.)
and buttonbush (*Cephalanthus occidentalis* L.). Scattered colonies of trees and shrubs, including maple (*Acer* spp.), sycamore (*Plantanus* sp.), alder (*Alnus* spp.), green ash (*Fraxinus pennsylvanica* L.), common persimmon (*Diospyros virginiana* L.) and especially willow (*Salix* spp.), are common.

During 1996, Dr. G. R Whittecar, Old Dominion University, Norfolk, VA assessed the groundwater flows at the site through geological observations of faults and regional groundwater flows, piezometric surface maps, and nested piezometers. He summarized his findings as follows: "Upward movement of groundwater along bedrock fractures is not likely to be a major source of water feeding the wetlands at the site. Analyses of topographic linears and fracture patterns in bedrock suggest that the orientations of valleys in the study area are strongly controlled by tectonic fractures. However, the orientation of the largest valley tributary to the wetland is not one of the most common fracture orientations in the region and thus is not expected to be a discharge point for regional groundwater flow systems. Head data from three piezometer nests placed in alluvium and in fractured saprolite at sites around the existing wetland support these conclusions" (Whittecar, 1997).

### 3.3 Overview of Site Instrumentation and Field Measurements

The Manassas site has been of interest to VDOT for several years. In March and April, 1992 they installed twenty-nine water table observation wells at the site using a truck-mounted hollow stem auger. The wells consisted of 10.2 cm (4") auger borings to depths ranging from 3 meters (10 ft) maximum or to rock if shallower. Well construction included a 10.2 cm (4") PVC point, a 12.7 cm (5 ft) length of 5.1 cm (2") inner diameter (ID) PVC pipe with 0.051 cm (0.02") slotted screen, and a 12.7 cm section of riser on top. Boreholes were advanced to 3 meters unless auger refusal occurred (Dan Redgate, VHB Inc., Williamsburg, VA, personal communication). Borehole annular spaces were not backfilled with well gravel and sealed with bentonite (W. Lee Daniels, Virginia Tech, personal communication). These wells were monitored weekly by VDOT from April - July, 1992.

In early 1996, Virginia Tech (VT) researchers installed 24 additional water table observation wells. These wells were hand augered and lined with slotted PVC 5.1 cm ID
pipe. A sand filter pack was used to fill the annular space around the pipe and a bentonite surface seal was used to limit surface water leakage into each well. The water levels in both the VDOT and VT wells were measured weekly by VT personnel from March 1996 - November 1997 and monthly from December 1996 - June 1997. A well location map is provided as Figure 2.

The United States Geological Survey (USGS) installed four wells, two stream gages, and a tipping bucket rainfall gage in early 1996. The USGS well used in this study is designated as USGS1 (Figure 2). The well was constructed from schedule 40, 5.1 cm (2") ID PVC pipe with 0.025 cm (0.01") slotted screen. It was installed by hand augering, backfilling with filter sand, and placing a bentonite seal on top. USGS1 is screened from the land surface to approximately 0.6 meters below, and is located in the emergent/open water wetland. Telog brand groundwater pressure transducers and data loggers were used to monitor the water levels in this well. Telog provides the whole system including software. The data were downloaded on either Telog supplied hand-held data transfer units or a laptop computer with Telog software. The pressure transducer was positioned slightly above the bottom of the well (Michael Focazio, USGS, personal communication). The sensitivity of the continuous water level recorders is 0.031 cm (0.001 feet). Water levels were monitored every hour in USGS1 since 10 July 1995 to 2 October 1996.

The stream gages were installed on Cockrell Branch at a point before the wetland (inflow) and at the point where the stream leaves the wetland (outflow). The stream gages had shaft encoders that were used to record stage. The encoders were connected to Campbell Scientific CR-10 brand data loggers. The shaft encoders were a float and pulley type mechanism that were hooked to a wheatstone bridge circuit and an analog to digital (A-D) converter. As the float moved with stream stage, the pulley moved a shaft that changed the resistance of the circuit. The signal was passed via the A-D converter to the CR-10 for storage. USGS software was used to download and manipulate the data with a laptop computer (Michael Focazio, USGS, personal communication).

Stream flow and precipitation data were measured from 9 May 1996 through 27 February 1997. The stream gages were designed to take half hour water level (stage) readings with
0.031 cm (0.001 feet) sensitivity. The wetted channel cross sectional area and stream velocity were measured to calculate discharge at these stream gages during several periods for varying stages. The channel area was measured by first stretching a measuring tape across the stream and dividing it into 12 vertical sections of equal width ($w_i$). The depth of water (stage height) of each section ($d_i$) was measured by a wading rod. A Price brand current meter was suspended on a cable and lowered into the stream with a streamlined weight to measure the velocity ($v_i$) of each section. The current meter was calibrated against the average of two stream velocities taken at 0.2 and 0.8 times the stream depth. The discharge of each of the 12 sections ($q_i$) at a particular stage height was calculated by multiplying the stream velocity by the area of the channel, i.e. $q_i = v_i d_i w_i$. The total discharge ($Q$) for the stream is the sum of the discharges ($q_i$) for each segment. When the stage was too low for this type of measurement, a Pygmy meter replaced the current meter calibrated against the velocity of the stream taken at 0.6 times the stream depth (Fetter, 1994; Michael Focazio, USGS, personal communication). Plotting the stage against flow discharge ($Q$) and fitting a second degree polynomial resulted in a rating curve. Stream discharge for any given set of stage heights can be determined from the curve.

The tipping bucket rain gage was installed on the outflow stream gage platform and connected to the stream gage's CR-10 datalogger. The volume of the tipping bucket of the rain gage was designed to record precipitation to an accuracy of 0.03 cm (0.01 inches). Twenty-four-hour precipitation totals were obtained by summing the gage readings which were accumulated every 30 minutes on the datalogger.

Two sets of nested piezometers and one shallow monitoring well were installed by Rich Whittecar of Old Dominion University. The findings in his report "Geological Controls on the Movement of Shallow Groundwater, VDOT Manassas Mitigation Wetland Site" were used in this study. Five wells were installed to act as sets of piezometers: one shallow well beside monitoring well VT119 in the stream valley of Cockrell Branch (AMRL03); a shallow (AMRL04) and a deep well (AMRL05) approximately 200 feet downstream of the inflow stream gaging station; and a shallow (AMRL01) and a deep well (AMRL02) next to VT106 (Figure 2). They were monitored on the days they were constructed in November 1996 and
3.4 Evaluation of the Modified Pierce Model for Estimating Water Budget Components

The modified Pierce water budget model utilizes common and relatively simple methods to calculate the monthly potential storage of water in a wetland for use in wetland construction. The monthly potential storage is the amount of water the wetland gains or loses assuming no surface outflows. The potential storage, expressed as a height of water, is used to design structures to keep the water level in the wetland continuously at or below a set maximum height considered optimal for supporting the wetland vegetation. The model calculates the potential storage of water as the difference between the inputs and any outputs of water that would not be controlled by the outflow structure. The inputs of water consist of precipitation and surface runoff and the outputs of water are evapotranspiration and groundwater seepage. The existing wetland boundary was used as the areal boundary for expressing all water budget components as height of water.

Measurements for precipitation, stream flow, evapotranspiration, and groundwater flow are necessary to assess the validity of the methods used to calculate each water budget component recommended in the modified Pierce model. The field measurements and the comparisons that were made for each component of the modified Pierce water budget model are detailed below.

3.4.1 Precipitation

In the modified Pierce model, precipitation data inputs are taken from the nearest rain gaging station. Daily precipitation records were obtained from the State Climatologist for Dulles Airport, Sterling, VA, approximately 20 miles northwest of the Manassas wetland site. This was the nearest weather station to the Manassas wetland site. The monthly on-site precipitation was compared to the Dulles precipitation for the same days using a double mass curve.

3.4.2 Runoff
The modified Pierce model estimates watershed runoff into the delineated wetland boundary by the SCS runoff method (Mokus, 1972). The runoff is the total surface flow into the wetland from the surrounding hills and watersheds. The SCS method relies on daily precipitation data, SCS soil hydrological classifications as defined by the SCS county soil survey, type of land cover, and the boundary area of the watershed contributing to the wetland from topographic maps.

Daily precipitation data obtained from the Dulles Airport weather station were used in runoff calculations of the modified Pierce model. The land cover was estimated based on aerial photographs and field reconnaissance. Runoff values were calculated using an antecedent moisture condition (AMC) II according to the TR-55 manual (USDA, 1986) as described in section 2.2.2. The watersheds that contribute water within the wetland boundary were delineated with the aid of USGS maps. The runoff from the largest watershed (W1) enters Cockrell Branch above the inflow stream gaging station. The remaining watersheds are smaller in size and have been combined as watershed 2 (W2). They include the watersheds draining from smaller streams located to the north, northeast, and south of the wetland and the area of land that surrounds the wetland. The boundary of all runoff contributions to the wetland extend no further than the outflow stream gage. The wetland itself was not included in the watershed area because the precipitation over the wetland is a direct water input. The volume of runoff produced from each watershed can be divided by the fixed delineated area of the wetland to obtain the equivalent depth of water available for storage in the wetland.

To determine the effect AMC has on runoff volume calculations using the SCS runoff method, the runoff volumes were computed using an AMC I, II, III, and a variable AMC determined from the 5-day rainfall before each storm event. These runoff volumes were calculated for W1 using the on-site precipitation data. The modified Pierce model also estimated runoff from W1 using off-site precipitation (Dulles weather station) with the SCS Runoff Method using an AMC II. These runoff values were compared to the volume of runoff that enters the stream channel from W1 as measured by the stream gage data after hydrograph separation. As already described, hydrograph separation allows partitioning of the total stream flow into base flow and runoff. These comparisons show which AMC
provides a runoff value closest to the gaged runoff. A comparison of runoff values calculated with on-site and off-site precipitation data for an AMC II was also studied to determine how these two precipitation data sources affect the runoff values. In addition, runoff was estimated by the SCS runoff method for both a historically wet and dry year.

3.4.3 Evapotranspiration

Potential evapotranspiration rates were calculated in the modified Pierce model by Thornthwaite's method using mean monthly temperature measurements. These temperature records were obtained from the Dulles weather station. The general equation:

\[ \text{PET} = 1.6 \times (10T/I)^a \]

was used to calculate PET where I is a heat factor, \( I = \sum (T/5)^{1.514} \), and T is the mean monthly temperature in degrees Celsius. The exponent \( a \) was expressed by the relation:

\[ a = 0.49239 + 0.01792 \times I - 7.71 \times 10^{-5} \times I^2 + 6.75 \times 10^{-7} \times I^3. \]

The PET rate was then adjusted to account for the variation of the number of hours in the day and the number of days in a month from 28 to 31 which varies with season and latitude (Thornthwaite et al., 1944; Chow, 1964). The adjusted monthly PET rates were calculated for May 1996 - February 1997.

The water table data from well USGS1 were graphed against time to depict the pattern of the diurnal cycles in groundwater levels. These diurnal cycles were used in the equation

\[ ET = y(24r \pm s) \]

for each day during dry periods when the water table was below land surface. Data for only these days in which diurnal cycles occurred, in combination with the specific yield estimates of the soil, were used to calculate the daily rate of ET.

The soil's specific yield used in the above equation was calculated as the difference in the volumetric water contents at saturation and 33.3 KPa. Specific yield was determined from core soil samples taken near the USGS1 well. Two samples were taken from the A horizon at depths of 0-5 cm and 10-15 cm and one from the B horizon at 50 cm. For each sample, the volumetric water content was measured at saturation and then at a pressure of minus 33.3 KPa using a pressure membrane apparatus.

The specific yield can have a large effect on the evapotranspiration value calculated. It is necessary to apply the specific yield at a soil depth equivalent to the depth that the cycle is
occurring. In calculating the evapotranspiration for the wetland, the estimated specific yield value that corresponded with the depth of cycle was used. Water level cycles were graphed and used in combination with the specific yield estimates of the soil to calculate a daily rate of ET.

Potential ET values obtained by the Thornthwaite algorithm and diurnal cycles in groundwater levels were compared. The daily ET values from the latter method for the set of suitable dates in a given month were averaged to obtain a daily rate. This average daily rate was then used to estimate a monthly ET rate for comparison with monthly Thornthwaite estimates.

### 3.4.4. Groundwater

The modified Pierce model uses Darcy's equation to estimate vertical groundwater seepage with a hydraulic gradient of one as discussed in section 2.3. The hydraulic conductivity of the soil was estimated from the particle size analysis of soil samples. Thirteen core soil samples were taken from the major soil types (Aden, Dulles, and Bermudian) in the wetland for particle size analysis. Three core soil samples were taken from Dulles, 6 from Aden, and 4 from Bermudian from depth of 0 to 50 cm. The average hydraulic conductivity of these soils was estimated from the particle size analysis results (Das, 1994). Darcy's equation was used to calculate the vertical groundwater seepage within the wetland boundary. For comparison, instead of a hydraulic gradient of one, the vertical hydraulic gradients of 0.39, 0.29, 0.026, and -0.035, calculated from piezometer data measured by Rich Whittecar, were used in Darcy's equation to estimate the vertical groundwater seepage.

Net groundwater flow also was estimated as the residual of the water budget during dry periods with no precipitation (P) and therefore, no surface runoff (RO). In this case the total water budget equation, \( P + RO + SI = ET + SO \pm \Delta S \pm \Delta GW \), is reduced to \( SI = ET + SO \pm \Delta S \pm \Delta GW \). Therefore, the groundwater flow rate (\( \Delta GW \) for inflow and -\( \Delta GW \) for outflow) was calculated as \( \Delta GW = SO - SI + ET \pm \Delta S \), where SO is total stream outflow, SI is total stream inflow, ET is evapotranspiration, and \( \Delta S \) is the change in storage (+\( \Delta S \) for increase and -\( \Delta S \) for decrease). The stream gage data and rating curves es