OBJECT-ORIENTED HYDROLOGIC AND WATER-QUALITY MODEL FOR HIGH-WATER-TABLE ENVIRONMENTS

By

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by

Christopher John Martinez
I would like to dedicate this dissertation and the time I have spent at the University of Florida furthering my education and professional and personal development to my mother. She would have liked to have seen this day.
ACKNOWLEDGMENTS

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A hydrologic and water quality model was developed for high water-table environments such as the flatwoods of Florida. The model was developed within the object-oriented framework of the ACRU2000 model. The model uses physical approximations suitable for highly conductive, poorly drained soils. The water quality component of the model uses nitrogen and phosphorus algorithms patterned after the GLEAMS model, with appropriate modifications for sandy, poorly drained, acid soils.

The hydrologic model operates on a daily time-step and assumes a hydrostatic distribution of soil moisture. Reference potential evapotranspiration can be estimated using the Penman-Monteith equation and the resulting atmospheric demand is applied in a top-down approach to intercepted water, ponded water on the ground surface, and to soil evaporation and plant transpiration. Vertical upward flow of soil moisture in response to evapotranspiration is approximated using a steady-state solution of Darcy’s
Groundwater flow can occur to or from a deep aquifer or an adjacent water body. Runoff from the land surface is assumed to occur via saturation-excess only.

The hydrologic component of the model was validated using observed data from three field sites. The validation established the model’s ability to predict water-table depths, soil moisture contents, evapotranspiration, and runoff volumes.

The water quality component of the model employs modifications for poorly drained, flatwoods soils that include the specification of optimal ranges of water contents affecting the rate of nutrient transformations, the effect of soil moisture on transformation rates under saturated or near-saturated conditions, the instantaneous, reversible sorption of phosphorus, and the extraction of nutrients into runoff water.

The water quality component of the model was validated for six experimental pastures. Model validation, while providing improved predictions of runoff nutrient loads compared to the model without modifications for shallow water-table environments, indicated several shortcomings of the model. These include the need for explicit representation of plant biomass and organic soil accretion and the need for more site- or region-specific information on nitrogen contents and the factors that control N and P cycling and retention in these soils.
CHAPTER 1
INTRODUCTION

Humid, shallow water-table environments such as the flatwoods of the southeastern United States are characterized by flat topography and moderately to poorly drained soils with high infiltration capacities, where significant interaction between surface water and groundwater occurs. Flatwoods occur throughout the southeastern coastal plain of the United States and cover approximately 50% of the land area of the state of Florida (Abrahamson and Harnett 1990) (Figure 1-1). The soils of the Florida flatwoods are composed primarily of Spodosols and to a lesser extent Alfisols. These sandy, acidic soils typically lack an abundance of the mineral components that are important to phosphorus (P) retention, Fe and Al oxides and aluminosilicate and metal-oxide clays (Mansell et al. 1995; Harris et al. 1996), particularly in surficial soil horizons. The loss of P from soil has been shown to be an important cause of eutrophication of surface water bodies. The low retention capacity of these soils is further exacerbated by the application of organic fertilizers. Organic acids can adsorb to mineral surfaces, reducing the capacity for adsorption and stabilization of P (Eghball et al. 1996; Graetz et al. 1999).

The mitigation of adverse effects from P losses from flatwoods soils to the aquatic environment requires

- Understanding the surface and subsurface hydrology of the flatwoods,
- understanding the fate of P in this environment, and
- disseminating knowledge in the form of best management practices.

Extensive laboratory and field research to accomplish these goals is expensive in terms of time and money, thus the interest in computer simulation models. Successful modeling
of the flatwoods requires accurate representation of the hydrology and P biogeochemical cycle. An accurate model of flatwoods systems should include an accurate representation of a shallow water-table, its contribution to evapotranspiration, and its effect on runoff generation. A successful model should also be capable of reflecting the factors that affect P retention and subsequent loss in runoff and groundwater. Many models suffer from a lack of specific knowledge of the most important environmental conditions affecting P retention when applied to specific locations. Thus models must be updated as our understanding of the environment grows; and models should be developed with such future expansions in mind. Such model design is enhanced by the concept of object-oriented programming whereby real world objects are presented more intuitively than they might be in procedural programming (Liang 2001). Few object-oriented models exist in the fields of hydrology and agricultural water quality. One object-oriented model, ACRU2000, is available for such model expansion.

**Objectives**

Using the object-oriented ACRU2000 model as the modeling platform to build upon, the main objectives of this work were twofold:

1. Develop and test a field-scale hydrologic module for shallow water-table environments, and

2. test the suitability of the nitrogen and phosphorus algorithms of the model with modifications for sandy, poorly drained flatwoods soils.

It is proposed here that an approximate hydrologic model, operating on a daily time-step and representing the major forcings affecting water-table depths and runoff generation, can effectively represent the hydrology of the flatwoods. The major forcings are rainfall, evapotranspiration, and groundwater flow to or from adjacent water bodies or boundary conditions. The approximate methods include
The assumption that hydrostatic soil moisture conditions prevail,
the vertical upward movement of soil moisture in response to evapotranspiration can be represented as a steady-state process,
runoff occurs only by saturation-excess.

The water-quality component of the ACRU2000 model uses nitrogen and phosphorus cycling algorithms from the Groundwater Loading Effects of Agricultural Management Systems (GLEAMS) model as a result of a previous model expansion (Campbell et al. 2001). Following the second main objective of this work, the nitrogen and phosphorus algorithms of the model are evaluated for application to shallow water-table environments with modifications governing

- the effect of soil moisture on nutrient transformation rates,
- the extraction of nutrients into ponded/runoff water,
- the instantaneous, reversible sorption of P to soil particles.

**Contribution of this Work: Additions Made to the ACRU2000 Model**

The Java-based, object-oriented hydrologic model ACRU2000 was not developed for use in humid, shallow water-table environments. However, due to its flexible structure the ACRU2000 model was chosen as the platform to implement the approximate hydrologic model described above. Specific hydrologic, field-scale modifications to ACRU2000 include

- Addition of the standardized Penman-Monteith equation (Allen et al. 1998) for estimating daily reference potential evapotranspiration,
- estimation of incoming solar radiation using the methods of Hargreaves and Samani (1982) and Samani (2000) when observations are unavailable,
- expansion of the number of soil layers represented by the model,
- explicit representation of ponded water on the ground surface,
- evaporation from ponded water,
• changes in the transpiration response to excess and limited soil moisture conditions,
• addition of a closed-form root distribution function,
• representation of upward gradients and flow between the water-table and the plant root zone,
• representation of a depth-variable specific yield using closed-form soil moisture characteristic equations to determine water-table depths and soil moisture contents,
• movement of groundwater to and from time-variable boundary conditions (both vertically and horizontally), and
• addition of a simple stage-discharge relationship to route runoff from the land surface.

Hydrologic processes retained from the original, unmodified ACRU2000 model include
• The ability to estimate reference potential evapotranspiration using a variety of methods,
• the choice of applying evaporative demand to the soil as a lumped quantity or as separate soil evaporation and plant transpiration,
• the interception of rainfall, and
• evaporation of intercepted water.

As mentioned, the water-quality component of ACRU2000 has been adapted, almost entirely, from the GLEAMS model. GLEAMS was developed to simulate edge-of-field and bottom-of-root-zone loadings of water, sediment, pesticides, and nutrients (Knisel et al. 1993). However, GLEAMS being a predominantly “upland” model may not respond appropriately to the moisture regime and soil conditions seen in shallow water-table environments such as the Florida flatwoods. For this reason the following modifications were made:

• maximum rates of nutrient transformations occur over specific ranges of soil moisture,
• mineralization and immobilization processes may occur under saturated or near-saturated conditions, but at a depressed rate,
• nutrients within ponded water are represented explicitly,
• the movement of nutrients from soil to ponded water occurs via mixing and in response to concentration gradients, and
• phosphorus partitioning coefficients are predicted based on factors that control P sorption in flatwoods soils.

**Organization of this Dissertation**

The representation of field-scale hydrologic processes by the (unmodified) ACRU2000 model are detailed in Chapter 2 of this document. Also in this chapter the validity of applying ACRU2000 to humid, shallow water-table environments is discussed.

In Chapter 3 of this document, the shallow water-table hydrologic modifications are presented. The model is developed by integrating a vadose zone component that uses an approximation of Richards’ equation, an evapotranspiration component that represents plant response to soil moisture conditions and accounts for upward gradients in the vadose zone, a Variable-Source-Area (VSA) runoff generation component, and a horizontal groundwater flow component.

In Chapter 4 the shallow water-table model proposed in Chapter 3 is validated for three experimental sites in the southeastern United States and its performance is compared to the Field Hydrologic And Nutrient Transport Model (FHANTM) and the original, unmodified ACRU2000 as described in Chapter 2. The first experimental field site was a wet prairie community within Paynes Prairie State Preserve in north-central Florida, the second, a dairy pasture in south-central Florida, and the third a beef cattle pasture in south-central Florida. The model performance is evaluated by comparing observed water-table depths, soil moisture contents, evapotranspiration, and runoff volumes to field observations.
In Chapter 5 of this document the N and P module of the ACRU2000 model are presented and its appropriateness for shallow water-table environments is discussed. In Chapter 6 modifications are proposed to the N and P module for shallow water-table environments and flatwoods soils.

In Chapter 7 the N and P module developed in Chapter 6 is evaluated using field observations from experimental pastures in south-central Florida. The model is also compared to the unmodified algorithms as described in Chapter 5. The model performance is evaluated by its ability to predict nitrogen and phosphorus loads in runoff.

In Chapter 8 the main findings from this study are summarized and recommendations for future work are made.

The appendices of this dissertation provide documentation for future model users and developers. Appendix A gives a list and short description of the hydrologic process and data objects added to the ACRU2000 model in the course of this work. Appendix B shows the Unified Modeling Language (UML) design diagrams for the hydrologic processes added to the model. Appendix C is an input and output variable reference that describes the new hydrologic input and output variables for future model users. Appendix D is a technical manual that details the workings of the hydrologic model. This appendix replicates parts of Chapter 3; however it refers to parameters as they are referenced in the input/output reference (Appendix C) and in the original, unmodified ACRU2000 model (Smithers and Schulze 1995) as well as providing some guidance for users in input parameter determination. Appendices E, F, and G cover the nitrogen and phosphorus model in a similar manner as for the hydrologic model. Appendix E is a short list and description of the new process and data objects, Appendix F presents the
UML diagrams of the objects, and Appendix G is a technical manual. Appendices H, I, J, and K detail a module for simulating the transport of a conservative solute (not implemented in this work). Appendix H is a short list and description of the new process, data, and interface objects; Appendix I shows the UML diagrams of the objects, Appendix J is an input/output variable reference, and Appendix K is a technical manual.
Figure 1-1. Flatwoods regions of Florida (adapted from United States Department of Agriculture Natural Resources Conservation Service [USDA/NRCS] 2002)
CHAPTER 2
FIELD-SCALE HYDROLOGY OF THE ACRU2000 MODEL

Introduction

The name ACRU began as an acronym for the Agricultural Catchment Research Unit of the Department of Agricultural Engineering (now the School of Bioresources Engineering and Environmental Hydrology) at the University of KwaZulu-Natal in Pietermaritzburg, South Africa. The model, recently redesigned into an object-oriented framework (Clark et al. 2001; Kiker and Clark, 2001a) and adopting the ACRU2000 moniker, operates on a daily time-step, uses a two-layer soil (referred to as the “A” and “B” soil horizons) to represent the water budget of a field or catchment, and can be operated as either a lumped field-scale or a distributed basin-scale model. This chapter details the field-scale hydrologic processes of the ACRU2000 model and discusses its suitability for shallow water-table environments. A detailed description of the entire ACRU model can be found in Schulze (1995) and Smithers and Schulze (1995). The structure and design of the model are presented in the next chapter in the context of the developments made in this work.

Evapotranspiration

The calculation of reference potential evapotranspiration in ACRU2000 can be determined by a variety of methods or inputted directly to the model (Schulze 1995). Reference potential evapotranspiration can also be determined using daily or average monthly meteorologic parameters. Calculation methods include the Penman (1948)
equation, the Hargreaves and Samani (1982; 1985) equations, the Blaney and Criddle (1950) equation, the Thornthwaite (1948) equation, and others (Schulze 1995).

Evaporative demand is applied in a top-down approach in ACRU2000; it is applied first to previously intercepted water on the plant canopy with the remaining demand being applied to soil evaporation and plant transpiration using Ritchie’s (1972) method or as a lumped quantity. When partitioned, potential transpiration \( T_p \) is estimated as a function of the leaf area index (\( LAI \)):

\[
T_p = (0.7 LAI^{0.5} - 0.21) ET_0 \quad \text{for} \quad LAI < 2.7 \quad (2-1a)
\]

\[
T_p = 0.95 ET_0 \quad \text{for} \quad LAI \geq 2.7 \quad (2-1b)
\]

with the remaining demand applied as potential soil evaporation, \( E_p \) to the A soil horizon. Potential soil evaporation \( E_p \) is adjusted for the percent surface cover by mulch or litter, \( C_s \):

\[
E_p = E_p \frac{C_s}{100} \quad (2-2)
\]

According to Ritchie’s (1972) method, actual evaporation from the soil surface continues at a maximum rate equal to the potential rate (Stage 1 evaporation) until the accumulated soil water evaporation exceeds the stage 1 upper limit, \( U_1 \), which is defined in units of mm:

\[
U_1 = (\alpha_s - 3)^{0.42} \quad (2-3)
\]

where \( \alpha_s \) is a soil water transmission parameter that is related to the texture of the soil (Ritchie 1972). After \( U_1 \) is exceeded soil water evaporation proceeds at a reduced (Stage 2) rate as a function of the square root of time:
\[ E = \alpha_s t_d - (t_d - 1)^{0.5} \]  
(2-4)

where \( t_d \) is the number of days since \( U_1 \) has been exceeded.

Transpiration (or lumped evapotranspiration) occurs in proportion to the user-specified fraction of roots contained in the two soil horizons and is adjusted using a crop coefficient. The reduction of transpiration in response to water excess or deficiency is assumed to occur at water contents above field capacity (water excess), taken as 100 cm of suction in the ACRU2000 model, or below a user-defined fraction of plant available water (water deficiency) where plant available water is defined as the water stored between field capacity and the wilting point (Figure 2-1).

**Interception**

The interception of rainfall by the plant canopy is represented as either a user-defined maximum storage capacity (mm) or as a function of the leaf area index and gross daily rainfall using the model of Von Hoyingen-Huene (1983) as cited in Schulze (1995):

\[ I = 0.30 + 0.27P + 0.13LAI - 0.013P \cdot LAI - 0.007LAI^2 \]  
(2-5)

where \( I \) is in units of mm and \( P \) is gross daily rainfall (mm).

**Infiltration and Runoff**

Runoff and infiltration are determined using a modified SCS curve number method (USSCS 1972; Schulze 1995):

\[ Q = \frac{(P_n - cS)^2}{P_n + (1-c)S} \]  
for \( P_n > cS \)  
(2-6a)

\[ Q = 0 \]  
for \( P_n \leq cS \)  
(2-6b)

where \( Q \) is the depth of runoff (mm), \( P_n \) is net rainfall (mm) (rainfall less intercepted water), \( c \) is a coefficient of abstraction (considered constant at 0.2 in the original SCS equation), and \( S \) is the potential maximum retention (mm). The term \( cS \) is the initial
abstraction, at $P_n$ values below which no runoff will occur. The coefficient of abstraction, $c$, is a user-supplied monthly value and may be as high as 0.4 immediately after plowing or under forested conditions or a low as 0.05 in regions of compacted soils according to Schulze (1995). As a variable parameter the coefficient of abstraction gives the user the ability to vary the runoff response to reflect different vegetation, site conditions, and management practices. The maximum retention, $S$, is determined from a soil water deficit prior to a rainfall event down to a user-defined critical soil depth, $D_c$:

$$S = \sum (\theta_s - \theta)D_c$$  \hspace{1cm} (2-7)

where $\theta_s$ and $\theta$ are the water content at saturation and the current water content, respectively. The critical soil depth varies with climatic, vegetative, and soil characteristics (Schulze 1995). A location with sparse vegetation, thin soils, and intense rainfall might have a relatively low value and a location of dense vegetation, deep soils, and low-intensity rainfall a high value (Figure 2-2). The ability to vary the critical soil depth, as well as the coefficient of abstraction, is an attempt in the ACRU2000 model to account for different runoff-producing mechanisms (Schulze 1995).

The runoff generated using equation 2-6 is routed from the field by specifying a “quickflow” fraction that runs off on the same day it was generated. The remainder is retained to the following day; however it is not available for infiltration or evaporation. This “delayed stormflow” is intended to act as a surrogate for interflow according to Schulze (1995) and the specified quickflow fraction only affects the timing of runoff, not the amount generated.
Percolation and Soil Moisture Redistribution

Downward percolation from a soil horizon can occur when the horizon is above field capacity according to a user-specified fraction (mm/day):

\[ q = K_{AB} (\theta - \theta_{fcA})d \]  

(2-8)

where \( K_{AB} \) is the fraction of water above field capacity in the A horizon that will drain to the B horizon, \( \theta_{fcA} \) is the water content of the A horizon at field capacity, and \( d \) is the thickness of the A horizon. An identical relationship is used for percolation out of the B horizon. Soil moisture may (as an option selected by the user) move in response to gradients at moisture contents below field capacity as a function of the gradient between the horizons downward (mm/day):

\[ q = 0.02 \theta_A \cdot d \left( \frac{\theta_A}{\theta_{fcA}} - \frac{\theta_B}{\theta_{fcB}} \right) \]  

(2-9)

and upward:

\[ q = 0.01 \theta_B \cdot d \left( \frac{\theta_B}{\theta_{fcB}} - \frac{\theta_A}{\theta_{fcA}} \right) \]  

(2-10)

where \( \theta_A \) and \( \theta_B \) are the moisture contents of the A and B horizons, \( \theta_{fcB} \) is the moisture contents of the B horizon at field capacity, and \( d \) is the thickness of the soil layer.

Baseflow

Water that percolates out of the B horizon enters the groundwater store. Water within the groundwater store can flow out to a nearby surface water body as baseflow. Baseflow is calculated as a function of the size of the groundwater store by assuming that a fraction of the groundwater store is released. The baseflow release fraction, \( F_{bf} \), is a
function of the size of the groundwater store, $S_{gw}$ (mm), and a base release coefficient, $F_{bf}$:

\[
F_{bf} = \begin{cases} 
0.8F_{bfi} & \text{for } S_{gw} < 0.015 \\
1.3F_{bfi} & \text{for } S_{gw} > 0.100 \\
F_{bfi} & \text{for } 0.015 \leq S_{gw} \leq 0.100 
\end{cases} \quad (2-11a)
\]

Application of the ACRU2000 Model to Shallow Water-Table Environments

The field-scale hydrology of the ACRU2000 model, as described above, was developed principally for arid locations where runoff generation occurs primarily via an infiltration-excess mechanism and the interaction of groundwater with nearby water bodies can be approximated as a one-way process. This is evidenced by

- The reduction of evapotranspiration at water contents above field capacity,
- the use of an SCS-type equation to estimate surface runoff,
- the assumption that groundwater flows only out from a catchment as baseflow.

While the reduction of evapotranspiration above field capacity and the inability to represent the flow of groundwater into the field or catchment from a nearby waterbody can be considered shortcomings of the model, the modeling approach of ACRU2000 does provide some flexibility in applying it to shallow water-table environments. This flexibility is entirely due to the ability to vary the parameters that determine runoff and infiltration (the coefficient of abstraction, $c$, and the critical soil depth, $D_c$). In applying the ACRU2000 model to locations with highly permeable, shallow water-table soils there are a few recommendations that can be made for parameterization of the model:

- The total soil depth represented in the model should be as deep as the deepest water-table observation,
- the response fraction for percolation from the B horizon to the groundwater store should be set to zero in order to mimic the poorly drained conditions caused by a shallow water-table,
the response fraction for percolation from the A to B horizon should be set to a value of 1.0 in the case of highly permeable soils. This allows the A horizon to drain to field capacity if sufficient storage is available in the B horizon where it can accumulate, mimicking the presence of a shallow water-table,

- plant roots must extend into the B horizon in order for the moisture contained within it to be available for evapotranspiration (as water may only move up from the B horizon at moisture gradients below field capacity),

- the coefficient of initial abstraction should have a value of 1.0. This will cause runoff to only occur when the entire soil profile is saturated,

- the critical soil depth from which the soil moisture deficit is calculated should be the entire depth of the soil.

As an alternative to allowing plant roots to extend into the B horizon and setting the response fraction from the A to B horizon to unity, the A horizon can be assumed to extend to the entire depth of the soil and contain all of the plant roots. Due to the reduction of evapotranspiration above field capacity the splitting of the soil between two horizons is arbitrary and depending on their relative thickness will produce greatly varying results when assuming no drainage out of the B horizon.

**Summary**

This chapter details the field-scale hydrology as simulated in the ACRU2000 model and its suitability for use in shallow water table environments. In the model, reference potential evapotranspiration can be determined by a variety of methods and is applied in a top-down approach to intercepted water and then to soil evaporation and plant transpiration. The model does not represent water ponded on the ground surface. Rainfall can be intercepted by the plant canopy, as represented by two different methods, and is partitioned between runoff and infiltration using a modified SCS curve number method. Water may percolate out of a soil horizon at water contents above field capacity and may move between soil horizons at water contents below field capacity. Water
percolating out of the B soil horizon is added to the groundwater store from which baseflow can occur according to a user-defined baseflow fraction and the size of the groundwater store. Runoff that is generated is split between a quickflow fraction, occurring on the day generated, and a delayed stormflow fraction.

The ACRU2000 model may not be appropriate for application to shallow water-table environments due to the reduction of evapotranspiration (or plant transpiration) at water contents above field capacity, the use of an infiltration-excess type procedure to determine runoff, and the inability of the model to represent groundwater flow into the model domain. However, the flexibility of the modified curve number procedure of the model offers some flexibility in representing the different runoff producing mechanisms of infiltration-excess and saturation-excess. A field dominated by saturation-excess runoff may be sufficiently represented by using parameter values that are outside of the recommended ranges.

Modifications to ACRU2000 that may be considered to be more appropriate for shallow water-table environments are described in the next chapter. This modified model and the original model described here are evaluated against field data from shallow water-table environments in Chapter 4.
Figure 2-1. Transpiration reduction due to soil moisture excess and deficiency as simulated in the ACRU2000 model (adapted from Schulze 1995)

Figure 2-2. Critical soil depth as related to climatic, vegetative, and soil conditions (adapted from Schulze 1995)
CHAPTER 3
FIELD-SCALE HYDROLOGIC MODEL FOR HUMID, SHALLOW WATER-TABLE ENVIRONMENTS: DEVELOPMENT

Introduction

Simulation of humid regions with highly permeable, shallow soils has been shown to be inconsistent with the concept of infiltration excess or “Hortonian” runoff generation (Dunne and Black 1970; Freeze 1972; Jayatilaka and Gillham 1996; Ogden and Watts 2000; Hernandez et al. 2003). In such environments, runoff is typically generated by saturation excess whereby the water-table rises to the ground surface, creating a Variable Source Area (VSA), a zone of saturation that expands and contracts seasonally as well as during individual storms. These VSAs often form where subsurface lateral flow converges, the ground slope changes, or the depth to a restrictive layer decreases (Frankenberger et al. 1999). Regions dominated by VSA runoff include much of the southeastern coastal plain of the United States, and the flatwoods regions of Florida, in particular. The flatwoods landscape is characterized by very flat topography with moderately to poorly drained, highly permeable, sandy soils that can often have standing water during wet weather. As a result, groundwater levels are heavily influenced by rainfall, evapotranspiration (ET), and nearby canal or stream stages (Yan and Smith 1994; Dukes and Evans 2003).

Management of surface and groundwater quality has become an environmental priority, particularly in agricultural watersheds. In managing water-resource quantity and quality, modeling is the most cost-effective way to evaluate the impact of management
alternatives. In the past, model development efforts often emphasized land-surface
processes or groundwater processes, but rarely both. The land-surface models emphasize
infiltration, ET, and surface-water movement while often ignoring or oversimplifying
saturated groundwater, and the groundwater models emphasize prediction of groundwater
levels in response to pumping, boundary conditions, and recharge while oversimplifying
ET and vadose zone processes. Both surface water and groundwater models work well
on their own, when used in areas where the interaction between surface and groundwater
is weak or insignificant (Yan and Smith 1994).

Several models have been developed and tested for use in flatwoods regions
including CREAMS-WT (Heatwole et al. 1987; Heatwole et al. 1988) based on the
CREAMS model (Knisel 1980); EAAMOD (Bottcher et al. 1998a); FHANTM (Tremwel
and Campbell 1992; Fraisse and Campbell 1996) based on the DRAINMOD model
(Skaggs 1980); and FLATWOODS (Sun et al. 1998) based on the MODFLOW model
(McDonald and Harbaugh 1988). All of these models were developed as field-scale
models with the exception of FLATWOODS, which happens to be the only one of these
models that (to date) does not contain a water quality component. The field-scale models
mentioned have been incorporated into distributed models (Heatwole et al. 1986;
Negahban et al. 1995; Bottcher et al. 1998b); however, they are used to determine “edge
of field” effects and, as such, are used in a “loose coupling” framework where individual
fields (or grid cells) do not interact with one another. Sun et al. (1998) showed the need
for distributed, fully-interactive modeling of flatwoods systems where groundwater
gradients may be strongly affected by heterogeneous distributions of vegetation type,
seasonality effects, and changing management practices.
To better simulate high water-table environments like the coastal plain flatwoods, new model components have been developed within the framework of the ACRU2000 model (Clark et al. 2001; Kiker and Clark 2001a). The new components allow for distributed, physically based modeling of high water-table environments. The objective of this work was to develop field-scale model components for the flatwoods landscape for use in the ACRU2000 distributed hydrologic model; and to demonstrate the advantages of adding model components within the object-oriented framework of the ACRU2000 model.

**Model Development**

**Background**

The ACRU model (originally written in the FORTRAN programming language) has its origins in a distributed catchment evapotranspiration study in the Natal Drakensberg region of South Africa in the early 1970s (Schulze 1995). Since then the model has undergone many revisions and additions to meet the water-related needs of the scientific-modeling community in South Africa and beyond. However, each consecutive improvement to the model has created a more difficult design and coding challenge for subsequent researchers. The many contributions made to the model over the years resulted in a framework in which it was relatively difficult to make new additions, and in some instances the model structure was unable to accommodate the desired additions at all. To better accommodate future model additions, the ACRU model was recently redesigned in an object-oriented framework (Clark et al. 2001; Kiker and Clark 2001a).

As mentioned in the previous chapter, the ACRU2000 model can be used as either a lumped field-scale model or as a distributed basin-scale model. The model operates on a daily time-step, using a modified SCS curve number procedure to generate daily runoff.
volumes (Schulze 1995). The model uses a two-layer soil to represent the water budget of the catchment, with any water above field capacity percolating out of a layer according to a user-defined fraction. Water draining out of the bottom soil layer enters the groundwater store, from which baseflow is generated as a function of the size of the store and a user-defined baseflow coefficient. Plant-canopy interception can be represented using two different methods and this intercepted water can (in turn) be evaporated back into the atmosphere. Reference potential evaporation can be determined using a variety of methods and applied as a lumped quantity, or it may be partitioned between soil evaporation and plant transpiration, according to the method of Ritchie (1972). Plant water-stress, and a corresponding reduction in transpiration, is determined to occur at some water content between field capacity and wilting point (set by the user) for water-limiting conditions, and is determined to occur at water contents above field capacity for water-excess conditions. Schulze (1995) and Smithers and Schulze (1995) give a more detailed description of the ACRU model.

Structure

The model was redesigned with the belief that the hydrologic system is complex, and thus the way we view it or model it should be seen as a learning process that may require periodic reevaluation. The model was restructured using an object-oriented methodology to produce a more flexible and extensible model structure. The new, object-oriented model is referred to as ACRU2000. The model was designed using the Unified Modeling Language (UML) and the model was implemented in the Java programming language. Using UML allows graphic design of objects with diagrams for object-oriented programming, before writing computer code. The UML provides a standardized notation to specify, design, visualize, and document object-oriented
software (Jacobsen et al. 1998). The extensibility of the ACRU2000 model framework has been demonstrated previously by Campbell et al. (2001), who added a module to simulate nitrogen and phosphorus transport and transformations, and by Kiker and Clark (2001b), who added a module to simulate southern African rangeland ecosystems.

Object orientation uses the concept of objects, where an object consists of a small, well-written piece of computer code that contains its own attributes, methods, and behavior (Figure 3-1). The attributes describe the object, in terms of physical characteristics or other traits. The methods describe the object’s internal functionality, and contain the equations the model uses to simulate various events. The behavior of the object describes how it interacts with other objects. Object orientation thus encourages the creation of models that are modular in structure. Three main object types in ACRU2000 are of interest to the researcher simulating environmental events: Component, Data, and Process objects. Component objects are physical components of the system, such as the climate, the soil, or a soil layer. Data objects are the descriptors, or attributes, of the Component such as the temperature, depth, or hydraulic conductivity. These data attributes of the Component objects are modeled as separate objects themselves, because as an object they can be reused or extended (rather than just being a simple variable). As an object, only a single Data object representing a specific trait needs to be written in code. This single object can then be associated with different objects (with different values) representing the particular trait of each Component object. Process objects, the third type of object of interest to the researcher, represent the action or event that involves one or more Component objects, such as interception of rainfall by a vegetation canopy or the infiltration of water into the soil. Each Process object, when
acting on one or more Component objects, uses the Data objects associated with that Component (Figure 3-1) (Clark et al. 2001; Kiker and Clark 2001a).

Objects can interact with each other in three different ways: inheritance, aggregation, and association (Figure 3-1). Objects may inherit properties from other objects. Inheritance indicates that one object is a “type of” another object. This inheritance of functionality between objects allows for the code that has already been written for the “parent” object to be used by the “child” object. Any difference between the two objects is made as needed in the “child” object. In this manner the “child” object can be a more specialized version of the “parent” object. Objects may also be aggregated, or be a “part of” other objects. In this role, one object may use the functionality of another. Another way that objects may interact is by association (using information from other objects). This “uses data from” relationship allows an object to access data that is “owned” by another object. These three relationships among objects encourage the development of modular code, and result in a flexible and extensible programming structure that encourages code re-use.

The structure of the object-oriented design of ACRU2000 allows new objects to be created and linked to the model without major revision to the existing code. To add a new module (a group of objects with a common overall purpose) the model developer needs to:

- Identify the Component, Process, and Data objects which will be used.
- Determine any new objects to be created.
- Define the relationships of all of the objects in the new module and to existing objects.
- Implement the design in Java code within the framework set in Steps 1-3.
Steps 1-3 are accomplished easily using UML. The developer only needs to write computer code at Step 4.

Adding the new hydrology module to ACRU2000 focuses on adding new Process objects, and to a lesser extent on adding any needed Data objects. This is because the Component objects (the climate, soil, soil layers, etc.) already exist, along with the Data objects associated with them. Each new Process object has a UML diagram associated with it (Figure 3-1). The use of UML as the standardized design tool provides a design that is easy to understand.

The following section describes the field-scale hydrologic processes of the shallow water-table module of ACRU2000. The processes that have been retained from the original model are noted. A short description of the process and data objects added to the model can be seen in Appendix A, UML design diagrams of the hydrologic processes can be seen in Appendix B, and Appendix C provides and input/output variable reference for future users of the model.

**Hydrologic Processes and Governing Equations**

The ACRU2000 model was originally developed for use in upland watersheds that are characterized by infiltration-rate limited (Hortonian) runoff generation and topographic gradients that drive overland and groundwater flow direction. To apply the model to humid, shallow water-table environments, several modifications were needed. These included the effects of a shallow water-table on evapotranspiration and runoff generation, a depth-variable specific yield, and the representation of surface water and groundwater gradients that may reverse in response to time-variable boundary conditions. In making these modifications the model was expanded to simulate up to ten soil layers.
Evapotranspiration

In addition to the methods previously included in the model for calculating reference potential evapotranspiration ($ET_0$), the standardized Penman-Monteith equation for grass reference potential evapotranspiration adopted by the Food and Agricultural Organization in the FAO Irrigation and Drainage Paper No. 56 (Allen et al. 1998) has been added to the model:

$$ET_0 = \frac{0.408\Delta(R_n - G) + \gamma_p \frac{T_{\text{mean}} + 273}{\Delta + \gamma_p (1 + 0.34u_2)} u_2 (e_s - e_a)}{900}$$

(3-1)

where $ET_0$ is in units of mm day$^{-1}$, $R_n$ is the incoming net radiation (MJ m$^{-2}$ day$^{-1}$) and is the difference between net incoming shortwave radiation, $R_{ns}$ and the net outgoing longwave radiation, $R_{nl}$, $G$ is the soil heat flux density (MJ m$^{-2}$ day$^{-1}$) and is assumed to be zero for daily calculations, $T_{\text{mean}}$ is the mean daily air temperature at 2 m height $[(T_{\text{max}} + T_{\text{min}})/2, ^\circ\text{C}]$, $u_2$ is the wind speed at 2 m height (m/s), $e_s$ is the saturated vapor pressure (kPa), $e_a$ is the actual vapor pressure (kPa), $\Delta$ is the slope of the vapor pressure curve (kPa/°C), and $\gamma_p$ is the psychrometric constant (kPa/°C). The net shortwave radiation, $R_{ns}$ is determined from the albedo, $\alpha$ (assumed to be 0.23) and the incoming shortwave radiation, $R_s$:

$$R_{ns} = (1 - \alpha)R_s$$

(3-2)

$R_s$ is supplied as daily input to the model or estimated from the Angstrom equation:

$$R_s = \left(a_s + b_s \frac{n}{N}\right)R_a$$

(3-3)

where $n$ is the number of sunshine hours, $N$ is the maximum possible number of sunshine hours, $R_a$ is the extraterrestrial radiation (MJ m$^{-2}$ day$^{-1}$), $a_s$ is a regression constant that
expresses the fraction of extraterrestrial radiation reaching the earth on overcast days \((n = 0)\), and the quantity \(a_s + b_s\) is the fraction of extraterrestrial radiation reaching the earth on clear days \((n = N)\). In the absence of regional values for \(a_s\) and \(b_s\), \(R_s\) may be estimated using the equation of Hargreaves and Samani (1982):

\[
R_s = (KT) \left( R_a \right) (TD)^{0.5}
\]

where \(TD\) is the difference between maximum and minimum air temperatures (°C) and \(KT\) is an empirical constant. Samani (2000) developed an equation to determine \(KT\) as a function of \(TD\):

\[
KT = 0.00185(TD)^2 - 0.0433TD + 0.4023
\]

using 25 years of data for the continental U.S. Daily values for \(\Delta\), \(e_s\), \(e_a\), \(\gamma_p\), \(N\), \(R_{as}\), and \(R_{nl}\) are calculated according to the equations given in Allen et al. (1998).

Evaporative demand is applied in a top-down approach in the model, with evaporation applied first to intercepted water, then to ponded water on the ground surface, then to soil evaporation and plant transpiration. As in the original ACRU2000 model (Chapter 2), soil evaporation and plant transpiration can be applied as a lumped quantity to soil layers containing plant roots or separated between soil evaporation and plant transpiration for conditions of incomplete cover using the methods of Ritchie (1972) where potential transpiration \((T_p)\) is estimated as a function of the leaf area index \((LAI)\):

\[
\begin{align*}
T_p &= \left(0.7LAI^{0.5} - 0.21\right)ET_0 \quad \text{for} \quad LAI < 2.7 \\
T_p &= 0.95ET_0 \quad \text{for} \quad LAI \geq 2.7
\end{align*}
\]

Plant transpiration is further adjusted for different plant albedo, and stomatal and aerodynamic resistances at various stages of growth with a crop coefficient. Water
extraction from each soil layer is determined as a function of the maximum transpiration rate $T_p$, a root distribution function $g(d)$, and soil moisture dependent reduction factor $\alpha(h)$:

$$ T = \alpha(h) g(d) T_p $$

(3-7)

The soil moisture reduction factor as a function of pressure head, $\alpha(h)$ used was proposed by Feddes et al. (1978) (Figure 3-2). At $h < h_1$ (oxygen deficiency), and $h > h_4$ (pressure greater than the wilting point) $\alpha(h)$ is zero. Between $h_2$ and $h_3$ transpiration occurs at the potential transpiration rate, $\alpha(h)$ is unity. Between $h_1$ and $h_2$ and $h_3$ and $h_4$ a linear reduction in transpiration is assumed. The ability to alter the point in which oxygen deficiency begins ($h_2$) is advantageous since many plant species in the flatwoods ecosystem are quite tolerant to wet conditions and may transpire at their maximum rate even when the soil is saturated. The linear root distribution function used here is that proposed by Hoogland et al. (1981):

$$ g(d) = \frac{e(2d - L) + L}{L^2} \quad -1 \leq c \leq 0, \quad d \leq L $$

(3-8)

where $c$ is an empirical parameter expressing the relative density of roots between the ground surface ($d = 0$) and the maximum depth of roots ($d = L$) and is shown in Figure 3-3. Upon integration with respect to $d$, Equation 3-8 defines the fraction of roots between the ground surface and a depth of $d$.

Potential soil evaporation is adjusted by a factor of 1.15 to account for the difference in albedo between bare soil and a vegetated surface as recommended by Allen et al. (1998). As in Chapter 2, potential soil evaporation $E_p$ is further adjusted for the percent surface cover by mulch or litter, $C_s$: 
\[ E_p = E_p \frac{C_s}{100} \]  

Soil water evaporation takes place to a user defined depth within the soil profile with recommended values ranging from 0.1 to 0.15 m depending on soil texture (Allen et al. 1998). According to Ritchie’s method actual evaporation from the soil surface continues at a maximum rate equal to the potential rate (Stage 1 evaporation) until the accumulated soil water evaporation exceeds the stage 1 upper limit, \( U_1 \), which is defined in units of mm:

\[ U_1 = (\alpha_s - 3)^{0.42} \]  

where \( \alpha_s \) is a soil water transmission parameter that is related to the texture of the soil (Ritchie 1972). After \( U_1 \) is exceeded soil water evaporation proceeds at a reduced (Stage 2) rate as a function of the square root of time:

\[ E = \alpha_s t_d - (t_d - 1)^{0.5} \]  

where \( t_d \) is the number of days since \( U_1 \) has been exceeded.

**Interception**

The interception of rainfall by the plant canopy is represented as either a maximum storage capacity (mm) or as a function of the leaf area index and gross daily rainfall using the model of Von Hoyingen-Huene (1983) as cited in Schulze (1995) and used in the unmodified ACRU2000 model:

\[ I = 0.30 + 0.27P + 0.13LAI - 0.013P \cdot LAI - 0.007LAI^2 \]  

where \( I \) is in units of mm and \( P \) is gross daily rainfall (mm).
**Infiltration**

In humid, shallow water-table regions the dominant mechanism of runoff generation has been shown to be saturation excess (Dunne and Black 1970; Freeze 1972; Hernandez et al. 2003). In these regions a shallow water-table rises, saturating the entire soil profile and inundating the ground surface, creating a Variable Source Area (VSA). These VSAs vary with space and time, expanding in the wet season and contracting in the dry season. In contrast to infiltration-rate limited or Hortonian runoff (Horton 1933), runoff from VSAs have shown little sensitivity to temporal variability of rainfall or rainfall intensity (Hernandez et al. 2003). Due to the highly conductive nature of the sandy soils of the Florida flatwoods the infiltration capacity of the soil is rarely, if ever, limiting. Thus runoff generation in the model is assumed to be solely storage-limited, with infiltration proceeding until the entire soil profile becomes saturated.

**Water-table depth and soil moisture distribution**

The total volume of water contained within pore spaces between the water-table \( z = 0 \) and the ground surface \( z = z_0 \) can be expressed as:

\[
V_w = \int_0^{z_0} \theta k d z \tag{3-13}
\]

where \( V_w \) is the volume of water per unit area (cm\(^3\) cm\(^{-2}\)) and \( \theta \) is the moisture content (cm\(^3\) cm\(^{-3}\)) at some height above the water-table \( z \) (cm). If the relationship between soil moisture and pressure head \( h \) within the soil profile is known and hydrostatic conditions can be adequately assumed, then closed form equations expressing \( \theta \) as a function of \( z \) can be used in equation (3-13) to define explicitly the volume of water between the ground surface and the water-table. In this hydrostatic approximation of Richards’ equation \( V_w \) is the basic state variable that is influenced by water movement into or out of
the soil profile. This assumption of a hydrostatic condition has been shown to be adequate for regions with shallow water-tables (Skaggs 1980; Koivusalo et al. 2000) and also specifically for shallow water-table regions with highly conductive soils such as the flatwoods regions of Florida (Rogers 1985). In general, the hydrostatic approximation is most accurate for higher conductivity soils and shallower water-tables. Under hydrostatic conditions the water content at any point can be described using the models of Brooks and Corey (1964), Hutson and Cass (1987), or van Genuchten (1980). For a soil profile at hydrostatic equilibrium the water content as a function of height above the water-table using the model of Brooks and Corey (1964) is

\[
\theta(z) = \theta_r + (\theta_s - \theta_r) \left( \frac{h_b}{z} \right)^{\lambda} \quad \text{for} \quad z > h_b
\]  

(3-14a)

\[
\theta(z) = \theta_s \quad \text{for} \quad z \leq h_b
\]  

(3-14b)

where \( \theta(z) \) is the moisture content (cm\(^3\) cm\(^{-3}\)) as a function of height above the water-table \( z \) (cm), \( \theta_r \) is the residual moisture content (cm\(^3\) cm\(^{-3}\)), \( \theta_s \) is the saturated moisture content (cm\(^3\) cm\(^{-3}\)), \( h_b \) is the bubbling pressure head, or air entry pressure head, of the soil (cm); and \( \lambda \) is the pore size distribution index (-). The model of Hutson and Cass (1987) which replaces the sharp discontinuity at \( h_b \) in the Brooks and Corey (1964) model with a parabolic segment can similarly be expressed as

\[
\theta(z) = \theta_r + (\theta_s - \theta_r) \left( \frac{h_b}{z} \right)^{\lambda} \quad \text{for} \quad z > h_i
\]  

(3-15a)

\[
\theta(z) = \theta_r + (\theta_s - \theta_r) \left[ 1 - \frac{h_b^2}{z^2 \left( 1 - \frac{\theta_i}{\theta_s} \right)^{2/\lambda}} \right] \quad \text{for} \quad z \leq h_i
\]  

(3-15b)
where $\theta_i$ is the water content (cm$^3$ cm$^{-3}$) at the inflection point between the exponential and parabolic portions of the water characteristic curve at capillary pressure head $h_i$ (cm). $\theta_i$ and $h_i$ are defined as

$$\theta_i = \frac{2\theta_s}{\lambda + 2} \quad (3-16)$$

$$h_i = h_b \left( \frac{\lambda + 2}{2} \right)^{\frac{1}{\lambda}} \quad (3-17)$$

Similarly the model of van Genuchten (1980) can be expressed as

$$\theta(z) = \theta_r + (\theta_s - \theta_r) \left[ \frac{1}{1 + (\alpha z)^n} \right]^m \quad (3-18)$$

where $\alpha$ is in units of cm$^{-1}$, and $n$, and $m$ are empirical parameters with the constraint $m = 1 - 1/n$. The volume of water within a given soil layer can then be expressed as the integral of either one of these models:

$$V_w = \int_{z_i}^{z_f} \theta(z) \, dz \quad (3-19)$$

Upon integration Equation (3-14) becomes:

$$\int \theta(z) \, dz = \theta_r z + \frac{(\theta_s - \theta_r) \left(\frac{h_b}{z}\right)^{\lambda}}{1 - \lambda} \quad \text{for } \lambda \neq 1 \text{ and } z > h_b \quad (3-20a)$$

$$\int \theta(z) \, dz = \theta_r z + (\theta_s - \theta_r) h_b \ln(z) \quad \text{for } \lambda = 1 \text{ and } z > h_b \quad (3-20b)$$

$$\int \theta(z) \, dz = \theta_s z \quad \text{for } z \leq h_b \quad (3-20c)$$

Similarly, upon integration Equation (3-15) becomes:

$$\int \theta(z) \, dz = \theta_r z + \frac{(\theta_s - \theta_r) \left(\frac{h_i}{z}\right)^{\lambda}}{1 - \lambda} \quad \text{for } \lambda \neq 1 \text{ and } z > h_i \quad (3-21a)$$

$$\int \theta(z) \, dz = \theta_r z + (\theta_s - \theta_r) h_i \ln(z) \quad \text{for } \lambda = 1 \text{ and } z > h_i \quad (3-21b)$$
The model of van Genuchten (1980), Equation (3-18), cannot be integrated analytically with the restriction that \( m = 1 - 1/n \) so it is integrated numerically in the model using a five-point Gauss-Legendre quadrature (Chapra and Canale 1998).

The removal of water from the plant root zone by evapotranspiration may cause a deviation from the hydrostatic profile. This deviation creates a depleted root zone that is represented explicitly in the model for each soil layer. This depleted root zone implies that an upward gradient is induced within the soil profile. Water may move upwards in the soil profile in response to this gradient. This upward movement of water defines the connectivity between a shallow water-table and evapotranspiration. This upward movement of water is simulated by assuming that a steady state condition exists between the water-table and an evaporating surface. Assuming steady-state, the upward movement of water can be found from Darcy’s Law, assuming the soil profile is homogeneous:

\[
\int \theta(z) \, dz = \theta(z) \, z + (\theta_s - \theta_r) \left[ \frac{z^2}{3h_i^2} \left( 1 - \frac{\theta_i}{\theta_s} \right) \left( \frac{\theta_s}{\theta_i} \right)^{2/\alpha} \right] \quad \text{for} \quad z \leq h_i \tag{3-21c}
\]

where \( q \) is the upward flux (m/day), \( K(h) \) is the hydraulic conductivity (m/day), \( h \) is the soil capillary pressure head, and \( z \) is the height of the evaporating surface above the water-table (m). Integration of Equation (3-22) yields:
\[
\int_{0}^{h} \frac{dh}{1 + q / K(h)} = z
\]

Assuming a relationship between \(K\) and \(h\) allows Equation (3-23) to be solved with a lower boundary condition of \(h = 0\) at \(z = 0\) (at the water-table). The complexity of analytical or numerical solutions to equation (3-23) depends on the choice of \(K(h)\) relationship. In developing analytical solutions an upper limit of integration of \(h = \infty\) is typically used for simplicity. Gardner (1958) showed that this upper limit is appropriate since upward flux quickly approaches a limiting value as soil suction increases. Anat et al. (1965) solved equation (3-23) using the Brooks and Corey (1964) hydraulic conductivity relationship and derived an approximate, algebraic solution that is explicit in \(q\):

\[
q = K_s \left[ 1 + \frac{1.886}{\eta^2 + 1} \right]^\eta \left( \frac{z}{h_b} \right)^{-\eta} \quad (3-24)
\]

where \(\eta = 2 + 3 \lambda\). The actual amount of upward flux occurring during a time step is determined by using the first depleted soil layer above the water-table as the upper boundary. The maximum upward flux calculated to this bottom-most depleted layer is retained as the limiting maximum upward flux for the entire profile on a given day. Should this bottom-most depleted layer become fully replenished the calculation of upward flux proceeds to the next layer until either the limiting maximum is reached, or the maximum upward flux for a subsequent layer is less than the amount to which that layer is depleted below its hydrostatic water content, or the entire root zone is replenished, whichever is smaller. The importance of such a time-varying upper boundary condition for upward flow from the water-table when simulating the
fluctuations of the water-table, particularly during periods of drought, has been demonstrated by Rogers (1985) and Desmond et al. (1996).

**Groundwater flow**

Groundwater may flow into or out of the model domain both horizontally and vertically. Vertical movement of groundwater can occur at a constant or time-varying specified rate or according to a constant or time-varying hydraulic head in a deep aquifer below a restrictive layer according to Darcy’s Law:

\[ q = -C_r(H_{wt} - H_d) \]  

(3-25)

where \( H_{wt} \) is the hydraulic head in the surficial aquifer (m), \( H_d \) is the hydraulic head in the deep aquifer (m), and \( C_r \) is the conductance of the restrictive layer (1/day) and is defined as:

\[ C_r = \frac{K_r}{L_r} \]  

(3-26)

where \( K_r \) and \( L_r \) are the hydraulic conductivity (m/day) and thickness (m) of the restrictive layer, respectively.

Horizontal flow in response to a constant or time-varying boundary condition is simulated using the Dupuit equation (Fetter, 1994):

\[ q' = \frac{K_H}{2L} \left( H_{wt}^2 - H_b^2 \right) \]  

(3-27)

where \( q' \) is the flow per unit width (m²/day), \( K_H \) is the horizontal hydraulic conductivity, \( L \) is the distance to the boundary (m), and \( H_b \) is the hydraulic head at the boundary (m).
Overland flow and depression storage

Overland flow is simulated with a simple stage-discharge relationship as used by Tremwel and Campbell (1992) and Kroes and van Dam (2003). The stage-discharge relationship is of the form:

\[ q = \frac{1}{\gamma} (h_{\text{pond}} - z_{\text{dep}})^\beta \]

(3-28)

where \( q \) is the runoff depth (mm), \( \gamma \) is the runoff resistance (days) and is typically calibrated to observed data, \( h_{\text{pond}} \) is the depth of ponded water on the ground surface (mm), \( z_{\text{dep}} \) is the depth of depression storage which must be filled before runoff can begin (mm), and \( \beta \) is an exponent to be calibrated (-) but is usually given a value of 1.67 (assuming Manning’s equation).

Summary

A field-scale hydrologic module for use within the ACRU2000 distributed hydrologic model was developed to simulate the hydrology of humid, shallow water-table regions such as the flatwoods of the southeastern United States. The module is intended to simulate the position of the water-table explicitly in order to accurately predict its contribution to evapotranspiration and the creation of Variable Source Area runoff. The standardized Penman-Monteith reference potential evapotranspiration equation recommended by the Food and Agricultural Organization (Allen et al. 1998) has been added to the model as well as the methods of Hargreaves and Samani (1982) and Samani (2000) to estimate incoming solar radiation. Evaporative demand is applied in a top-down approach to intercepted water, ponded water, and then applied as a lumped quantity to the soil or partitioned to soil evaporation and plant transpiration. Lumped evapotranspiration or partitioned transpiration is applied in proportion to the density of
plant roots which is defined using the linear root distribution function of Hoogland et al. (1981). The response of plant transpiration or lumped evapotranspiration to water stress is represented as a function of soil water pressure head using the functional relationship of Feddes et al. (1978). The model approximates soil moisture as having a hydrostatic distribution. Soil moisture contents may be represented by one of three soil moisture characteristic models. Soil moisture within the root-zone may drop below the hydrostatic water content due to evapotranspiration. This reduction in root-zone water contents induces upward flow which is represented using an approximate, steady-state solution to Darcy’s Law. Saturated groundwater can flow in or out to a deep aquifer and/or horizontally in response to a time-varying boundary condition. Runoff is assumed to occur via saturation-excess only and moves from the field according to a simple stage-discharge relationship.

Application of the model, at the field-scale, is limited to the scale at which model parameters can be appropriately considered to be homogeneous and to areas with highly permeable soils where runoff occurs primarily by saturation-excess.

The object-oriented design of the ACRU2000 model made it an ideal candidate for adding such model components in a straightforward and consistent manner. Object design was made using UML to define new objects and their relationships. UML diagrams of the hydrologic module described here are shown in Appendix B and are accompanied by a short description of the objects in Appendix A, and input/output variable reference in Appendix C, and a technical manual in Appendix D. The resulting model design provides a modular and easily extensible model structure. The new
hydrologic module is validated in Chapter 4 and its performance compared to the original, unmodified model described in Chapter 2.
Figure 3-1. Sample UML diagram showing Component, Process, and Data objects and the inheritance, aggregation, and association relationships (adapted from Kiker and Clark 2001)
Figure 3-2. Transpiration reduction factor as a function of soil pressure head.

Figure 3-3. Root density distribution function $g(d)$ of Hoogland et al. (1981). Maximum depth of roots ($L$) is 80 cm in this example.
CHAPTER 4
FIELD-SCALE HYDROLOGIC MODEL FOR HUMID, SHALLOW WATER-TABLE ENVIRONMENTS: VALIDATION

Introduction

A physically based model for humid, shallow water-table environments was developed in Chapter 3, this chapter validates the model, evaluates the sensitivity of the model to input parameters, and makes recommendations for model improvement. The model validation is conducted using observed data from three experimental sites. For the first experimental site validation is made by comparing model simulations to observed data and to a numerical, finite difference model that solves Richards’ equation. For the second experimental site validation is made by calibrating the model to a portion of the observed data and then using the remaining observations for verification of the model calibration. The model’s performance is also compared to the original, unmodified ACRU2000 model (Chapter 2) and the field-scale Field Hydrologic And Nutrient Transport Model (FHANTM) for the second experimental site. For the third experimental site validation is made by calibrating the model to a portion of the observed data and then using the remaining observations for verification of the model calibration.

In addition to evaluating the model’s performance by making visual comparisons between simulated results and field observations the model performance is evaluated using error measures of mean absolute error:

\[
MAE = \frac{1}{n} \sum_{i=1}^{n} |P_i - O_i|
\]

(4-1)

and root-mean square error:
by pair-wise comparison between model predicted \( (P_i) \) and observed \( (O_i) \) daily values.

These measures are significant in that they provide a quantification of the error in units of the variable in question. Using \( RMSE \) in conjunction with \( MAE \) is useful in that the degree to which \( RMSE \) exceeds \( MAE \) is an indicator of the extent that outliers or variance in the difference between simulated and observed values exist in the data since \( RMSE \) is more sensitive to extreme values due to the squaring of the differences between observations and predictions (Legates and McCabe 1999). Additionally, model performance is evaluated using a relative error measure, the Nash-Sutcliffe coefficient of efficiency (Nash and Sutcliffe 1970):

\[
E = 1 - \frac{\sum_{i=1}^{n} (P_i - O_i)^2}{\sum_{i=1}^{n} (O_i - \bar{O})^2}
\]  

(4-3)

where \( \bar{O} \) is the mean of the observations. \( E \) ranges from negative infinity to 1.0 with higher values indicating better agreement. Since \( E \) is a ratio of the mean square error to the variance of the observed data, subtracted from 1.0, \( E \) is equal to 0.0 if the squares of the differences of predicted and observed values is as large as the variability in observed data. This indicates that the observed mean \( \bar{O} \) is as good a predictor as the model. A value of \( E < 0.0 \) indicates that \( \bar{O} \) is a better predictor (Legates and McCabe, 1999).
Case Study 1: Paynes Prairie State Preserve

Site Description and Experimental Design

Paynes Prairie State Preserve is a 5600 ha highland marsh system in north-central Florida (Figure 4-1). Approximately 4100 ha of the Preserve is wetland and is a surface expression of the surficial aquifer. Experimental data were collected by Jacobs et al. (2002) from a wet prairie community within the Preserve (29°34'14"N, 82°16'46"W). Detailed site description and instrumentation information can be found in Jacobs et al. (2002) and Whitfield (2003), however a brief summary is included here. The wet prairie is a flat plain with emergent, herbaceous species such as maiden cane (*Panicum hemitomon* Schultes), mild water-pepper (*Polygonum hydropiperoides* Michx.), mock bishop’s weed (*Ptilimnium capillaceum* Michx), and dog fennel (*Eupatorium capillifolium* Lam.) (Jacobs et al. 2002). Observations by Jacobs et al. (2002) showed that the majority of plant roots were contained in the top 10 cm of soil with 95% of roots within the top 25 cm. The marsh hydrogeology consists of a sandy surficial aquifer that is separated from the Floridan aquifer by the Hawthorne Formation, a semi-confining clay unit that is approximately 1 meter below ground surface. The predominant soil in the wet prairie was found to be Wauberg sand, a loamy, siliceous, hyperthermic Arenic Albaqualfs (Liu et al. 2005).

Micrometeorological, soil moisture within the top 25 cm of soil, and water-table measurements were made during the experimental period. Water-table measurements were made between February 1 and June 30, 2001 and soil moisture measurements were made between April 10 and June 26, 2001. Measurements of actual evapotranspiration were made directly using an energy budget variation of the eddy correlation approach (Jacobs et al. 2002). These measurements were made between May 1 and July 20, 2001.
Micrometeorological measurements of rainfall, net radiation, temperature, relative humidity, and wind speed were made during the entire study period in order to calculate reference potential evapotranspiration.

**Results and Discussion**

Due to the extremely flat topography and the underlying semi-confining unit, vertical and horizontal flow into the model domain by overland or groundwater flow were assumed to be negligible. Using such assumptions the wet prairie can be represented as a single lumped element for modeling purposes where accumulation of water occurs via precipitation and losses via evapotranspiration only. For comparison purposes the wet prairie was also simulated using the Soil-Water-Atmosphere-Plant (SWAP) model, a one-dimensional, finite-difference model that solves Richards’ equation (van Dam and Feddes 2000; Kroes and van Dam 2003). The SWAP model was chosen for comparison because it shares several of the same algorithms with ACRU2000, specifically the determination of plant water stress, the partitioning of potential evapotranspiration between soil evaporation and plant transpiration, and the transition from stage I to stage II soil evaporation.

Following a prior modeling effort of this site the leaf area index was 2.7 and the plant root density distribution was represented as decreasing linearly from the ground surface to a depth of 0.325 meters below ground which satisfies the observation of 95% of the roots within the top 0.25 meters of soil (Jacobs et al. 2002; Liu et al. 2005). Soil moisture characteristic parameter values are shown in Table 4-1 and are adapted from Liu et al. (2005). Reference potential evapotranspiration was calculated using the Penman-Monteith equation as described by Jacobs et al. (2002) and was used along with measured precipitation as the climate forcings in the models. Reference potential
evapotranspiration, as determined by Jacobs et al. (2002) was not adjusted with a crop coefficient. No additional model calibration, beyond that done by Liu et al. (2005) was made here.

Observed and simulated water-table depths are shown in Figure 4-2. The simulated results agree with the measured data with the exception of days during and immediately following a large rain event, where the simulated rise of the water-table is not as intense. This observation is consistent with the response caused by entrapped air and is supported by the rapid decline in the measured water-table soon after the rain event (Fayer and Hillel 1986; Nachabe et al. 2004). Both models deviated from observations when the observed water-table fell below approximately 1 meter in depth. During this period (roughly between 5/25 and 6/15), as the wilting point of the top 25 cm of soil was approached (Figure 4-3), a reduction in the simulated evapotranspiration can be seen as compared to observed values (Figure 4-4). This indicates that the soil, as represented in the model, was not capable of supplying adequate water vertically upwards during this period of dry-down.

Observed and simulated soil moisture content in the top 0.25 m of soil is shown in Figure 4-3. ACRU2000 tended to under-predict the soil moisture content in the top soil during much of the simulation period, however this may be due to the nature of the model. Since the model does not compute water balance components simultaneously, but rather calculates them sequentially, the low soil moisture contents reported by the model at the end of the simulation day are likely due to the fact that evapotranspiration removes water from the root zone after any upward flux (into the root zone) occurs within the soil profile.
Simulated and observed evapotranspiration are shown in Figures 4-4 and 4-5. ACRU2000 often over-predicted on days with high observed ET and SWAP often under-predicted on days with low observed ET.

The goodness of fit between measured and simulated daily values is shown in Table 4-2. Both models performed comparably according to the $MAE$, $RMSE$, and $E$, with SWAP appearing to perform slightly better in predicting water-table depths and actual evapotranspiration and ACRU2000 appearing to perform slightly better in predicting soil moisture contents. Values of the coefficient of efficiency, $E$, compared well to those found by Liu et al. (2005) using daily average potential evapotranspiration inputs. Liu et al. (1995) found values of 0.888, 0.902, and 0.605 for the water-table depth, soil moisture content, and evapotranspiration, respectively.

**Case Study 2: W.F. Rucks Dairy**

**Site Description and Experimental Design**

W.F. Rucks Dairy is a low density improved dairy pasture located in south-central Florida within the Kissimmee River Basin (27°27’N 80°56’W) (Figure 4-6). The pasture is approximately 3.9 ha with an average ground slope of 0.14%. The pasture contained primarily bahia grass ($Paspalum notatum$). The soil of this site was found to be a dominated by Myakka fine sand, a sandy, siliceous, hyperthermic, Aeric Hapludands (Capece 1994). As part of a study to better understand the hydrologic and contaminant transport characteristics of the river basin the site was hydrologically isolated by the construction of a low earthen berm. Surface water flows from the site were measured using a critical-depth trapezoidal flume installed in a breach in the berm. Field ditches within the sites were blocked to mimic undrained, natural conditions. This site has been the focus of prior modeling studies using the FHANTM model (Campbell et al. 1995;
Zhang et al. 1995; Zhang and Gornak 1999) and was thus chosen as a good candidate for testing the ACRU2000 model.

Groundwater level measurements were taken on an approximately weekly basis at 51 well stations throughout the site with each station composed of 2, 3, or 4 wells screened over various depths (Campbell et al. 1995). Rainfall, wind speed, solar radiation, temperature, and relative humidity were measured at the site. The form of the Penman (Penman 1948) equation developed by Jones et al. (1984a) for Florida conditions was used to estimate daily reference potential evapotranspiration (Tremwel 1992). Micrometeorological, groundwater level and runoff measurements were made for 33 months from April 1, 1989 to December 31, 1991.

Results and Discussion

Because of the extremely flat topography and low groundwater gradients measured during the study period (Campbell et al. 1995) model simulation was conducted by treating the pasture as a single, lumped element. Groundwater flow into or out of the model domain were assumed to be negligible. Groundwater level measurements were spatially averaged for comparison to model outputs. Soil physical properties and water characteristics for the site are based on the data of Tremwel (1992) and Sodek et al. (1990) and are shown in Table 4-3. Monthly crop coefficients were taken from the work of Tremwel (1992) and interpolated to daily values by Fourier analysis in the ACRU2000 model (Table 4-4). The plant root distribution was assumed to decrease linearly from the ground surface to a depth of 0.9 m for bahia grass (Fraisse and Campbell 1997).

Model calibration was conducted using the first 17 months of data (April 1, 1989 to August 31, 1990) leaving the final 16 months for model verification (September 1, 1990 to December 31, 1991). Model calibration consisted of adjusting the runoff resistance
coefficient and the depression storage to match observed daily runoff timing and magnitude as well as minor changes to the water characteristic parameters of the soil layers within the range expected for the soil type and location as ascertained from the data of Sodek et al. (1990) in order to match both water-table levels and the timing of runoff generation. Model calibration was made by graphical comparison between observed and simulated daily values. The model was (subjectively) considered to be sufficiently calibrated when successive parameter adjustments appeared to provide little or no improvement in graphically matching observations.

The observed and predicted water-table depths by ACRU2000 and by FHANTM, as reported by Tremwel (1992), are shown in Figure 4-7 for the entire 33 month period. Both models generally followed the observations. Both models deviated from the observations in early 1991. FHANTM deviated from the observations in late 1990 and 1991 as well.

Figure 4-8 shows the simulated and observed daily runoff for the calibration and verification periods for the shallow water-table version of ACRU2000. The timing of the runoff events correspond with periods where the water-table reached the ground surface causing saturation excess overland flow. Simulated and observed daily runoff for the FHANTM and unmodified ACRU2000 models is shown in Figures 4-9 and 4-10. As with the prediction of water-table depths (Figure 4-7) the FHANTM model performed similarly to the modified ACRU2000 model in predicting total runoff (Figure 4-11), however daily runoff events were not predicted as well as by ACRU2000 (Figures 4-8 and 4-9). As seen in Figure 4-10, the prediction of daily runoff by the unmodified ACRU2000 model was poor. This poor prediction is due, almost entirely, to the
reduction of evapotranspiration by the model at soil water contents above field capacity (taken as 100 cm of suction) as detailed in Chapter 2. The original, unmodified ACRU2000 model could predict daily runoff much more accurately if the values of field capacity are set artificially high (near porosity) allowing evapotranspiration to continue at the potential rate under very wet conditions. This result is not shown here. The simulated and observed runoff is also shown in annual cumulative plots in Figure 4-11 for the three models and for the three years of the study.

The statistical measures of the model performance on daily predictions are shown in Table 4-5. Based on the absolute error measures and the coefficients of efficiency the model proposed in Chapter 3 appears to be a better predictor in capturing the water-table dynamics and the generation of saturation excess runoff of the pasture as modeled as a single, lumped element compared to the FHANTM and original, unmodified ACRU2000 models.

Case Study 3: MacArthur Agro-Ecology Research Center at Buck Island Ranch

Site Description and Experimental Design

MacArthur Agro-Ecology Research Center (MAERC) at Buck Island Ranch is a full-scale working cattle ranch owned by The John D. and Catherine T. MacArthur Foundation and leased to Archbold Biological Station (Swain 1998). The site is located in south-central Florida (27° 9’ N and 81° 11’ W), approximately 21 km northwest of Lake Okeechobee (Figure 4-12). As part of a major integrated research project to address the effects of best management practices on nutrient loads in runoff 16 experimental pastures were hydrologically isolated in order to quantify runoff volumes leaving the pastures. For this case study a single 30.2-ha semi-native pasture is simulated as an example of the model’s performance at Buck Island Ranch. The pasture simulated is
identified as Winter Pasture 7. The terrain of the site is extremely flat, with a slope of no greater than 0.02 % (Hendricks 2003) to the north towards the C41 (Harney Pond) canal, a major regional conveyance linking Lake Istokpoga to the north and Lake Okeechobee to the south. Shallow wetlands are interspersed throughout the site covering 4.3% of the land area of the pasture (MAERC 2004), most within 30 m of shallow drainage ditches. The stage of Harney Pond Canal is managed by the South Florida Water Management District (SFWMD) at the S70 spillway located approximately 4 km downstream from the site. The pasture contains semi-native vegetation composed primarily of broomsedge (*Andropogon virginicus*), carpet grass (*Axonopus furcatus*), and bahia grass (*Paspalum notatum*), the wetlands are vegetated primarily with grasses such as carpet grass and maidencane (*Panicum hemitomon*) and with miscellaneous wetland species (MAERC, 2004). Soil surveys of the area were conducted by the USDA-NRCS in June 1997, at a 0.5-ha resolution. Soils in the pasture are predominantly (95.7%) Pineda fine sand, a loamy, siliceous, hyperthermic Arenic Glossaqualfs, with 90% coverage of a thin (2.5-15 cm) muck layer (MAERC 2004).

The pasture is hydrologically isolated with a low earthen berm that forces all runoff from the pasture to exit through a trapezoidal flume located at the downstream end of the pasture. Existing ditches were interconnected to route runoff through the exit flume. Runoff was determined at the trapezoidal flume from water level measurements made in stilling wells at both the upstream and downstream end of the flume in 20 minute intervals (Capece et al. 1999). Meteorological data were collected on an hourly basis at a weather station adjacent to the pasture as well as at 3 nearby stations. Rainfall, temperature, solar radiation, relative humidity, wind speed, and wind direction were
collected at each station. Groundwater levels were measured at 15-minute intervals in a 2-inch monitoring well installed in the center of the pasture. The well extended to a depth of 18 ft below ground surface with the screened portion beginning at 5 ft below ground surface. Runoff and climatic data collection began in May 1998 and groundwater level measurements began in September 2000. Data collection continued until the end of 2003.

Results and Discussion

Due to the extremely low groundwater gradients observed within the pastures (data not shown), and the extremely flat topography, model simulation was conducted by treating the pasture as a single, lumped element. Figure 4-13 shows the canal stage as reported at the S70 spillway located four kilometers downstream as compared to the groundwater level recorded at the center of the experimental pasture. As can be seen, the gradient between the pasture and the canal reverses direction, with groundwater flowing towards the canal during wet periods (typically summer) and canal water flowing towards the pasture during dryer periods. The daily time series of canal stage serves as an input to the model. Groundwater flow between the pasture and the adjacent upland were assumed to be negligible. Soil physical properties and water characteristics for the site are based on the data of Sodek et al. (1990) and Gathumbi et al. (2005) and are shown in Table 4-6 (calibrated values). The plant root density distribution was assumed to decline linearly from the ground surface to a depth of 0.8 meters for the combination of bahia and native grasses (Fraisse and Campbell 1997). Daily reference potential evapotranspiration was calculated using the standardized Penman-Monteith equation recommended by Allen et al. (1998). Crop coefficients used are shown if Table 4-4 and are based on the work of Smajstrla (1990) and Allen et al. (1998).
Model calibration was conducted using the observed runoff and groundwater level data from 1998 to 2001. The length of the calibration period was chosen in order to include adequate groundwater level data (data collection having started in September, 2000). This period allows for 16 months of groundwater level data to be used in calibration. The remaining two years, 2002 and 2003, are used to verify the model calibration. Model calibration was performed by changing the runoff resistance coefficient, increasing the crop coefficients reported by Smajstrla (1990) and Allen et al. (1998) for the winter months, reducing the saturated water content and $n$ soil parameters slightly from the fitted values determined from the data of Sodek et al. (1990), and reducing the hydraulic conductivity by one order of magnitude of the A, E, and Bw soil layers from that reported by Sodek et al. (1990) (adjusted parameters shown in Table 4-6) in order to better replicate the influence of the stage in the adjacent canal on groundwater levels within the pasture. Model calibration was made by graphical comparison between observed and simulated daily values. The model was (subjectively) considered to be sufficiently calibrated when successive parameter adjustments appeared to provide little or no improvement in graphically matching observations.

Observed and simulated groundwater levels are shown in Figure 4-14. The discrepancy between observations and simulated water-table depths, particularly during periods of deeper observed water-tables, may be due to the representation of the canal’s influence on groundwater levels within the pasture, the uncertainty of soil hydraulic parameters due to the lack of field-collected data, and the assumption of no groundwater inflow or outflow from upland areas. No stage measurements of the canal were made at the site, stage measurements at the S70 spillway were assumed to represent the local
canal stage by assuming level pool conditions over the 4 km distance. No groundwater measurements were made in areas upland from the pastures. As can be seen the model performed reasonably well in following the general trend in water-table fluctuations, with both periods of over and under prediction. As can also be seen in Figure 4-14, the model predicts the water-table reaching the ground surface on numerous occasions (at which time runoff may occur by saturation excess) while this does not always appear in the observed data. This is due to the installation of the flume in a shallow ditch which is contiguous with the other shallow ditches in the pasture (land elevation near the groundwater well was approximately 8.47 m and the flume bottom elevation was at 8.08 m). This experimental design causes runoff to occur from the field without the water-table being at the ground surface as reported at the monitoring well since the well is not located within one of these ditches. These shallow ditches are not explicitly represented in the model.

Observed and simulated daily runoff can be seen in Figure 4-15 and reported as cumulative annual plots in Figure 4-16. During the study period there were a few instances of backflow from the canal into the pasture as recorded at the flume (Figure 4-15). These few instances were ignored in generating the cumulative plots in Figure 4-16. As shown the calibration (1998-2001) followed the observations, with the exception of the very dry year of 2000 where runoff events were simulated that were not observed. Verification (2002-2003) matched the observed data less well, but general trends in runoff timing and magnitude were satisfactory.

The goodness of fit between measured and simulated daily values is shown in Table 4-7. The coefficient of efficiency for both the water-table depth and runoff were
lower for this site than for W.F. Rucks Dairy, however the values were greater than zero, indicating that the model is of value as a predictor of water-table depth, saturation excess runoff, and groundwater flow to a time-varying canal stage when applied as a single, lumped element to a site of this size.

**Sensitivity Analysis**

Exact values for model input variables can be costly if not nearly impossible to obtain for even a field-scale model when applied to sites where spatial variation in properties is likely. The accuracy of model input parameter values is usually proportional to the time and resources invested in their determination. Since model results will be more sensitive to certain inputs compared to others it is important to perform a sensitivity analysis in order to establish priorities in collecting and determining model parameters.

An analysis was performed to determine the sensitivity of model simulation of runoff, evapotranspiration, and groundwater flow to the hydrologic input parameters in Table 4-8. The sensitivity analysis was performed using the six-year simulation of the experimental pasture at MAERC. Model sensitivity was determined for ± 25, 50, 75, and 100% of the base input value (shown in Table 4-8). For cases where this range of variation was infeasible, or unrealistic, the results were omitted. Input parameters that had a zero value in the simulation of the experimental pasture (depression storage and the transpiration reduction factor for oxygen deficiency) were changed to the non-zero values shown in Table 4-8 for the sensitivity analysis. Model sensitivity is reported as the percent difference of model results as compared to the base simulation.

The sensitivity of runoff, evapotranspiration, and groundwater flow to the hydrologic parameters in Table 4-8 is shown in Table 4-9. Input parameters showing relatively low sensitivity on all three outputs were root depth, root distribution parameter,
the transpiration reduction factor due to water excess, vertical saturated hydraulic conductivity, upward flux exponent, and interception capacity. Parameters showing little sensitivity with the exception of very large positive or negative changes include depression storage, the transpiration reduction factor due to oxygen deficiency, runoff resistance, bubbling pressure head, and soil moisture shape parameter \( n \). The runoff exponent showed little sensitivity except for at the -25% change. This value was limited to values greater than one (and is a value of 1.67, the value used for the base simulation, for Manning’s equation). The initial depth of the water-table was insensitive on runoff and evapotranspiration, but had a large effect on groundwater flow. This depth was not increased due to the relatively deep initial value of the base simulation. The remaining parameters, the crop coefficient, horizontal hydraulic conductivity, saturated water content, and soil moisture shape parameter \( \alpha \), caused the greatest changes in runoff, evapotranspiration, and groundwater flow. The sensitivity of these parameters on runoff is presented in Figure 4-17.

Considering the uncertainty associated with the parameters that showed the greatest sensitivity on runoff volumes, and the understanding that model calibration rarely results in a single set of “optimal” parameters, a few conclusions can be made. Crop coefficients are usually reported in the literature, if available at all, for single plant species and are often region-specific. Their application to locations other than where they were developed and where there exists a heterogeneous distribution of vegetation species necessitates their use as calibratable parameters within defendable limits. Considering the sensitivity of the soil parameters on runoff volumes, and the use of a set of soil parameters for each soil layer, the model proposed here may be over-parameterized.
Even under circumstances where considerable effort is made to collect soil physical properties there exists a degree of uncertainty in assigning effective model parameter values. For this reason it is postulated that future shallow water-table model development should attempt to reduce the number of parameters required by the model. These simplifications could include the representation of soil moisture retention with a single curve using a single set of parameter values. Horizontal hydraulic conductivity of the soil could be approximated using a single average value, or as decreasing linearly or exponentially with depth as used by Beven and Kirkby (1979). The model could also benefit from the representation of upward flux by a single set of parameters, rather than one for each soil layer representing the combined effects of the layers below it. Such a relationship could be approximated as an exponential decrease with water-table depth and produce similar results to the analytic expressions developed for homogeneous soils (Gardner 1958; Anat et al. 1965; Cisler 1969; Warrick 1988). While approximate, these simplifications would allow for more straightforward parameter adjustment and model calibration.

**Summary and Conclusions**

A field-scale module for use within the ACRU2000 distributed hydrologic model was developed to simulate the hydrology of humid, shallow water-table environments such as the flatwoods of Florida. The field-scale validation of the model for three experimental sites indicated the appropriateness of the physical approximations made by the model, including the assumption of hydrostatic conditions in the unsaturated zone and the use of a daily time-step when simulating regions where saturation-excess is the dominant runoff producing mechanism. This is supported by the model’s ability to satisfactorily replicate field observations of evapotranspiration, soil moisture contents,
water-table depths, and saturation-excess runoff timing and magnitude as well as its ability to produce similar results compared to a numerical, one-dimensional, finite-difference model.

The model was compared to the original, unmodified ACRU2000 model as well as to the field-scale FHANTM model. The shallow-water-table model performed similarly to the FHANTM model. The original, unmodified ACRU2000 model greatly over-predicted runoff due to the simulated reduction of evapotranspiration below the potential rate at soil moisture contents above field capacity as detailed in Chapter 2.

Model sensitivity to parameter values on runoff volumes, evapotranspiration, and groundwater flow was shown to be greatest to crop coefficients and soil hydraulic parameters. Future field experimentation should focus on collecting these parameters to facilitate greater certainty in model simulation. In lieu of this, it is recommended that future model development should explore the simplification of the model by reducing the number of parameters required.
Table 4-1. Wauberg sand soil characteristics and van Genuchten (1980) soil moisture model parameters

<table>
<thead>
<tr>
<th>Layer</th>
<th>Layer depth (cm)</th>
<th>$K_s$ (cm/hr)</th>
<th>$\theta_s$ (cm$^3$/cm$^3$)</th>
<th>$\theta_r$ (cm$^3$/cm$^3$)</th>
<th>$\alpha$ (cm$^{-1}$)</th>
<th>$n$ (-)</th>
<th>$m$ (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td>OA</td>
<td>0-5</td>
<td>246.0</td>
<td>0.58</td>
<td>0.10</td>
<td>0.0305</td>
<td>1.53</td>
<td>0.346</td>
</tr>
<tr>
<td>A</td>
<td>5-15</td>
<td>248.5</td>
<td>0.38</td>
<td>0.03</td>
<td>0.100</td>
<td>1.66</td>
<td>0.398</td>
</tr>
<tr>
<td>E</td>
<td>15-33</td>
<td>85.0</td>
<td>0.45</td>
<td>0.03</td>
<td>0.033</td>
<td>3.1</td>
<td>0.677</td>
</tr>
<tr>
<td>Btg</td>
<td>33-145</td>
<td>0.05</td>
<td>0.30</td>
<td>0.03</td>
<td>0.022</td>
<td>3.0</td>
<td>0.667</td>
</tr>
</tbody>
</table>

Table 4-2. Paynes Prairie State Preserve error measures of daily outputs

<table>
<thead>
<tr>
<th>Parameter</th>
<th>MAE ACRU2000</th>
<th>SWAP</th>
<th>RMSE ACRU2000</th>
<th>SWAP</th>
<th>E(-) ACRU2000</th>
<th>SWAP</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water-table depth</td>
<td>0.083[a]</td>
<td>0.070[a]</td>
<td>0.110[a]</td>
<td>0.098[a]</td>
<td>0.927</td>
<td>0.940</td>
</tr>
<tr>
<td>Soil moisture content</td>
<td>0.022[b]</td>
<td>0.030[b]</td>
<td>0.031[b]</td>
<td>0.044[b]</td>
<td>0.863</td>
<td>0.718</td>
</tr>
<tr>
<td>Evapotranspiration</td>
<td>0.570[c]</td>
<td>0.499[c]</td>
<td>0.729[c]</td>
<td>0.632[c]</td>
<td>0.503</td>
<td>0.626</td>
</tr>
</tbody>
</table>

[a] Units of m.  [b] Units of cm$^3$ cm$^{-3}$.  [c] Units of mm/day.

Table 4-3. Myakka fine sand soil characteristics and van Genuchten (1980) soil moisture model parameters

<table>
<thead>
<tr>
<th>Layer</th>
<th>Layer depth (cm)</th>
<th>$K_s$ (cm/hr)</th>
<th>$\theta_s$ (cm$^3$/cm$^3$)</th>
<th>$\theta_r$ (cm$^3$/cm$^3$)</th>
<th>$\alpha$ (cm$^{-1}$)</th>
<th>$n$ (-)</th>
<th>$m$ (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0-10</td>
<td>23.0</td>
<td>0.38</td>
<td>0.05</td>
<td>0.0199</td>
<td>3.17</td>
<td>0.685</td>
</tr>
<tr>
<td>E</td>
<td>10-33</td>
<td>23.7</td>
<td>0.36</td>
<td>0.04</td>
<td>0.0221</td>
<td>3.42</td>
<td>0.708</td>
</tr>
<tr>
<td>Bh</td>
<td>33-46</td>
<td>12.4</td>
<td>0.32</td>
<td>0.16</td>
<td>0.0218</td>
<td>3.12</td>
<td>0.679</td>
</tr>
<tr>
<td>Bw</td>
<td>46-56</td>
<td>16.0</td>
<td>0.32</td>
<td>0.11</td>
<td>0.0231</td>
<td>3.15</td>
<td>0.683</td>
</tr>
<tr>
<td>Cg</td>
<td>56-203</td>
<td>17.8</td>
<td>0.38</td>
<td>0.07</td>
<td>0.0199</td>
<td>2.59</td>
<td>0.614</td>
</tr>
</tbody>
</table>
Table 4-4. Crop coefficients for W.F. Rucks and MacArthur Agro-Ecology Research Center (MAERC) at Buck Island Ranch

<table>
<thead>
<tr>
<th>Month</th>
<th>W.F. Rucks</th>
<th>MAERC</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>0.40</td>
<td>0.65</td>
</tr>
<tr>
<td>February</td>
<td>0.45</td>
<td>0.75</td>
</tr>
<tr>
<td>March</td>
<td>0.50</td>
<td>0.85</td>
</tr>
<tr>
<td>April</td>
<td>0.50</td>
<td>0.90</td>
</tr>
<tr>
<td>May</td>
<td>0.60</td>
<td>0.95</td>
</tr>
<tr>
<td>June</td>
<td>0.65</td>
<td>1.00</td>
</tr>
<tr>
<td>July</td>
<td>0.85</td>
<td>1.00</td>
</tr>
<tr>
<td>August</td>
<td>0.85</td>
<td>1.00</td>
</tr>
<tr>
<td>September</td>
<td>0.85</td>
<td>1.00</td>
</tr>
<tr>
<td>October</td>
<td>0.75</td>
<td>0.90</td>
</tr>
<tr>
<td>November</td>
<td>0.75</td>
<td>0.80</td>
</tr>
<tr>
<td>December</td>
<td>0.60</td>
<td>0.70</td>
</tr>
</tbody>
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Table 4-5. W.F. Rucks error measures of daily outputs

<table>
<thead>
<tr>
<th></th>
<th>MAE</th>
<th>RMSE</th>
<th>E(-)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ACRU 2000</td>
<td>FHANTM</td>
<td>Original</td>
</tr>
<tr>
<td>Water-table depth</td>
<td>0.095</td>
<td>0.108</td>
<td>0.132</td>
</tr>
<tr>
<td>Runoff</td>
<td>3.32</td>
<td>8.55</td>
<td>8.33</td>
</tr>
</tbody>
</table>

[a] Units of m.  [b] Units of mm.

Table 4-6. Pineda fine sand soil characteristics and van Genuchten (1980) soil moisture model parameters

<table>
<thead>
<tr>
<th>Layer</th>
<th>Depth (cm)</th>
<th>Ks (cm/hr)</th>
<th>θs (cm³/cm³)</th>
<th>θr (cm³/cm³)</th>
<th>α (cm⁻¹)</th>
<th>n (-)</th>
<th>m (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0-10</td>
<td>4.32</td>
<td>0.42</td>
<td>0.10</td>
<td>0.0287</td>
<td>1.96</td>
<td>0.490</td>
</tr>
<tr>
<td>E</td>
<td>10-30</td>
<td>4.14</td>
<td>0.34</td>
<td>0.08</td>
<td>0.0224</td>
<td>2.57</td>
<td>0.611</td>
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<td>Bw</td>
<td>30-80</td>
<td>2.70</td>
<td>0.32</td>
<td>0.07</td>
<td>0.0234</td>
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<td>0.448</td>
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<td>Btg</td>
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<td>1.37</td>
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<td>Cg</td>
<td>140-160</td>
<td>2.95</td>
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<td>0.06</td>
<td>0.0106</td>
<td>2.06</td>
<td>0.515</td>
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Table 4-7. MacArthur Agro-Ecology Research Center at Buck Island Ranch error measures from the experimental pasture for daily outputs

<table>
<thead>
<tr>
<th>Parameter</th>
<th>MAE</th>
<th>RMSE</th>
<th>E (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water-table depth</td>
<td>0.253(^a)</td>
<td>0.310(^a)</td>
<td>0.631</td>
</tr>
<tr>
<td>Runoff</td>
<td>0.472(^b)</td>
<td>1.53(^b)</td>
<td>0.572</td>
</tr>
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</table>

\(^{a}\) Units of m.  \(^{b}\) Units of mm.
Table 4-8. Hydrologic input parameters included in the sensitivity analysis

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Unit</th>
<th>Base Value</th>
<th>Description</th>
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<td>Kc</td>
<td>–</td>
<td>0.875</td>
<td>Crop coefficient</td>
</tr>
<tr>
<td>L</td>
<td>m</td>
<td>0.8</td>
<td>Maximum depth of roots</td>
</tr>
<tr>
<td>c</td>
<td>–</td>
<td>-1</td>
<td>Root distribution shape parameter</td>
</tr>
<tr>
<td>h₂</td>
<td>cm</td>
<td>20</td>
<td>Transpiration reduction due to O₂ deficiency</td>
</tr>
<tr>
<td>h₃</td>
<td>cm</td>
<td>10000</td>
<td>Transpiration reduction due to water excess</td>
</tr>
<tr>
<td>ZDep</td>
<td>mm</td>
<td>10.0</td>
<td>Depression storage</td>
</tr>
<tr>
<td>I</td>
<td>mm</td>
<td>1</td>
<td>Interception capacity</td>
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<tr>
<td>γ</td>
<td>1/day</td>
<td>200</td>
<td>Runoff resistance</td>
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<tr>
<td>β</td>
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<td>1.67</td>
<td>Runoff exponent</td>
</tr>
<tr>
<td>dₜ</td>
<td>m</td>
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<td>Initial depth to water-table</td>
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<td>θₛ</td>
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<tr>
<td>θₑ</td>
<td>cm³/cm³</td>
<td>0.32</td>
<td></td>
</tr>
<tr>
<td>θₑ</td>
<td>cm³/cm³</td>
<td>0.35</td>
<td></td>
</tr>
<tr>
<td>θₑ</td>
<td>cm³/cm³</td>
<td>0.30</td>
<td></td>
</tr>
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<td>α</td>
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<td>Soil moisture shape parameter of van Genuchten (1980)</td>
</tr>
<tr>
<td>α</td>
<td>1/cm</td>
<td>0.0224</td>
<td></td>
</tr>
<tr>
<td>α</td>
<td>1/cm</td>
<td>0.0234</td>
<td></td>
</tr>
<tr>
<td>α</td>
<td>1/cm</td>
<td>0.0177</td>
<td></td>
</tr>
<tr>
<td>α</td>
<td>1/cm</td>
<td>0.0106</td>
<td></td>
</tr>
<tr>
<td>n</td>
<td>–</td>
<td>1.96</td>
<td>Soil moisture shape parameter of van Genuchten (1980)</td>
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<td>–</td>
<td>2.57</td>
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<td>n</td>
<td>–</td>
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</tr>
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<td>n</td>
<td>–</td>
<td>1.67</td>
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</tr>
<tr>
<td>n</td>
<td>–</td>
<td>2.06</td>
<td></td>
</tr>
<tr>
<td>Kₛ,ₜ</td>
<td>cm/h</td>
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<td>Horizontal saturated hydraulic conductivity</td>
</tr>
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<td>Kₛ,ₜ</td>
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<td></td>
</tr>
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<tr>
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<td>Vertical saturated hydraulic conductivity used in the upward flux relationship of Anat et al. (1965)</td>
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Table 4-9. Sensitivity of runoff, evapotranspiration, and groundwater flow to hydrologic parameters (reported as percent difference of base simulation result). Values in left-hand column are percent changes in input parameters.

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Figure 4-1. Location of Paynes Prairie State Preserve.

Figure 4-2. Measured and predicted water-table depths at Paynes Prairie State Preserve.
Figure 4-3. Measured and predicted soil moisture contents within the top 25 cm of soil at Paynes Prairie State Preserve.

Figure 4-4. Measured and predicted evapotranspiration at Paynes Prairie State Preserve.
Figure 4-5. Measured vs. predicted evapotranspiration at Paynes Prairie State Preserve. $r^2 = 0.83$ and 0.74 for SWAP and ACRU2000, respectively.

Figure 4-6. Location of W.F. Rucks Dairy.
Figure 4-7. Measured and predicted water-table depths at W.F. Rucks Dairy.

Figure 4-8. Measured and modified ACRU2000 predicted daily runoff at W.F. Rucks Dairy. For the calibration (top) and verification (bottom) periods.
Figure 4-9. Measured and FHANTM predicted daily runoff at W.F. Rucks Dairy. For the calibration (top) and verification (bottom) periods.

Figure 4-10. Measured and unmodified ACRU2000 predicted daily runoff at W.F. Rucks Dairy. For the calibration (top) and verification (bottom) periods.
Figure 4-11. Measured and predicted cumulative annual runoff at W.F. Rucks Dairy.
Note: Observed runoff events during 1990 fell within the calibration period.

Figure 4-12. Location of MacArthur Agro-Ecology Research Center at Buck Island Ranch.
Figure 4-13. Groundwater level and adjacent canal stage in the experimental pasture at the MacArthur Agro-Ecology Research Center at Buck Island Ranch. Canal stage is the stage in the C41 canal as measured at the S-70 spillway located 4 km downstream and the groundwater level is from the 2-inch well (center of pasture).

Figure 4-14. Measured and predicted water-table depths at the experimental pasture at the MacArthur Agro-Ecology Research Center at Buck Island Ranch.

Figure 4-17. The parameters most sensitive on runoff volumes.
CHAPTER 5
FIELD-SCALE NITROGEN AND PHOSPHORUS MODULE OF THE ACRU2000 MODEL

Introduction

The nitrogen (N) and phosphorus (P) components of the Groundwater Loading Effects of Agricultural Management Systems (GLEAMS) model were incorporated into the ACRU2000 model in a previous model expansion (Campbell et al. 2001). GLEAMS, a commonly used field-scale hydrology and water quality model, represents the major N and P components and transformations (Knisel et al. 1993). The P algorithms used in GLEAMS are largely incorporated from the EPIC (Erosion-Productivity Impact Calculator) model (Jones et al. 1984b). This chapter details the N and P algorithms of the ACRU2000 model, as added by Campbell et al. (2001), and discusses their suitability for shallow water-table environments.

Nutrient Models

The algorithms of the model describe the mobilization and transport of dissolved forms of N and P. Sediment-bound nutrients are not currently represented explicitly in the model since sediment yield is not, to date, separated by particle-size class. In the case of N the soluble component is split between nitrate-N and ammonium-N, with mineralization of organic forms represented as a two-stage process. Soluble organic N is not simulated in the model. P is represented as a single soluble form, labile P. Both ammonium-N and labile P undergo fast, reversible sorption to soil particles. Transport of soluble nutrients occurs between completely mixed soil layers. The exchange of
nutrients with runoff water is assumed to occur within the top one centimeter of soil.

Most nutrient transformations are mediated by soil moisture and soil temperature.

**Phosphorus Model**

The P model is represented by six main pools (Figure 5-1), a labile pool ($P_l$), an “active” inorganic pool ($P_a$), a slowly changing “stable” inorganic pool ($P_s$), a fresh organic pool ($P_f$) representing plant residue, a stable organic (organic humus) pool ($P_h$), and a plant pool ($P_p$). Of the six pools, only $P_l$ is mobile. The mass balance of the pools can be written as:

\[
\frac{dP_l}{dt} = P_{\text{rain}} + P_{\text{fert}} + 0.75R_{awP} + R_{alP} + R_{flP} + R_{hlP} - R_{uP} - P_{ro} - P_{gw} - P_{perc} \quad (5-1)
\]

\[
\frac{dP_a}{dt} = R_{sap} - R_{alP} \quad (5-2)
\]

\[
\frac{dP_f}{dt} = -R_{sap} \quad (5-3)
\]

\[
\frac{dP_f}{dt} = R_{pfP} - R_{flP} - R_{fhP} \quad (5-4)
\]

\[
\frac{dP_p}{dt} = 0.25R_{awP} + R_{fhP} - R_{hlP} \quad (5-5)
\]

\[
\frac{dP_p}{dt} = R_{up} - R_{pfP} \quad (5-6)
\]

where $P_{\text{rain}}$ is the quantity of P in rainfall (kg/ha/day), $P_{\text{fert}}$ is the rate of application of P in inorganic fertilizer (kg/ha/day), $R_{awP}$ is the rate of decay of P in animal waste on the ground surface (kg/ha/day), $R_{alP}$ is the rate of transformation from $P_a$ to $P_l$ (kg/ha/day), $R_{flP}$ is the rate of transformation from $P_f$ to $P_l$ (kg/ha/day), $R_{hlP}$ is the rate of transformation from $P_h$ to $P_l$ (kg/ha/day), $R_{uP}$ is the rate of plant uptake (kg/ha/day), $R_{sap}$ is the rate of transformation from $P_s$ to $P_a$ (kg/ha/day), $R_{pfP}$ is the rate of transformation of $P_p$ to $P_f$ (kg/ha/day), $R_{fhP}$ is the rate of transformation from $P_f$ to $P_h$ (kg/ha/day), and $P_{ro}$,
$P_{gw}$, and $P_{perc}$ are the quantities of P lost to runoff, groundwater flow, and percolation (kg/ha/day), respectively.

**Mineralization**

Mineralization of organic forms of P is represented as first-order processes. The fast-cycling fresh organic pool ($P_f$) consists of surface crop residues resulting from harvest and sub-surface root residues (C:P ratios generally greater than 200). The slow-cycling organic humus pool ($P_h$) consists of more recalcitrant organic forms (C:P ratios between 125 and 200). The mineralization of animal waste ($R_{awP}$), fresh organic P ($R_{fip}$), and organic humus P ($R_{hlp}$), are defined as, respectively (Knisel et al. 1993):

$$R_{awP} = C_{NP} k_d P_{aw} \sqrt{f_{\theta \text{min}} f_{T \text{min}}}$$  (5-7)

$$R_{fip} = C_{NP} k_d P_f \sqrt{f_{\theta \text{min}} f_{T \text{min}}}$$  (5-8)

$$R_{hlp} = f_{\text{act}} k_{hf} P_h \sqrt{f_{\theta \text{min}} f_{T \text{min}}}$$  (5-9)

where $P_{aw}$ is the P content in animal waste (kg/ha), $f_{\text{act}}$ is the fraction of total N that is active N which is used to infer the fraction of $P_h$ that is mineralizable, $k_{hf}$ is the rate of organic humus P decomposition under optimum conditions (assumed to be 0.0001 day$^{-1}$), $C_{NP}$ is a factor that varies from 0 to 1 that is a function of C:N and C:P ratios of the organic material (Jones et al. 1984b):

$$C_{NP} = \min \left\{ \exp \left[ -0.693 \frac{C : N - 25}{25} \right], \frac{1}{1.0} \exp \left[ -0.693 \frac{C : P - 200}{200} \right] \right\}$$  (5-10)

where the C:N and C:P ratios are determined as the ratios of fresh residue and animal waste mass to the mass of organic and inorganic N and P present in fresh residue and animal waste, $k_d$ is an organic matter composition factor and represents the age of the
decomposing material by assuming the first 20% is carbohydrate-like material, the final
10% is lignin, and the intermediate is cellulose (Jones et al. 1984b):

\[ k_d = 0.8 \quad \text{for} \quad f_{\text{dec}} > 0.8 \]  
\[ k_d = 0.05 \quad \text{for} \quad 0.8 \geq f_{\text{dec}} > 0.1 \]  
\[ k_d = 0.0095 \quad \text{for} \quad f_{\text{dec}} \leq 0.1 \]  

where \( f_{\text{dec}} \) is the fraction of the initial material remaining, \( f_{\theta\text{min}} \) and \( f_{T\text{min}} \) are soil moisture and soil temperature response factors that vary from 0 to 1 and are described below.

Mineralization of animal waste P, \( P_{aw} \), and fresh organic P, \( P_f \), are assumed to be partitioned to \( P_l (75\%) \) and \( P_h (25\%) \) (Knisel et al. 1993).

**Immobilization**

The high C:P ratios of fresh crop residue (generally greater than 200) results in the immobilization of labile P by soil microbes during the decomposition process. The rate of uptake of labile P by decomposing material is dependent upon the stage of decomposition, the C:N and C:P ratios of the residue, the concentrations of P in fresh organic matter, and labile P (Knisel et al. 1993):

\[ R_{IP} = C_{NP} k_d P_f \sqrt{f_{\theta\text{min}} f_{T\text{min}}} (0.16 f_{IP} - C_{IP}) \]  

where the value 0.16 results from assuming that carbon is 40% of fresh organic matter and that 40% of the carbon can be assimilated by soil microbes (Jones et al. 1984b), \( C_{IP} \) is the concentration of fresh organic P (kg/kg) and \( f_{IP} \) is a phosphorus immobilization factor:

\[ f_{IP} = 0.01 + 0.001 C_{IP} \quad \text{for} \quad C_{IP} \leq 10.0 \]  
\[ f_{IP} = 0.02 \quad \text{for} \quad C_{IP} > 10.0 \]
and $C_{IP}$ is the concentrations of labile P (mg/kg). The rate of immobilization can be limited by either N or P if the amount of N or P immobilized is less than that available (Knisel et al. 1993).

**Inorganic transformations**

As P is added to soil solution the equilibrium between mobile and immobile forms is disturbed. Following P additions, net P movement occurs into immobile forms. The initial rate of these adsorption/fixation reactions is rapid, leaving newly formed immobile forms relatively unstable and readily returned to solution (McGechan and Lewis 2002). The movement between the labile P and active P pool is considered to be a rapid equilibrium (several days to weeks) (Jones et al. 1984b). P moves between the pools as a function of the relative size of the pools, moisture content, and temperature (Jones et al. 1984b):

$$R_{alP} = k_{al}f_{θmin}f_{Tal}[P_1 - P_ασ_{al}]$$

(5-14)

where $k_{al}$ is the rate constant at optimal conditions (assumed to be 0.1 day$^{-1}$), and $f_{θmin}$ and $f_{Tal}$ are soil moisture and soil temperature response factors (defined below), the equilibrium constant of proportionality between the two pools, $σ_{al}$ is estimated from a P sorption parameter, $PSP$ (Jones et al. 1984b):

$$σ_{al} = \frac{PSP}{1 - PSP}$$

(5-15)

The P sorption parameter, also referred to as the P availability index, is defined as the fraction of P added to a soil sample that remains labile after a long incubation period:

$$PSP = \frac{P_{FinalLabile} - P_{InitialLabile}}{P_{added}}$$

(5-16)

Following Sharpley et al. (1984), Sharpley and Williams (1990) defined $PSP$ by dividing
soils into three groups based on soil taxonomy and weathering. The three groups defined were calcareous, slightly weathered, and highly weathered soils. For slightly weathered soils such as Spodosols and Alfisols (except Ultic subgroups) PSP was defined as a function of base saturation, $B_{SAT} (%)$ and $pH$:

$$PSP = 0.0054B_{SAT} + 0.116pH - 0.73$$

(5-17)

and is constrained between values of 0.05 and 0.75. This results in values between 0.05 and 3 for $\sigma_{al}$, defining $P_l$ as 0.05 to 3 times as large as $P_a$ at equilibrium.

The active P pool is considered to be in a slow equilibrium with the stable P pool. The differentiation between “active” and “stable” inorganic P pools is made in order to account for the initial rapid decrease in labile P typically seen after P application followed by a much slower decrease in observed labile P over long periods (Jones et al. 1984b). This representation is a simplification of a continuum of time-dependent adsorption or fixation reactions (McGechan and Lewis 2002). The transfer of P between active P and stable P pools is a function of the relative size of the pools (Jones et al. 1984b):

$$R_{sap} = k_{sa} [\sigma_{sa} P_a - P_s]$$

(5-18)

where $\sigma_{sa}$ is an equilibrium constant of proportionality and is assumed to be a value of 4 (Jones et al. 1984b), and $k_{sa}$ is a rate constant that is defined in the GLEAMS model as a function of $PSP$ for non-calcareous soils (Jones et al. 1984b):

$$k_{sa} = \exp(-1.77PSP - 7.05)$$

(5-19)

In addition to the rapid adsorption/fixation reactions between the labile and active P pools, the labile P pool undergoes instantaneous, reversible sorption governed by a linear
isotherm in order to determine the portion of labile P that is in solution and available for transport into runoff, groundwater, and percolating water:

\[ S = K_d C \]  

(5-20)

where \( S \) is the portion of labile P adsorbed (mg/kg), \( C \) is the portion in solution (mg/L) and \( K_d \) is the partitioning coefficient (L/kg). The partitioning coefficient is assumed to be a function of the clay content of the soil, \( CL (\%) \) (Knisel et al. 1993):

\[ K_d = 100 + 2.5CL \]  

(5-21)

where \( K_d \) is in units of L/kg.

**Plant uptake**

Plant uptake of P is assumed to occur from each soil layer from which transpiration occurs, with the total uptake limited to a calculated plant demand. Plant demand is determined from the plant growth rate and plant nutrient characteristics (Knisel et al. 1993).

**Nitrogen Model**

The nitrogen model is represented by six main pools, nitrate-N \( (N_{NO3}) \), ammonium-N \( (N_{NH4}) \), active organic N \( (N_a) \), stable organic N \( (N_s) \), fresh organic N \( (N_f) \), and plant N \( (N_p) \) (Figure 5-2). The mass balance of the pools can be written as:

\[
\frac{dN_{NO3}}{dt} = N_{\text{rain}} + N_{\text{fertNO3}} + R_{\text{nit}} - R_{\text{denit}} - R_{\text{immNO3}} - R_{\text{wNO3}} - N_{\text{roNO3}} - N_{\text{gwNO3}} - N_{\text{percNO3}} \\
\frac{dN_{NH4}}{dt} = N_{\text{fertNH4}} + 0.8R_{\text{awN}} + R_{\text{ammonN}} + 0.8R_{\text{amminfN}} - R_{\text{nit}} - R_{\text{wN}} - R_{\text{immNH4}} - R_{\text{wNH4}} - N_{\text{roNH4}} - N_{\text{gwNH4}} - N_{\text{percNH4}} \\
\frac{dN_a}{dt} = 0.2R_{\text{awN}} + 0.2R_{\text{amminfN}} + R_{\text{saN}} - R_{\text{ammonN}} \\
\frac{dN_s}{dt} = -R_{\text{saN}}
\]  

(5-22)

(5-23)

(5-24)

(5-25)
\[
\frac{dN_f}{dt} = R_{pfN} + R_{ammNO3} + R_{ammNH4} - R_{ammfN} \\
\frac{dN_p}{dt} = R_{uNO3} + R_{uNH4} - R_{pfN}
\]

(5-26)

(5-27)

where \(N_{rain}\) is the quantity of N in rainfall (kg/ha/day), and is assumed to occur completely as nitrate-N, \(N_{fertNO3}\) and \(N_{fertNH4}\) are nitrate-N and ammonium-N quantities in inorganic fertilizer (kg/ha/day), \(R_{awN}\) is the rate of animal waste N decay (kg/ha/day), \(R_{nit}\) is the rate of nitrification of ammonium-N to nitrate-N (kg/ha/day), \(R_{denit}\) is the rate of loss of nitrate-N to the atmosphere by denitrification (kg/ha/day), \(R_{amnaN}\) and \(R_{ammfN}\) are the rates of ammonification of the active and fresh organic N pools (kg/ha/day), respectively, \(R_{vol}\) is the rate of loss of ammonium-N in manure to the atmosphere by ammonia volatilization (kg/ha/day), \(R_{sawN}\) is the rate of transformation of stable N to active N (kg/ha/day), \(R_{ammNO3}\) and \(R_{ammNH4}\) are, respectively, the rates of immobilization of nitrate-N and ammonium-N by fresh organic matter (kg/ha/day), \(R_{uNO3}\) and \(R_{uNH4}\) are the rates of plant uptake of nitrate-N and ammonium-N (kg/ha/day), \(R_{pfN}\) is the rate of transformation of plant N to fresh organic N (kg/ha/day), \(N_{roNO3}\) and \(N_{roNH4}\) are the quantities of nitrate-N and ammonium-N lost in runoff (kg/ha/day), \(N_{gwNO3}\) and \(N_{gwNH4}\) are the amount of nitrate-N and ammonium-N lost in groundwater flow (kg/ha/day), and \(N_{percNO3}\) and \(N_{percNH4}\) are the quantities if nitrate-N and ammonium-N in percolating water (kg/ha/day).

**Mineralization**

Mineralization of organic N forms is represented as a first-order process in a similar manner as for P. Mineralization of fresh organic N and animal waste N is partitioned to ammonium-N (80%) and active organic-N (20%). Mineralization of active organic N is assumed to be completely converted to ammonium-N. The rates of
The immobilization of nitrate-N and ammonium-N is represented in a similar manner as that for labile P:

\[
R_{imm} = C_{NP} k_d N_f \sqrt{f_{\theta \min} f_{T \min}} \left(0.016 - C_{fN}\right)
\]  

where \(C_{fN}\) is the concentration of N in fresh residue (kg/kg) and the constant 0.016 comes from assuming that carbon is 40% of fresh organic matter and that 40% of the carbon can
be assimilated by soil microbes and that the microbial biomass and its products have a C:N ratio of 10 (Knisel et al. 1993). Nitrate-N and ammonium-N are immobilized in proportion to their availability.

**Atmospheric loss of nitrogen**

Both ammonium-N and nitrate-N may be lost to the atmosphere via the volatilization of ammonia gas and the denitrification of nitrate. Ammonia volatilization is assumed to occur from the portion of $N_{NH4}$ contained in animal waste and only for a period of one week following application. The rate of volatilization is given by (Knisel et al. 1993):

\[
R_{vol} = N_{NH4} \exp(-k_v t)
\]  

where \( t \) is time in days and \( k_v \) is volatilization rate constant \( \text{(day}^{-1}) \):

\[
k_v = 0.409(1.08)^{t-20}
\]  

The rate of denitrification is defined as (Knisel et al. 1993):

\[
R_{denit} = N_{NO3} \left[ 1 - \exp\left( -k_{denit} f_{\theta \min} f_{\theta \denit} \right) \right]
\]  

where \( f_{\theta \denit} \) is the soil moisture response factor to denitrification (defined below), \( k_{denit} \) is the rate coefficient for denitrification which is a function of active soil carbon, \( SC \) (mg/kg) (Knisel et al. 1993):

\[
k_{denit} = 0.106SC + 0.202
\]  

The active soil carbon is estimated from the active N pool (Knisel et al. 1993):

\[
SC = \frac{0.018N_a}{M_{soil}}
\]
Inorganic nitrogen

Nitrate-N is assumed to be completely conservative in solution. Ammonium-N is assumed to undergo instantaneous, reversible sorption following a linear isotherm (Equation 5-15) where the partitioning coefficient is assumed to be a function of the clay content of the soil (Knisel et al. 1993):

\[ K_d = 1.34 + 0.083CL \]  (5-39)

Plant uptake

Plant uptake of N occurs in the same manner as for labile P, from each soil layer that transpiration occurs. The total uptake is limited to a calculated plant demand in the same manner as for P. Uptake of nitrate-N and ammonium-N is assumed to occur in proportion to their availability (Knisel et al. 1993).

Nutrient Transformation Response to Soil Moisture

In the model, there are three soil moisture functions employed for various processes. For ammonification, P mineralization, and mineral N and P immobilization the soil moisture response function is of the form (shown in Figure 5-3) (Knisel et al. 1993):

\[ f_{\theta_{min}} = \frac{\theta - \theta_{wp}}{\theta_{fc} - \theta_{wp}} \quad \text{for} \quad \theta \leq \theta_{fc} \]  (5-40a)
\[ f_{\theta_{min}} = 0 \quad \text{for} \quad \theta > \theta_{fc} \]  (5-40b)

where \( \theta \), \( \theta_{wp} \), and \( \theta_{fc} \) are the moisture content of a soil layer (cm\(^3\) cm\(^{-3}\)), the moisture content at the wilting point (cm\(^3\) cm\(^{-3}\)), and the moisture content at field capacity (cm\(^3\) cm\(^{-3}\)), respectively. As can be seen in figure 5-3, the response to soil moisture rises from zero at the wilting point to an optimum value at field capacity. Immediately above field
capacity the response reduces to zero, implying a complete cessation of microbial activity under wet conditions.

For nitrification the soil moisture response function is at an optimum value at field capacity and decreases linearly to zero at saturation and at the wilting point (Figure 5-3) (Knisel et al. 1993):

\[
f_{\text{init}} = \frac{\theta - \theta_{wp}}{\theta_{fc} - \theta_{wp}} \quad \text{for } \theta \leq \theta_{fc} \tag{5-41a}
\]
\[
f_{\text{init}} = 1 - \frac{\theta - \theta_{fc}}{\theta_s - \theta_{fc}} \quad \text{for } \theta_{fc} < \theta < \theta_s \tag{5-41b}
\]

where \( \theta_s \) is the water content at saturation (cm\(^3\) cm\(^{-3}\)).

For denitrification the soil moisture response function begins when the water content is 10% above field capacity and increases linearly to saturation (Figure 5-3) (Knisel et al. 1993):

\[
f_{\text{denit}} = \frac{\theta - [\theta_{fc} + 0.1(\theta_s - \theta_{fc})]}{\theta_s - \left[\theta_{fc} + 0.1(\theta_s - \theta_{fc})\right]} \quad \text{for } \theta \geq \theta_{fc} + 0.1(\theta_s - \theta_{fc}) \tag{5-42a}
\]
\[
f_{\text{denit}} = 0 \quad \text{for } \theta < \theta_{fc} + 0.1(\theta_s - \theta_{fc}) \tag{5-42b}
\]

**Nutrient Transformation Response to Temperature**

The model uses two temperature response factors (Figure 5-4) to adjust transformation rates. For P mineralization, P immobilization, N ammonification, N immobilization, and denitrification the relationship is (Knisel et al. 1993):

\[
f_{T_{\text{min}}} = \frac{T}{T + \exp(9.93 - 0.312T)} \tag{5-43}
\]

where \( T \) is the temperature (°C). For the transformation between \( P_l \) and \( P_o \) the soil temperature factor is (Jones et al. 1984b):

\[
f_{T_{al}} = \exp(0.115T - 2.88) \tag{5-44}
\]
Extraction of Nutrients into Runoff

The approach to the extraction of nutrients into runoff by the model assumes a constant rate of rainfall ($R$) falling on an area of saturated soil with negligible surface ponding. If the mixing of rainfall with the soil water can be considered complete and instantaneous, the mass balance of an adsorbing nutrient can be written as (Steenhuis and Walter 1980):

$$\frac{d(\theta_s Cd + K_d \rho_b Cd)}{dt} = -Rt$$  \hspace{1cm} (5-45)

where $d$ is the depth of soil considered to be completely mixed with runoff water (sometimes referred to as the effective depth of interaction, EDI), $C$ is the concentration of the nutrient in solution, $\rho_b$ is the bulk density of the soil, and $t$ is time. Integration of Equation 5-45 results in:

$$C = C_o \exp \left( \frac{-Rt}{\theta_s d + K_d \rho_b d} \right)$$  \hspace{1cm} (5-46)

where $C_o$ is the initial concentration in solution. In the model the effective depth of interaction is assumed constant at 1 cm, however mixing is assumed to be incomplete. This is accomplished by defining the mass of nutrient within the mass of soil, $Y$, as the product of the concentration in soil water as defined in Equation 5-46 and the soil mass per unit volume of overland flow, $\beta$ (Leonard et al. 1987):

$$Y = C \beta$$  \hspace{1cm} (5-47)

as the nutrient equilibrates between soil and runoff (Leonard et al. 1987):

$$Y = C_r V + S \beta$$  \hspace{1cm} (5-48)

where $C_r$ is the concentration in runoff, $V$ is the volume of water unit volume of runoff interface, and $S$ is the amount of nutrient in the sorbed phase. By disregarding the soil
mass compared to the volume of water (making the $V = 1$), substituting Equation 5-20 for $S$ ($S = K_dC_{ro}$), and equating Equations 5-47 and 5-48 (Leonard et al. 1987):

$$C_{ro} = \frac{C\beta}{1 + K_d\beta}$$  \hspace{1cm} (5-49)

where in the case of a conservative substance, $C_{ro} = C\beta$, and for very large values of $K_d$, $C_{ro}$ approaches 0. As the parameter $\beta$ is unmeasureable in practice, it is referred to as the extraction coefficient (Knisel et al. 1993). The extraction coefficient is assumed to vary between 0.1 and 0.5 as a function of the partitioning coefficient. For ammonium N and labile P, $\beta$ is defined as (Knisel et al. 1993):

$$\begin{align*}
\beta &= 0.5 & \text{for} & & K_d \leq 1.0 & \hspace{1cm} (5-50a) \\
\beta &= 0.598\exp(-0.179K_d) & \text{for} & & 1.0 < K_d < 10.0 & \hspace{1cm} (5-50b) \\
\beta &= 0.1 & \text{for} & & K_d \geq 10.0 & \hspace{1cm} (5-50c)
\end{align*}$$

Application of the Nitrogen and Phosphorus Module to Shallow Water-Table Environments

From the above description of the N and P module of the ACRU2000 model, several potential shortcomings of the model can be noted. These shortcomings apply to simulating shallow water-table environments in particular, and varied field, soil, and management conditions in general. The GLEAMS model, from which the N and P models have been adapted, has been evaluated under different field and management conditions with varied results (e.g. Stone et al. 1998; Bakhsh et al. 2000; Dukes and Ritter 2000; Chinkuyu and Kanwar 2001). General shortcomings of the model include:

- the use of constant, hard-coded maximum reaction rates and proportionality coefficients for various transformations and nutrient pools,
- the assumption that linear partitioning coefficients can be sufficiently described by soil clay content only, and
• the assumption that no plant litter (residue) is produced until harvest operations are simulated.

It is evident from these shortcomings that the GLEAMS model was developed using a “one size fits all” approach. It should be noted however that field-specific data of maximum rates of reaction, equilibrium states between nutrient pools, and sorption characteristics are rarely, if ever, available and the approach used by GLEAMS is an attempt to simplify the use of the model. Additionally, the development of the GLEAMS model envisioned its use primarily for cropping systems, for which the approximation that crop residue is only produced at harvest is likely acceptable.

From the description of the model some specific potential shortcomings concerning its application to shallow-water table environments can also be noted:

• the response of nutrient transformations to soil moisture conditions, and

• the functional approach used by the model to represent the extraction of nutrients into runoff.

The assumption that the nutrient transformation rates of mineralization and immobilization cease completely at soil moisture contents above field capacity is known to be untrue (Brady and Weil 1996). The use of such a relationship by a model like GLEAMS is indicative of the field conditions for which it was developed, upland sites. For an upland site, where extreme soil wetness is a transient occurrence, the assumption that microbially mediated reactions cease under wet conditions may be an acceptable approximation of the expected depressed microbial activity. Additionally, it has been made quite clear in the literature that many transformations occur at maximal rates over a range of soil water contents and not at a singular value of field capacity (Rodrigo et al. 1997).
The form of the equation used to determine the extraction of nutrients into runoff (5-46) has been shown to be inappropriate under saturated soil conditions (Ahuja 1982). Under such conditions the extraction of solutes into runoff does not follow an exponential curve and the assumption that only the top one centimeter of soil interacts with runoff has been shown to be inaccurate (Ahuja 1982). Additionally, neglecting the presence of ponded water and nutrient concentrations in ponded water (and concentration gradients between ponded water and the soil water) may be problematic when applied to a shallow water-table environment where runoff occurs by saturation-excess and where low ground slopes retain ponded water for large times. Over larger time periods concentration gradients between ponded/runoff water and soil water may affect the relative movement of solutes. The concentration gradients will also be of importance should the model be applied in a distributed framework where runoff from one land segment becomes “run-on” to another. In such a situation it might be expected that less transfer of nutrients would take place into the surface water due to a low concentration gradient between the soil and “run-on” water. Additionally, the meaning of the extraction coefficient and its relationship to partitioning coefficients does not appear to be well supported for N and P in the literature (Knisel et al. 1993).

Summary

This chapter describes the field-scale N and P transformations and transport as simulated by the ACRU2000 model (Campbell et al. 2001) using algorithms from the GLEAMS model (Knisel et al. 1993). The model represents N and P using six main pools. Phosphorus consists of a labile, an immobile active inorganic, an immobile stable inorganic, and immobile fresh, stable humus, and plant organic pools. Nitrogen consists of two mobile forms, ammonium and nitrate, and organic pools representing fresh, active,
stable and plant pools. The mobile forms, labile P and ammonium N, undergo
instantaneous, reversible sorption to soil particles. Most of the nutrient transformations
are affected by soil moisture and temperature.

The model may not be appropriate for shallow water-table environments due to the
representation of nutrient transformations in response to soil moisture conditions and due
to the relationship defining the extraction of nutrients into runoff. The model assumes
many transformation rates to cease at water contents above field capacity and uses a
relationship that has been shown to be inappropriate in estimating the extraction of
solute into runoff/ponded water under conditions of saturation-excess. Additionally, the
model assumes that plant residue is only produced in response to harvest operations and
that the instantaneous, reversible sorption of P can be defined from the clay content of the
soil only.

Modifications to ACRU2000 that may be more appropriate for shallow-water table
environments and sandy flatwoods soils are described in the next chapter. Only some of
the potential issues of the model detailed here are addressed in the following chapter.
The remaining are addressed by current research not included in this dissertation. The
modified model proposed in the next chapter is compared to the model as described here
in Chapter 7.
Figure 5-1. Nitrogen cycle of the ACRU2000 model (adapted from Knisel et al. 1993)

Figure 5-2. Phosphorus cycle of the ACRU2000 model (adapted from Knisel et al. 1993)
Figure 5-3. Soil moisture response functions from the GLEAMS model

Figure 5-4. Soil temperature response functions from the GLEAMS model
CHAPTER 6
MODIFICATION OF THE FIELD-SCALE NITROGEN AND PHOSPHORUS
MODULE OF ACRU2000 FOR SHALLOW WATER-TABLE ENVIRONMENTS

Introduction

The nitrogen (N) and phosphorus (P) module of the ACRU2000 model may be most appropriate for application to upland environments as detailed in the previous chapter. Potential shortcomings of the model in general, and for application to shallow water-table environments in particular, were detailed. In this chapter modifications of the model for shallow water-table environments and for Florida flatwoods soils are made. These modifications address only a few of the potential shortcomings noted (Chapter 5). Specifically, modifications proposed include the effect of soil moisture contents on transformation rates, the extraction of nutrients into ponded/runoff water, and the factors that control the instantaneous, reversible sorption of P. Simplifications are made to the model in describing the mineralization of plant nutrients and the immobilization of nutrients by decomposing plant matter. A more rigorous description of plant growth, nutrient uptake, senescence, and residue deposition is currently under development but is not a part of this dissertation.

Nutrient Models

The nutrient models describe the mobilization and transport of dissolved forms of nitrogen and phosphorus. Sediment-bound nutrients are not represented in the model. In the case of N the soluble component is split between nitrate-N and ammonium-N, with mineralization of organic forms represented as a two-stage process. Soluble organic
nitrogen is not simulated in the model. P is represented as a single soluble form. Both ammonium-N and labile P undergo fast, reversible sorption to soil particles. Transport of soluble nutrients occurs between completely mixed soil layers and between the soil and adjacent water bodies. The exchange of nutrients with ponded/runoff water is assumed to occur within the top few centimeters of soil. The processes of litter fall, litter accumulation, and litter decay are assumed to be in equilibrium with plant uptake and residue immobilization. This assumption ignores the seasonal variation of these processes, but is considered to be a reasonable simplification for situations where harvest operations are largely absent.

The following sections describe the modifications made to the N and P module of the ACRU2000 model. A short description of the process and data objects added to the model can be found in Appendix E. UML diagrams of the process objects are presented in Appendix F. Appendix G is a technical manual for the N and P module.

**Phosphorus Model**

The P model is represented by three main pools (Figure 6-1), a labile pool \( P_l \), and two immobile forms, an “active” pool \( P_a \) and a slowly changing “stable” pool \( P_s \). This representation is much simpler compared to those proposed (e.g. Jones et al. 1984b; Knisel et al. 1993; Groenendijk and Kroes 1999) where organic forms are further described by additional pools which transform at various rates. In this model it is assumed that the transfer between the two immobile pools accounts for the net effects of slow adsorption/desorption of inorganic forms and mineralization/immobilization between organic and inorganic forms. The mass balance of the three pools can be written as:
\[
\frac{dP_r}{dt} = P_{\text{rain}} + P_{\text{fert}} + 0.75R_{awP} + R_{alP} - P_{\text{ro}} - P_{\text{gw}} - P_{\text{perc}}
\]
\[
\frac{dP_s}{dt} = R_{saP} - R_{alP}
\]
\[
\frac{dP_a}{dt} = -R_{saP} + 0.25R_{awP}
\]

where \( P_{\text{rain}} \) is the quantity of P in rainfall (kg/ha/day), \( P_{\text{fert}} \) is the rate of application of P in inorganic fertilizer (kg/ha/day), \( R_{awP} \) is the rate of decay of P in animal waste on the ground surface (kg/ha/day), \( R_{alP} \) is the rate of transformation from \( P_a \) to \( P_l \) (kg/ha/day), \( R_{saP} \) is the rate of transformation from \( P_s \) to \( P_a \) (kg/ha/day), and \( P_{\text{ro}}, P_{\text{gw}}, \) and \( P_{\text{perc}} \) are the quantities of P lost to runoff, groundwater flow, and percolation (kg/ha/day), respectively. The equations that govern the transformation rates are described in Chapter 5.

The reversible adsorption of P in sandy soils has been found to be related to soil properties such as double-acid-extractable magnesium, oxalate-extractable aluminum, oxalate-extractable iron, and organic carbon (Nair et al. 1998; Schoumans and Groenendijk 2000; Vadas and Sims 2002). Based on prior research in the Lake Okeechobee basin by Fraisse and Campbell (1997), the partitioning coefficients for A, E, Bh and Bw soil horizons are assumed to be a function of double-acid-extractable magnesium (mg/kg), \( M_{DA} \), oxalate-extractable aluminum (mg/kg), \( A_{OX} \), and organic carbon (%), OC. The linear partitioning coefficient for A horizons is defined as:

\[
K_d = e^{2.390} \text{ for } M_{DA} \geq 103.2
\]
\[
K_d = e^{-0.223} \text{ for } M_{DA} < 103.2 \text{ and } OC \geq 1.865
\]
\[
K_d = e^{1.442} \text{ for } M_{DA} < 103.2 \text{ and } OC < 1.865
\]

for E horizons:

\[
K_d = e^{4.241} \text{ for } A_{OX} \geq 496.45
\]
for Bh horizons:

\[ K_d = e^{3.751} \quad \text{for} \quad Al_{OX} \geq 1327.5 \]  
\( K_d = e^{2.195} \quad \text{for} \quad Al_{OX} < 1327.5 \)  
(6-6b)

For Bw horizons:

\[ K_d = e^{3.212} \quad \text{for} \quad Al_{OX} \geq 570.8 \]  
\[ K_d = e^{1.604} \quad \text{for} \quad Al_{OX} < 570.8 \]  
(6-7b)

Nitrogen Model

The nitrogen model is represented by four main pools, nitrate-N (\(N_{NO3}\)), ammonium-N (\(N_{NH4}\)), active organic N (\(N_a\)), and stable organic N (\(N_s\)) (Figure 6-2). As with the P model, the N model represented here is much simpler compared to others that have been proposed (Johnsson et al. 1987; Hansen et al. 1991; Knisel et al. 1993; Groenendijk and Kroes 1999). The mass balance of the four pools can be written as:

\[
\frac{dN_{NO3}}{dt} = N_{rain} + N_{fertNO3} + R_{nit} - R_{denit} - N_{roNO3} - N_{gwNO3} - N_{percNO3} 
\]  
(6-9)

\[
\frac{dN_{NH4}}{dt} = N_{fertNH4} + 0.8R_{awN} + R_{amminN} - R_{nit} - R_{vol} - N_{roNH4} - N_{gwNH4} - N_{percNH4} 
\]  
(6-10)

\[
\frac{dN_a}{dt} = 0.2R_{awN} + R_{saN} - R_{amminN} 
\]  
(6-11)
\[ \frac{dN_s}{dt} = -R_{saN} \]  

(6-12)

where \( N_{\text{rain}} \) is the quantity of nitrogen in rainfall (kg/ha/day), and is assumed to occur completely as nitrate, \( N_{\text{fertNO3}} \) and \( N_{\text{fertNH4}} \) are nitrate and ammonium quantities in inorganic fertilizer (kg/ha/day), \( R_{awN} \) is the rate of animal waste nitrogen decay (kg/ha/day), \( R_{nit} \) is the rate of nitrification of ammonium to nitrate (kg/ha/day), \( R_{\text{denit}} \) is the rate of loss of nitrate to the atmosphere by denitrification (kg/ha/day), \( R_{\text{ammon}} \) is the rate of ammonification of the active nitrogen pool to ammonium (kg/ha/day), \( R_{\text{vol}} \) is the rate of loss of ammonium in manure to the atmosphere by ammonia volatilization (kg/ha/day), \( R_{saN} \) is the rate of transformation of stable N to active N (kg/ha/day), \( N_{\text{roNO3}} \) and \( N_{\text{roNH4}} \) are the quantities of nitrate and ammonium lost in runoff (kg/ha/day), \( N_{\text{gwNO3}} \) and \( N_{\text{gwNH4}} \) are the amount of nitrate and ammonium lost in groundwater flow (kg/ha/day), and \( N_{\text{percNO3}} \) and \( N_{\text{percNH4}} \) are the quantities if nitrate and ammonium in percolating water (kg/ha/day). The equations describing the transformation rates are detailed in Chapter 5.

**Nutrient Transformation Response to Soil Moisture**

The use of soil moisture response factors in simulation models is inherently approximate as the response of microbial activity to soil moisture conditions is a function of a number of factors. The relationship between soil moisture and microbial activity has been shown to vary between soils, depending on the shape of the soil moisture curve, the abundance of organic matter, pH, and depth (Goncalves and Carlyle 1994; Rodrigo et al. 1997; Leiros et al. 1999).

There are several mechanisms that cause a decrease in microbial activity in dry soil. These include reduced mobility of both soluble substrate and microbes, and a direct effect
of dryness on microbial growth and survival. Under low soil moisture conditions a reduced rate of decomposition is caused by two factors: first, as the pores within the soil dry and the water film coating the particle surfaces becomes thinner, diffusion path lengths become more tortuous and the rate of both substrate and microbe diffusion declines; second, low water contents correspond to low water potentials that lower intracellular water potentials which in turn reduce hydration and enzymatic activity (Porporato et al. 2003).

Under wet conditions a decrease in aerobic microbial activity is caused by a reduction of oxygen diffusion (Grant and Rochette 1994). During periods of high soil moisture anoxic conditions prevent bacteria from aerobically oxidizing organic matter. Anaerobic respiration has been shown to be approximately one-third of aerobic respiration, regardless of substrate quality (DeBusk and Reddy 1998).

Rodrigo et al. (1997) note that defining functions in terms of water pressure allows for comparison between soils of different textures, and that using soil water contents may be more useful in describing processes that can limit microbial activity in soils such as solute and oxygen diffusion, while expressing functions in terms of water filled pore space, or relative saturation, appears to be the best indicator of aerobic/anaerobic microbial activity.

The soil moisture conditions that yield optimal decomposition and mineralization rates have been reported to occur at soil water pressure heads between 100 and 500 cm (Rodrigo et al. 1997). Kladivko and Keeney (1987) have shown that mineralization rates could be well represented as a linear function of water content or a logarithmic function of soil water pressure head. Soil moisture response functions for ammonification and P
mineralization are described as logarithmic functions of soil water pressure head (Figure 6-3) as is done in the models of Hansen et al. (1991), Rijetma and Kroes (1991), and Vanclooster et al. (1996). Expressed in units of $pF$ ($\log_{10}$ of negative pressure head in units of cm):

$$f_{\theta_{\text{min}}} = \frac{pF_{wp} - pF}{pF_{wp} - 2.7} \quad \text{for} \quad pF > 2.7 \quad (6-13a)$$

$$f_{\theta_{\text{min}}} = 0.6 + 0.4 \frac{pF - pF_{s}}{2 - pF_{s}} \quad \text{for} \quad pF < 2 \quad (6-13b)$$

$$f_{\theta_{\text{min}}} = 1 \quad \text{for} \quad 2 \leq pF \leq 2.7 \quad (6-13c)$$

where $pF_{wp}$ is the $pF$ value at the wilting point (15000 cm), $pF_{s}$ is the $pF$ near saturation (taken as 1 cm for mathematical reasons). The response to soil moisture rises from zero at the wilting point to an optimum between $pF$ of 2 and 2.7 (100 cm and 500 cm of soil water pressure head, respectively), followed by a decrease to a minimum of 0.6 under saturated conditions.

Similarly, nitrification is represented as a logarithmic function of soil water pressure head (Hansen et al. 1991; Rijetma and Kroes 1991; Vanclooster et al. 1996) but reduces to zero as saturated conditions are approached (Linn and Doran 1984; Skopp et al. 1990) (Figure 6-3):

$$f_{\text{nit}_{\text{min}}} = \frac{pF_{wp} - pF}{pF_{wp} - 2.7} \quad \text{for} \quad pF > 2.7 \quad (6-14a)$$

$$f_{\text{nit}_{\text{min}}} = \frac{pF - pF_{s}}{2 - pF_{s}} \quad \text{for} \quad pF < 2 \quad (6-14b)$$

$$f_{\text{nit}_{\text{min}}} = 1 \quad \text{for} \quad 2 \leq pF \leq 2.7 \quad (6-14c)$$

The soil moisture response of denitrification is simulated as a function of relative saturation as proposed by Johnsson et al. (1987) as used by Vanclooster et al. (1996) and similar to that developed by Rolston et al. (1984) (Figure 6-3):
\[ f_{\text{dil}} = \left[ \frac{\theta - \theta_d}{\theta_s - \theta_d} \right]^d \] 

(6-15)

where \( \theta_d \) is a threshold water content which defines the water content above which denitrification occurs and is assumed to correspond to an effective saturation of 0.8, and \( d \) is an empirical exponent assumed to be a value of 2 (Vanclooster et al. 1996).

**Extraction of Nutrients into Runoff**

During runoff events water entrains some of the soil porewater, this “extraction” of porewater solutes has been shown to occur principally near the soil surface and rapidly diminishes with depth (Ahuja et al. 1981). The exchange of solutes between the soil and ponded or runoff water has been simulated as a convective mass transfer across a thin boundary layer at the soil-water interface (Wallach et al. 1988; Havis et al. 1992) or as an enhanced diffusion process within the soil (Parr et al. 1987; Ahuja 1990). The rate of mass transfer has been related to the depth of overland flow, solute diffusion coefficients, soil permeability, runoff shear velocities, and the energy of rain impact (Parr et al. 1987; Richardson and Parr 1988; Wallach et al. 1989; Gao et al. 2004). However these models, both numerical and analytical, have to date only been applied to individual, controlled laboratory events and have yet to be integrated into a continuous field-scale model.

In lieu of such a complex approach several approximate models have been adopted. Early modeling efforts assumed that soil water within a thin zone of surface soil mixes completely and instantaneously with runoff (Crawford and Donigian 1973; Steenhuis and Walter 1980). The depth of this surface zone was often a calibrated parameter and has been shown to vary between values of 0.3 cm to over 5 cm in laboratory experiments (Ahuja and Lehman 1983; Snyder and Woolhiser 1985) with values increasing as free drainage is reduced during the runoff event. Some observations showed that solute
concentrations were usually much lower in runoff water compared to those in soil solution even at low infiltration rates (Snyder and Woolhiser 1985). In response, some modeling efforts considered a surface soil zone that mixes incompletely with runoff with the degree of mixing determined by an empirical extraction coefficient (Frere et al. 1980; Leonard et al. 1987). Ahuja and Lehman (1983) suggested that this degree of soil water mixing should be an exponential function of depth, however the maximum depth of interaction and the rate of decrease in the extraction coefficient with depth being, in practice, calibrated parameters.

The extraction of solutes into runoff in the ACRU2000 model is assumed to occur from the top 1 cm of soil. The runoff and soil water are assumed to be incompletely mixed with the degree of mixing defined by an extraction coefficient, $\beta$ which ranges from 0.1 to 0.5 as a function of the partitioning coefficient (Equation 5-50). This extraction of solutes into runoff assumes the runoff (or ponded) water to have an initial solute concentration of zero (Ahuja and Lehman 1983). Additionally, runoff and the solutes within are assumed to leave the field on the same day they were generated. To better reflect the slow runoff response typically seen in flatwoods sites (often lasting several days), the exchange of solutes between water on the ground surface and water in the top soil is considered to be a function of the concentration differences between ponded and soil water. Thus the concentration of solutes in runoff or ponded water is assumed to be the equilibrium concentration between water on the ground surface and an incompletely mixed top soil layer, or effective depth of interaction ($EDI$):

$$C_{eq} = \frac{C_s \theta_s \beta \cdot EDI + C_w H}{\theta_s \beta \cdot EDI + H}$$

where $C_{eq}$, $C_s$, and $C_w$ are the solution concentrations (mg/L) at equilibrium, in the top
soil layer, and in ponded water, respectively, $\theta_s$ is the saturated water content of the top soil layer (cm$^3$/cm$^3$), $EDI$ is the thickness of the top soil layer (also referred to as the effective depth of interaction) (mm), $H$ is the depth of ponded water on the ground surface (mm), and $\beta$ is the extraction coefficient which is assumed to be a constant value of 0.5. Mathematically $\beta$ can be interpreted as the fraction of the effective depth of interaction that is completely mixed with runoff water. Due to the experimental evidence showing that the depth of soil that interacts with runoff increases as infiltration rates decrease (Ahuja and Lehman 1983) the effective depth of interaction ($EDI$) is allowed to vary and is a calibrated parameter. This is considered to be appropriate for regions where runoff is generated primarily by saturation-excess such as the flatwoods compared to the infiltration-limited runoff generation mechanism as simulated by the ACRU2000 model (Chapter 2).

**Summary**

Modifications to the field-scale nitrogen and phosphorus module of the ACRU2000 model were proposed for shallow water-table environments and sandy flatwoods soils. The response of nutrient transformations to soil moisture conditions was changed in order to define an optimal range of water contents and to better reflect the effect of wet conditions on mineralization processes. The extraction of nutrients into ponded/runoff water was changed in order to reflect concentration gradients between soil water and ponded/runoff water and to define the effective depth of soil that interacts with surface water as a calibratable parameter. Additionally, the factors that affect the instantaneous, reversible sorption of P were changed to better reflect the sandy, low-clay soil horizons
of flatwoods soils. Finally, the model was simplified by assuming that plant uptake of nutrients is in equilibrium with the mineralization of plant residue.

The modified N and P module is evaluated for use in shallow water-table flatwoods sites in the next chapter. The model proposed here is also compared to the unmodified model described in Chapter 5 in order to gauge improvements.
Figure 6-1. Conceptual model of the phosphorus cycle

Figure 6-2. Conceptual model of the nitrogen cycle
Figure 6-3. Soil moisture response functions for ammonification (and P mineralization), nitrification, and denitrification
CHAPTER 7
FIELD-SCALE VALIDATION OF THE NITROGEN AND PHOSPHORUS MODULE OF THE ACRU2000 MODEL FOR SHALLOW WATER-TABLE ENVIRONMENTS

Introduction

The eutrophication of Lake Okeechobee in southern Florida has been attributed primarily to phosphorus (P) loads in agricultural runoff (Anderson and Flaig 1995) and has been shown to impact the ecological condition of the lake (Steinman et al. 1999). A large proportion of the P loading has been identified as coming from drainage areas north of the lake (Federico et al. 1981; SFWMD 1997; USEPA and SFWMD 1999). Beef cow-calf operations make up 51% of the Lake Okeechobee Basin, making them a large contributor of P to the lake (Hiscock et al. 2003). Several remediation projects have been initiated in order to restore the lake by improving water quality and lake ecosystem functions. In order to meet the target in-lake P level set for Lake Okeechobee (40 ppb) these projects are aimed at reducing both the internal and external P loads to the lake (USEPA and SFWMD 1999).

In order to determine which management practices will prove most effective in reducing external P loading to the lake, long-term, continuous models are often used to evaluate Best Management Practices (BMPs) prior to implementation. Due to the unique hydrology of the flatwoods watersheds (flat topography and poorly drained, sandy soils with shallow water-tables) that make up a large portion of the lake’s drainage basin the traditional models used in other locations in the U.S. often perform poorly (Heatwole et al. 1987). In addition, many models have had difficulty in predicting P loads in
flatwoods regions due to the limited P retention capacity of the region’s soils (Graetz and Nair 1995).

The ACRU2000 model is a distributed, daily time-step, object-oriented model that has recently been expanded for use in shallow water-table environments (Chapter 3). Hydrologic modifications made to the model include the simulation of a fluctuating water-table by assuming soil moisture to be in hydrostatic equilibrium with the water-table as has been done in similar models (Skaggs 1980; Koivusalo et al. 2000). The contribution of a shallow water-table to an upward gradient induced by evapotranspiration is simulated using the approximate, algebraic relationship of Anat et al. (1965). Reference potential evapotranspiration can be simulated using the standardized Penman-Monteith equation adopted by the Food and Agricultural Organization (Allen et al. 1998). Plant root distributions are simulated using the relationship proposed by Hoogland et al. (1981). Plant evapotranspiration response to soil moisture is simulated as a function of soil water pressure head within the root zone using the relationship of Feddes et al. (1978). Runoff is assumed to occur by saturation excess only and is routed from the land surface using a simple power law relationship (Kroes and van Dam 2003). Groundwater inflow/outflow is simulated in response to a time-varying boundary condition using the Dupuit equation (Fetter 1994) or to or from a deep aquifer below a restrictive layer using Darcy’s Law.

The ACRU2000 model represents the major N and P components and transformations using algorithms from the GLEAMS model (Chapter 5; Knisel et al. 1993). Modifications to these algorithms that may be considered more appropriate for shallow water-table environments and Florida flatwoods soils were proposed in Chapter
6. The objective of this work is to test the suitability of the ACRU2000 model, with the proposed modifications, to predict N and P loads from field sites in the flatwoods of the Lake Okeechobee Basin. The model is validated using observed N and P runoff load data from six experimental sites. Model evaluation is made by its ability to predict annual N and P loads in runoff. The model, as described in Chapter 6 is also compared to the unchanged algorithms in Chapter 5 in order to judge any improvement in the model’s predictive ability.

Model Validation

Site Description

The sites chosen for validation of the model are experimental pastures within the MacArthur Agro-Ecology Research Center (MAERC) at Buck Island Ranch in Highlands County, Florida (Figure 7-1). The site is situated at 27° 7.9’ N and 81° 12.3’ W, approximately 21 km northwest of Lake Okeechobee in the C41 basin. The Center consists of an array of 16 pastures divided between two typical land uses, improved and semi-improved pasture (Figures 7-2 and 7-3). The improved pastures are vegetated primarily with bahia grass (*Paspalum notatum*) and the semi-improved pastures are composed of a mixture of bahia grass and native herbaceous vegetation including carpet grass (*Axonopus furcatus*), broomsedge (*Andropogon virginicus*), bluestem (*Andropogon glomeratus*), and field paspalum (*Paspalum laeve*) (MAERC 2004). Historically the improved pastures were fertilized with recommended amounts of N, P, and K from the 1970s to 1987. After 1987 the improved pastures were fertilized each year in the spring with N only (56 kg N/ha) (MAERC 2004). The improved pastures had also been periodically limed every 3-5 years. The semi-improved pastures were believed to have never been fertilized (MAERC 2004). Improved pastures are approximately 20.2 ha and
semi-improved pastures are approximately 32.4 ha. The improved pastures are grazed primarily in the summer wet season (May – October) and the semi-improved in the winter dry season (November – April) (hereafter referred to as summer and winter pastures).

The terrain of the pastures is nearly flat (<0.1% slope) and is drained by a series of shallow ditches with the summer pastures more intensely drained compared to the winter pastures. The elevation of the pastures range from 7.9 to 8.5 m above mean sea level with the summer pastures generally 10-15 cm lower than the winter pastures (MAERC 2004). The pastures slope gently towards Harney Pond Canal, a major regional conveyance linking Lake Istokpoga to the north and Lake Okeechobee to the south.

Soil surveys of the area were conducted by the USDA-NRCS in June 1997, at a 0.5-ha resolution. Soils in the summer pastures are predominantly Felda fine sand, a loamy, siliceous, hyperthermic Arenic Endoaqualfs and are predominantly Pineda fine sand, a loamy, siliceous, hyperthermic Arenic Glossaqualfs, with 60% coverage of a thin (2.5-15 cm) muck layer in the winter pastures (MAERC 2004).

**Experimental Design**

The pastures are divided by an earthen berm (4 m wide, 0.5 m above grade). The original drainage ditches were connected to two existing ditches and to new collection ditches allowing runoff from each pasture to flow through an exit flume. Flume elevations were set at 7.99 and 8.08 m above mean sea level in the summer and winter pastures, respectively. Runoff from each pasture was determined at the trapezoidal flumes from water level measurements made in stilling wells at both the upstream and downstream end of the flume in 20 minute intervals (Capece et al. 1999). Water samples were taken using ISCO automatic water samplers at intervals based on flow volume and
hydrograph shape (Tremwel et al. 1996). Samples were analyzed for total phosphorus, nitrate/nitrite, ammonium, and total Kjeldahl nitrogen. Meteorologic data were collected on an hourly basis at weather stations adjacent to the pastures. Rainfall, temperature, solar radiation, relative humidity, wind speed, and wind direction were collected at each station. Groundwater levels were measured at 15-minute intervals in monitoring wells in the pastures. Three pastures in both the winter and summer pastures were instrumented with two monitoring wells, one 4-inch diameter well at the downstream end of the pasture near the flume and one 2-inch diameter well located in the center of the pasture (Figures 7-2 and 7-3). The remaining ten pastures were instrumented with one monitoring well, a 2-inch diameter well located in the center of the pasture (Figures 7-2 and 7-3). Wells extended to a depth of 18 ft below ground surface with the screened portion beginning at 5 ft below ground surface. Limited groundwater quality measurements were made. Runoff and climatic data collection began in May 1998 and groundwater level measurements began in September 2000. Data collection continued until the end of 2003.

Each pasture type had two replicates of four different cattle stocking rate treatments during the study period: 0, 15, 20, and 35 cow-calf pairs. The stocking rate of each pasture is shown in Figures 7-2 and 7-3. Cattle were stocked in the summer and winter pastures for a total of 1025 and 677 days during the study period, respectively. During the study period nitrogen fertilizer was applied at 56 kg N/ha in May 2000, April 2001, and March 2003 to the summer pastures. No fertilizer was applied to the winter pastures. Prescribed burning was conducted in the winter pastures in November 1998 and February 2002 and in the summer pastures in February 1999 and April 2002. For validation of the
model six pastures were selected for simulation. These include three pastures from each
land use, one control pasture (no stocking), one low stocked pasture (15 pairs), and one
high stocked pasture (35 pairs). The control pastures are winter pasture 7 (WP7) and
summer pasture 1 (SP1), the low stocked pastures are winter pasture 6 (WP6) and
summer pasture 4 (SP4), and the high stocked pastures are winter pasture 5 (WP5) and
summer pasture 3 (SP3) (Figures 7-2 and 7-3). These pastures were selected in order to
first evaluate the model’s ability to predict nutrient loads resulting from nutrients already
present in the soil and then to evaluate the model’s ability to predict the contribution from
grazing cattle.

Model Calibration

Model hydrologic calibration for each pasture was conducted using the observed
runoff and groundwater level data from 1998 to 2001. The length of the calibration
period was chosen in order to include adequate groundwater level data (data collection
started 9/2000). This period allowed for 16 months of groundwater level data to be used
in the calibration. The remaining two years, 2002 and 2003, were used to verify the
model calibration. Hydrologic calibration was conducted (as described in Chapter 4) first
for WP6 where values of saturated water contents ($\theta_s$) and van Genuchten (1980) “n” soil
parameters (Table 7-3) were reduced slightly from curve fits of published moisture
contents for the Pineda fine sand in neighboring Glades county (Sodek et al. 1990).
These changes were made in order to better reflect the magnitude of observed water-table
fluctuations and were considered to be appropriate as no field data were collected. The
changes made fell within the observed ranges for Pineda fine sand and closely associated
Felda, Malabar, Myakka, and other soils reported by Sodek et al. (1990). Hydraulic
conductivities of the A, E, and Bw soil layers (Table 7-1) were reduced one order of magnitude from that reported by Sodek et al. (1990) in order to more accurately represent the effect of the stage in the adjacent canal. The remaining hydrologic calibration consisted of changing the runoff resistance in order to better match runoff event peaks and duration. Model calibration was made by graphical comparison between observed and simulated daily values. WP7 and WP5 were then calibrated with only slight changes to the runoff resistance only. Hydrologic calibration of SP1, SP4, and SP3 consisted of manipulating the runoff resistance. No changes were made to hydraulic conductivities or the soil moisture relationships of the Felda fine sand (Table 7-2 and 7-4) fitted from the data of Sodek et al. (1990) for the summer pastures.

Observed groundwater level data from the two wells in WP6 and SP1 are shown in Figures 7-4 and 7-5. Observations were quite similar at both locations, near the flume and at the center of the pasture. As can be seen there are periods where there is a notable gradient between the two wells, indicating that groundwater level is affected by the neighboring canal stage. Figures 7-6 and 7-7 show the canal stage reported at the S70 spillway located four kilometers downstream compared to the groundwater level recorded at the 4-inch well located near the flume for WP6 and SP1 and the 2-inch well located mid-pasture for WP7, WP5, SP4, and SP3. As observed, the gradient between the groundwater level and the canal reverses direction, with groundwater flowing towards the canal during wet periods (typically summer) and canal water flowing towards the pasture during dryer periods. The daily time series of canal stage serves as an input to the model.

Soil physical and chemical properties are shown in Tables 7-1, 7-2, 7-5, and 7-6 for the Pineda fine sand found in the winter pastures and the Felda fine sand found in the
summer pastures. Soil phosphorus values were reported by Capece et al. (2003) for the A horizon of each pasture and were reported for the subhorizons by Hill (2003). Phosphorus pools were initialized according to the procedures of Fraisse and Campbell (1997) and Knisel et al. (1993). Initial pool sizes of labile P compared well to the values determined by Hill (2003) and compared well to double-acid extractable P values as reported by MAERC (2004). Values of double-acid-extractable Mg and oxalate-extractable Al were assumed to be the same for horizons in both pastures and were assumed to have similar values in A, E, and Bw horizons as Spodosols in the region. Values reported by Nair et al. (1998) were used since no field data were collected from the pastures. No soil N data were collected from the pastures. Ratios of total N to total P in the pastures were assumed to be the same as those reported by Nair et al. (1995) for pasture landuse in the Lake Okeechobee basin and a single value was assumed for the winter and summer pastures. N and P contents in rainfall were determined to be 0.20 and 0.03 ppm, respectively based on data from NADP (2005) and USEPA and SFWMD (1999).

Nutrient calibration of the model was first conducted for the control pastures WP7 and SP1. The depth of soil interacting with runoff water was assumed to be 1 cm, as in the original ACRU2000 model (Chapter 5), for WP7. For SP1 this depth was increased to 3 cm in order to better match field data. This increase may be justified due to the more intensive drainage network of the summer pastures (MAERC 2004), allowing for a more direct route for shallow subsurface water into the ditches and out of the pasture as runoff. These values were used for simulating the remaining pastures without further calibration. Application of the original, unmodified N and P module (Chapter 5) was made using the
same approximations as that proposed in the modified model (Chapter 6), that the uptake of nutrients by the plant was in equilibrium with the mineralization of fresh organic material. This was done in order to provide results that would be more directly comparable.

**Results**

Observed and simulated groundwater levels are shown in Figures 7-8 to 7-13. The discrepancy between observations and simulated water-table depths, particularly during periods of deeper observed water-tables, may be due to the representation of the canal’s influence on groundwater levels within the pasture, the uncertainty of soil hydraulic parameters due to the lack of field-collected data, and the assumption of no groundwater inflow or outflow from upland areas. No stage measurements of the canal were made at the site, stage measurements at the S70 spillway were assumed to represent the local canal stage by assuming level pool conditions over the 4 km distance. No groundwater measurements were made in areas upland from the pastures. As can be seen the model followed the general trend in water-table fluctuations, with both periods of over and under prediction. Observed and simulated daily runoff can be seen in Figures 7-14 to 7-19 and reported as cumulative annual plots in Figures 7-20 to 7-25. As shown the calibration (1998-2001) matched observations, with the exception of the very dry year of 2000. Verification (2002-2003) matched the observed data less well compared to the calibration period, but general trends in runoff timing and amounts were followed with larger variability observed in the winter pastures.

Annual runoff totals from each pasture are shown in Tables 7-7 and 7-8. Mean absolute error (MAE), root-mean square error (RMSE), and Nash-Sutcliffe coefficients of efficiency (E) (Nash and Sutcliffe 1970) using these annual values are reported in Table
MAE and RMSE values ranged from 28.2 to 81.2 mm and 33.4 to 103.1 mm, respectively. Nash-Sutcliffe efficiency values indicate that the model was a valuable tool for estimating annual runoff depths for all pastures (range between 0.722 and 0.987).

Observed and simulated N and P loads from WP7 (control) are shown in Figures 7-26 and 7-27. N loads generally matched field observations, while P loads were underpredicted in 2001 and 2002. Observed and simulated N and P loads from WP6 (low stocking) are shown in Figures 7-28 and 7-29. N loads followed observations, while P was under-predicted in 2001. N and P loads from WP5 (high stocking) are shown in Figures 7-30 and 7-31. N loads again generally followed observations, while P was under-predicted in 2001, 2002, and 2003. It is also of interest to note the apparent influence on cattle stocking in this pasture during the winter of 2002-2003, where larger magnitude loads are simulated during the stocking period.

Observed and simulated N and P loads from SP1 (control) are shown in Figures 7-32 and 7-33. The simulated N load agreed with observed data with the exception of 2001, however there was a large under-prediction of P in 2001 and 2002. In Figures 7-34 and 7-35, the effects of stocking can be seen on SP4. Over-predictions can be seen of N and P in 1999 and 2003. P was also under-predicted in 2001 and 2002. The effects of stocking on simulated loads is further illustrated for SP3 in Figures 7-36 and 7-37 where N is over-predicted in all years and P is over-predicted in 1999 and 2003, but (interestingly) matches observed values well in 2001 and 2002.

Annual totals of N and P loads are shown in Tables 7-7 and 7-8. Annual average N and P concentrations (determined as the annual load divided by the annual runoff depth and converted to units of mg/L) are shown in Tables 7-9 and 7-10. Mean absolute error
(MAE), root-mean square error (RMSE), and Nash-Sutcliffe coefficients of efficiency using these annual values are reported in Table 7-11. MAE and RMSE values for N and P loads were greater for the summer pastures compared to the winter. Coefficients of efficiency values indicate the model was a good predictor of N and P loads for all winter pastures and a poor predictor of both N and P for all summer pastures with the exception of P in SP4. E values also show that the model was a poor predictor of average N and P concentrations in all pastures.

For comparison, the annual simulated loads using the unmodified N and P algorithms of the ACRU2000 model, as described in Chapter 5, are also shown in Tables 7-7 and 7-8. The simulated annual average N and P concentrations in runoff are shown in Tables 7-9 and 7-10. N and P loads and concentrations predicted by the unmodified ACRU2000 were consistently lower than observations and that simulated by the modified model. Mean absolute error (MAE), root-mean square error (RMSE), and Nash-Sutcliffe coefficients of efficiency using these annual values are reported in Table 7-12. MAE and RMSE values of N and P loads were higher compared to the modified model (Table 7-11) with the exception of N and P loads in SP1 and SP3 and N load in SP4. MAE and RMSE were higher for N loads in the winter pastures and higher for P loads in the summer pastures. Coefficients of efficiency show the model to be a poor predictor for N and P loads and N and P concentrations for all pastures with the exception of N load in SP4. Coefficients of efficiency were higher than that for the modified model (Table 7-11), indicating that the unmodified model was a poorer predictor, with the exception of N loads in SP4 and SP3 and N and P concentrations in SP3.
Sensitivity Analysis

Obtaining accurate values for model input variables can be costly or impractical for field-scale models, particularly for large sites where spatial variation in properties is likely. The accuracy of model input parameter values is usually proportional to the time and resources invested in determination. Since model results will be more sensitive to certain inputs compared to others it is useful to perform a sensitivity analysis in order to establish priorities in collecting and determining model parameters.

An analysis was performed to determine the sensitivity of model simulation output of N and P loads to the hydrologic input parameters in Table 7-13 and to the nutrient input parameters in Table 7-14. The sensitivity analysis was performed by using the six-year simulation of SP4. Model sensitivity was determined for ± 25%, 50%, 75%, and 100% of the base input value (shown in Tables 7-13 and 7-14). For cases where this range of variation was unfeasible, or unrealistic, the results were omitted. Model sensitivity is reported as the percent difference of model results as compared to the base simulation.

The sensitivity of N and P loads to the hydrologic parameters in Table 7-13 is shown in Table 7-15. Input parameters showing very low sensitivity on N and P loads were root depth, root distribution parameter, vertical saturated hydraulic conductivity, upward flux exponent, the transpiration factor for water deficiency, interception capacity, and the initial depth of the water-table (this value was not increased due to the relatively deep initial value used). Parameters showing relatively low sensitivity with the exception of very large positive or negative changes include the soil moisture shape parameter n, bubbling pressure head, the transpiration reduction factor due to oxygen deficiency, and runoff resistance. The remaining parameters, in approximate order of increasing
sensitivity were the horizontal hydraulic conductivity, the runoff exponent (limited to values greater than 1), soil shape parameter \( \alpha \), saturated water content, and the crop coefficient.

The sensitivity of N and P loads to soil parameters (not including soil N and soil P parameters) was greatest to the application rate of manure and the effective depth of interaction, and was large for P loads for large increases in pH due to the effect pH has on the P availability index (PSP) as shown in Equation 5-17 (Table 7-16). Base saturation, clay content, and organic matter content showed no sensitivity over the range of values tested.

The sensitivity of N and P loads to N and P parameters is shown in Table 7-17 and 7-18. N loads showed greatest sensitivity to nitrogen concentrations in rainfall, and to total N and active N contents. N loads were slightly sensitive to the nitrate extraction coefficient and ammonium partitioning coefficient. P loads were not affected by the N parameters. P loads were most sensitive to total phosphorus, stable phosphorus, and the phosphorus availability index (Table 7-18). With the exception of the concentration of P in the nearby stream, P loads were slightly sensitive to all other P parameters with the largest sensitivity occurring at the extreme high or low values. N loads showed no sensitivity to the P parameters over the range of values tested.

To summarize, the hydrologic parameters with the largest sensitivity on N and P loads in runoff were the crop coefficient, saturated water contents, \( \alpha \) soil water shape parameters, runoff exponent, depression storage, and horizontal saturated hydraulic conductivities (Figures 7-38 and 7-39). In addition to these parameters, N loads were most sensitive to manure application rates, the effective depth of interaction, nitrogen
concentrations in rainfall, and total and active nitrogen contents (Figure 7-40). In addition to the hydrologic parameters, P loads were most sensitive to high values of pH, manure application rates, the effective depth of interaction, total and stable phosphorus contents, and the P availability index (Figure 7-41).

The sensitivity of N and P loads to the hydrologic parameters in Table 7-13 is largely due to the sensitivity of runoff volumes to these parameters. Decreasing crop coefficients causes a decrease in evapotranspiration, resulting in greater runoff. Increases in $\theta_s$ and $\alpha$ increases the volume of empty soil pores at a given water table depth, decreasing the amount of runoff produced and resulting nutrient loads. An increase in $z_{dep}$ also increases the volume that must be filled prior to runoff occurring, decreasing the total volume of runoff and loads in runoff.

As would be expected, N and P loads both increased with increased manure application rates. Interestingly, P loads increased with increasing EDI, while N loads decreased. This indicates that N became more diluted when the effective depth of soil that interacts with ponded water increased. The sensitivity of N loads on rainfall concentrations indicates its relatively large contribution to the nitrogen balance in these pastures. N loads also decreased with increasing values of TN and $N_a$. This is likely due to the use of the ratio of $N_a$ to TN in determining the transfer of N from $N_a$ to $N_s$ (Equation 5-32). P loads increased with increasing TP and $P_s$, showing the effect of P already present in the soil and the contribution of stable forms to produce labile P. The large sensitivity of P loads to PSP indicates the importance of the equilibrium between labile P and active P forms.
Discussion

While the prediction of runoff volumes by the model were in general agreement with observations, and that of water-table depths adequate considering the lack of field-collected soil hydraulic data and the approximation of the pastures as single lumped elements in the model, the validation of the model elucidates three major shortcomings in simulating N and P loads. First, the model’s inability to capture the large loads of P observed in most pastures during 2001. Second, the prediction of elevated loads of N and P in response to the increased stocking rates compared to observations. Third, the overall poor performance of the model and the relatively poorer performance of the model in simulating the summer pastures as compared to the winter. The model did however produce annual N and P load and concentration predictions that were in better agreement with observations compared to the original N and P module of ACRU2000 as described in Chapter 5.

The inability of the model to match elevated observed loads in 2001 may be due to the lack of a complete representation of organic nutrient cycling in the model. A complete representation would include the partitioning of biomass between above ground and below ground components. These components would respond to environmental conditions such as water-excess or deficient stress, temperature as it affects the plant growth stage, and nutrient availability. In response to environmental stresses, portions of the above ground and below ground biomass should experience die-back and senescence, contributing to decomposable biomass and nutrient pools (where it may mineralize or immobilize nutrients). It is also important to note the pastures contain wetlands which may accrete and store biomass and nutrients during periods of anaerobiosis. During dryer periods this accreted organic matter would decompose at an accelerated rate by aerobic
microbes. As 2000 was a relatively dry year it is postulated that organic soils that had been accreting within the wetlands and drainage ditches within the pasture were mineralized at an enhanced rate. The biomass built in 1998 and 1999 (and prior to the beginning of the model simulation) may have largely decomposed, releasing large quantities of nutrients in the process that were subsequently flushed from the pasture in 2001. This may be supported by the fact that, as can be seen in Tables 7-9 and 7-10, 2000 and 2001 generally had the highest average observed N and P concentrations for most pastures. After 2000, the process of organic soil accumulation may have reinitiated, binding nutrients in the process and reducing the simulated loads in later years. As simulated here the process of mineralization of nutrients from decomposing plant material was assumed to be in equilibrium with plant uptake and immobilization by organic material. In ACRU2000, as based on the GLEAMS algorithms, fresh organic matter is currently only available for decomposition at the beginning of the simulation period and after harvest operations (Chapter 5) rather than being represented as a continuum of growth, senescence, litter fall, organic soil accretion and decomposition.

The prediction of elevated loads in later years in the stocked pastures may have several possible causes. For both N and P it may be due to a lack of organic matter building as described above. There may also be specific reasons for both N and P.

For the case of nitrogen, the elevated loads in later years may be due to under-prediction of ammonia volatilization and denitrification of nitrate. The nitrogen model allows for ammonia volatilization to occur only from ammonium from animal waste, and only for a period of one week (Knisel et al. 1993). However, it is well known that volatilization can be a large component of nitrogen loss from wetland environments.
Denitrification, which occurs under reducing conditions, may also be enhanced in wetland environments. The accurate prediction of these two processes is further complicated by the inability of the nitrogen algorithms, as incorporated from the GLEAMS model, to adequately predict the nitrification of ammonium to nitrate (data not shown). The model largely over-predicted nitrate in runoff loads, indicating a large over-prediction in nitrification. The process of nitrification can only occur in the presence of oxygen, indicating the importance of representing reducing conditions. Due to the lack of field collected N data and/or site- or region-specific observed process rates or responses to abiotic factors such as temperature or soil moisture it is not possible to determine the source of the error.

In the case of P the increased simulated loads in higher stocked pastures may be due to the rate at which labile P is transformed into more immobile forms. The apparent sorption of P is usually conceptualized as a combination of a fast (almost instantaneous) reversible true sorption onto particle surfaces and various slower time-dependent processes where P diffuses into and is deposited below surfaces that may not be completely reversible (McGechan and Lewis 2002). It has been shown that the fast, reversible sorption of P is highly non-linear and should be represented mathematically with non-linear relationships such as the Freundlich or Langmuir isotherms (Graetz and Nair 1995; Mansell et al. 1995; McGechan and Lewis 2002). The slower time-dependent processes are sometimes divided into a relatively faster component (the exchange between the labile and active pools) and a very slow component (exchange between the active and stable pools). The rate at which this slow adsorption or desorption occurs is represented as a function of PSP in the GLEAMS algorithms, with the same equation.
used for both adsorption and desorption. However, Barrow (1979) and others have shown that the desorption process does not retrace the path of the sorption process, particularly after long incubation periods. This indicates that the rate constants $k_{al}$ and $k_{al}$ in Equations 5-14 and 5-18 should have different values depending on the direction of the reaction. This is a method employed by the Soil and Water Assessment Tool (SWAT) model, which also uses P algorithms almost entirely from GLEAMS, where the transfer from active P to labile P and from stable P to active P is assumed to be one-tenth that of the reverse reaction (Neitsch et al. 2002). Additionally, PSP is determined as a function of the base saturation and pH of the soil for slightly weathered soils (such as the Spodosols and Alfisols found in the flatwoods). Vadas (2001) recently showed that for sandy soils in the Delaware coastal plain this P availability index is better represented as a function of oxalate-extractable aluminum and labile P concentrations. It is likely a similar relationship could be developed for Florida flatwoods soils.

The model showed poorer predictions from the summer pastures compared to the winter, and was overall a poor predictor for all pastures. The poorer predictions in the summer pastures is partially due to the increased effects of stocking on these pastures since the cattle density is larger (due to the smaller pasture size compared to the winter pastures) and the cattle are present for larger periods of each year. The poor predictions of the model are likely also due to the “one-size-fits-all” approach of the algorithms as adopted from GLEAMS. These algorithms use constant, hard-coded maximum rate constants and proportionality coefficients that are based on limited data. Due to the relative uncertainty in these parameters it may be more appropriate for these parameters to be inputted, calibratable values. As inputted values the model can more easily take
advantage of the best available data and the sensitivity of the model to these parameters can be evaluated.

The algorithms of the original, unmodified N and P module (Chapter 5) predicted consistently lower N and P loads and concentrations for all pastures and for all years of simulation (Tables 7-7 to 7-10). For the case of P this is partly due to the relatively large partitioning coefficients predicted by the model from the soil clay content (Equation 5-21). These partitioning coefficients have been noted to overpredict P adsorption in flatwoods soils in Florida (Fraisse and Campbell 1997). The relatively large partitioning coefficient results in a relatively low concentration of P in runoff water as determined by Equation 5-49 due to the application of the partitioning coefficient in Equation 5-46 and 5-50. For the case of both N and P the low predicted loads and concentrations are due to the cessation of transformations that could supply labile forms. The rate of transformation from active P (Equation 5-14), active N (Equation 5-30), animal waste P (Equation 5-7), and animal waste N (5-28) are all affected by a soil moisture response factor (Equation 5-40) that reduces the transformations to zero at water contents above field capacity. Thus upon saturation of the soil, and the removal of labile nutrients by a runoff event, no new labile forms are produced in the surface soil layer beyond that entering via rainfall, upward flow, and groundwater flow until soil moisture contents are reduced below field capacity.

The shortcomings in the prediction of N and P loads would likely be overcome by a more accurate representation of organic forms and the use of site-specific or region-specific relationships. The relationships used by the GLEAMS model were determined using a limited number of soils from throughout the continental U.S. and may not be
representative of conditions of the Florida flatwoods. The need for such specific relationships limits the predictive ability of the model for application to ungauged sites. For most practical applications, where site-specific soil data are unavailable, a simplified process model (as applied here and described in Chapter 6) that is more flexible in terms of rate constants and coefficients of proportionality may be more suitable. Such rate constants and coefficients could be inputs to the model that could be based on the best available data and/or calibratable parameters.

**Conclusions**

The ACRU2000 model, using nitrogen and phosphorus cycling algorithms that are patterned after those used in the GLEAMS model, was modified to better represent the hydrology and water quality of shallow water-table flatwoods regions of the Lake Okeechobee basin. Modifications were made to the response of N and P transformations to soil moisture, the estimation of linear partitioning coefficients that govern the adsorption of labile phosphorus, and to the of the extraction of N and P into runoff.

Validation results, while generally in agreement with observations in terms of the prediction of runoff, indicate several shortcomings in the representation of N and P processes by the model. The model proved incapable of reproducing large observed loads following a drought year, possibly due to the lack of an explicit representation of biomass production, senescence, and organic soil accretion and mineralization. The model produced increased loads from increased cattle stocking rates that were not observed. This may be due to the model’s representation of atmospheric losses of nitrogen and the rate that P transforms into more stable, recalcitrant forms. The model also produced generally worse simulations on one set of experimental pastures compared to the other. This was likely due to the more intensive simulated stocking on the pastures
as well as to the assumed reaction rates and assumed proportionality between nutrient pools by the model.

Compared to the original, unmodified algorithms of the ACRU2000 N and P module (Chapter 5), the model simulated loads and concentrations that were generally in better agreement with observations. These were in better agreement due to a more appropriate prediction of sorption of P in flatwoods soils, affect of soil moisture on transformation rates, and extraction of solutes into ponded/runoff water as proposed in Chapter 6.

Model sensitivity of N and P loads to hydrologic parameters indicated the importance of correct hydrologic prediction in order to accurately predict nutrient loads. The hydrologic parameters having the largest effects on loads were largely the same as those having the largest effect on runoff prediction (Chapter 4). Model sensitivity of N and P loads to nutrient parameters showed the model to be most sensitive to manure application rates, the depth of soil that interacts with runoff, concentrations of N in rainfall, total and active soil N contents, total and stable P contents, and the phosphorus availability index.

It is recommended that the model incorporate a more detailed plant/organic carbon representation in order to better capture the effects of plant residue production, organic soil formation and mineralization and associated accumulation/release of nutrients. It is also recommended that maximum reaction rates and constants that define the equilibrium between nutrient pools should be user-variable parameters where the most current knowledge can be used.
Table 7-1. Pineda fine sand soil physical properties

<table>
<thead>
<tr>
<th>Layer \ Layer Depth$^{[a]}$ (cm)</th>
<th>Bulk Density (g/cm$^3$)</th>
<th>Hydraulic Conductivity (cm/hr)</th>
<th>Organic Matter (%$^{[b]}$)</th>
<th>Silt (%)</th>
<th>Clay (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0-10</td>
<td>1.22, 1.20, 1.01$^{[b]}$</td>
<td>4.32</td>
<td>14.6, 16.4, 18.8$^{[b]}$</td>
<td>0.4</td>
</tr>
<tr>
<td>E</td>
<td>10-30</td>
<td>1.63</td>
<td>4.14</td>
<td>0.24</td>
<td>0.8</td>
</tr>
<tr>
<td>Bw</td>
<td>30-80</td>
<td>1.69</td>
<td>2.70</td>
<td>0.07</td>
<td>0.7</td>
</tr>
<tr>
<td>Btg</td>
<td>80-140</td>
<td>1.72</td>
<td>1.37</td>
<td>0.67</td>
<td>2.6</td>
</tr>
<tr>
<td>Cg</td>
<td>140-160</td>
<td>1.66</td>
<td>2.95</td>
<td>0.11</td>
<td>3.2</td>
</tr>
</tbody>
</table>

$^{[a]}$ From Gathumbi et al. (2005).
$^{[b]}$ From Capece et al. (2003). Values are for winter pastures 5, 6 and 7, respectively. All other values adapted from Sodek et al. (1990).

Table 7-2. Felda fine sand soil physical properties

<table>
<thead>
<tr>
<th>Layer \ Layer Depth$^{[a]}$ (cm)</th>
<th>Bulk Density (g/cm$^3$)</th>
<th>Hydraulic Conductivity (cm/hr)</th>
<th>Organic Matter (%$^{[b]}$)</th>
<th>Silt (%)</th>
<th>Clay (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0-18</td>
<td>1.13, 1.27, 1.25$^{[b]}$</td>
<td>25.35</td>
<td>14.6, 13.5, 15.8$^{[b]}$</td>
<td>1.3</td>
</tr>
<tr>
<td>E</td>
<td>18-60</td>
<td>1.59</td>
<td>8.76</td>
<td>0.12</td>
<td>2.4</td>
</tr>
<tr>
<td>Btg</td>
<td>60-100</td>
<td>1.52</td>
<td>0.76</td>
<td>0.08</td>
<td>5.9</td>
</tr>
<tr>
<td>Cg</td>
<td>100-170</td>
<td>1.66</td>
<td>4.31</td>
<td>0.05</td>
<td>3.6</td>
</tr>
</tbody>
</table>

$^{[a]}$ From Gathumbi et al. (2005).
$^{[b]}$ From Capece et al. (2003). Values are for summer pastures 1, 3, and 4, respectively. All other values adapted from Sodek et al. (1990).

Table 7-3. Pineda fine sand saturated water content ($\theta_s$), residual water content ($\theta_r$), and van Genuchten (1980) $\alpha$, $n$, and $m$ parameters

<table>
<thead>
<tr>
<th>Layer \ Layer Depth (cm)</th>
<th>$\theta_s$ (cm$^3$/cm$^3$)</th>
<th>$\theta_r$ (cm$^3$/cm$^3$)</th>
<th>$\alpha$ (cm$^{-1}$)</th>
<th>$n$ (-)</th>
<th>$m$ (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0-10</td>
<td>0.42</td>
<td>0.10</td>
<td>0.0287</td>
<td>1.96</td>
</tr>
<tr>
<td>E</td>
<td>10-30</td>
<td>0.34</td>
<td>0.08</td>
<td>0.0224</td>
<td>2.57</td>
</tr>
<tr>
<td>Bw</td>
<td>30-80</td>
<td>0.32</td>
<td>0.07</td>
<td>0.0234</td>
<td>1.81</td>
</tr>
<tr>
<td>Btg</td>
<td>80-140</td>
<td>0.35</td>
<td>0.15</td>
<td>0.0177</td>
<td>1.67</td>
</tr>
<tr>
<td>Cg</td>
<td>140-160</td>
<td>0.30</td>
<td>0.06</td>
<td>0.0106</td>
<td>2.06</td>
</tr>
</tbody>
</table>

Parameter values adapted from the data of Sodek et al. (1990).
Table 7-4. Felda fine sand saturated water content ($\theta_s$), residual water content ($\theta_r$), and van Genuchten (1980) $\alpha$, $n$, and $m$ parameters

<table>
<thead>
<tr>
<th>Layer</th>
<th>Depth (cm)</th>
<th>$\theta_s$ (cm$^3$/cm$^3$)</th>
<th>$\theta_r$ (cm$^3$/cm$^3$)</th>
<th>$\alpha$ (cm$^{-1}$)</th>
<th>$n$ (-)</th>
<th>$m$ (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0-18</td>
<td>0.45</td>
<td>0.14</td>
<td>0.0158</td>
<td>3.52</td>
<td>0.716</td>
</tr>
<tr>
<td>E</td>
<td>18-60</td>
<td>0.35</td>
<td>0.11</td>
<td>0.0150</td>
<td>3.13</td>
<td>0.681</td>
</tr>
<tr>
<td>Btg</td>
<td>60-100</td>
<td>0.42</td>
<td>0.18</td>
<td>0.0055</td>
<td>1.26</td>
<td>0.206</td>
</tr>
<tr>
<td>Cg</td>
<td>100-170</td>
<td>0.36</td>
<td>0.11</td>
<td>0.0114</td>
<td>3.38</td>
<td>0.704</td>
</tr>
</tbody>
</table>

Parameter values adapted from the data of Sodek et al. (1990).

Table 7-5. Pineda fine sand soil chemical properties

<table>
<thead>
<tr>
<th>Layer</th>
<th>Depth (cm)</th>
<th>pH$^a$</th>
<th>Base Saturation$^b$ (%)</th>
<th>Mg$_{DA}$ (mg/kg)</th>
<th>Al$_{OX}$ (mg/kg)</th>
<th>Total Nitrogen$^c$ (mg/kg)</th>
<th>Total Phosphorus$^a$ (mg/kg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0-10</td>
<td>4.3, 4.3, 4.1$^d$</td>
<td>41</td>
<td>99</td>
<td>201</td>
<td>500</td>
<td>125, 104, 119$^d$</td>
</tr>
<tr>
<td>E</td>
<td>10-30</td>
<td>4.5</td>
<td>39</td>
<td>16</td>
<td>158</td>
<td>100</td>
<td>5.6</td>
</tr>
<tr>
<td>Bw</td>
<td>30-80</td>
<td>5.0</td>
<td>22</td>
<td>17</td>
<td>1472</td>
<td>100</td>
<td>1.5</td>
</tr>
<tr>
<td>Btg</td>
<td>80-140</td>
<td>5.9</td>
<td>51</td>
<td>-</td>
<td>-</td>
<td>100</td>
<td>2.0</td>
</tr>
<tr>
<td>Cg</td>
<td>140-160</td>
<td>6.0</td>
<td>99</td>
<td>16</td>
<td>158</td>
<td>100</td>
<td>1.8</td>
</tr>
</tbody>
</table>

$^a$ Data for A Horizon from Capece et al. (2003), other horizons from Hill (2003).
$^b$ From Sodek et al. (1990).
$^c$ From Nair et al. (1995).
$^d$ Values are for winter pastures 5, 6 and 7, respectively.

Table 7-6. Felda fine sand soil chemical properties

<table>
<thead>
<tr>
<th>Layer</th>
<th>Depth (cm)</th>
<th>pH$^a$</th>
<th>Base Saturation$^b$ (%)</th>
<th>Mg$_{DA}$ (mg/kg)</th>
<th>Al$_{OX}$ (mg/kg)</th>
<th>Total Nitrogen$^c$ (mg/kg)</th>
<th>Total Phosphorus$^a$ (mg/kg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0-18</td>
<td>4.7, 4.9, 4.5$^d$</td>
<td>14</td>
<td>99</td>
<td>201</td>
<td>900</td>
<td>212, 130, 199$^d$</td>
</tr>
<tr>
<td>E</td>
<td>18-60</td>
<td>4.9</td>
<td>45</td>
<td>16</td>
<td>158</td>
<td>100</td>
<td>5.6</td>
</tr>
<tr>
<td>Btg</td>
<td>60-100</td>
<td>5.7</td>
<td>86</td>
<td>-</td>
<td>-</td>
<td>100</td>
<td>2.0</td>
</tr>
<tr>
<td>Cg</td>
<td>100-170</td>
<td>7.7</td>
<td>93</td>
<td>16</td>
<td>158</td>
<td>100</td>
<td>1.8</td>
</tr>
</tbody>
</table>

$^a$ Data for A Horizon from Capece et al. (2003), other horizons from Hill (2003).
$^b$ From Sodek et al. (1990).
$^c$ From Nair et al. (1995).
$^d$ Values are for summer pastures 1, 3 and 4, respectively.
Table 7-7. Annual observed and simulated runoff and N and P loads in runoff for winter pastures

<table>
<thead>
<tr>
<th>Pasture and Year</th>
<th>Runoff (mm)</th>
<th>N Load (kg/ha)</th>
<th>P Load (kg/ha)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Obs.</td>
<td>Sim. [a]</td>
<td>Sim. [b]</td>
</tr>
<tr>
<td>WP7</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1998</td>
<td>244.6</td>
<td>198.1</td>
<td>0.570</td>
</tr>
<tr>
<td>1999</td>
<td>149.3</td>
<td>148.8</td>
<td>0.335</td>
</tr>
<tr>
<td>2000</td>
<td>9.8</td>
<td>81.4</td>
<td>0.065</td>
</tr>
<tr>
<td>2001</td>
<td>367.1</td>
<td>373.7</td>
<td>0.929</td>
</tr>
<tr>
<td>2002</td>
<td>380.3</td>
<td>484.8</td>
<td>0.751</td>
</tr>
<tr>
<td>2003</td>
<td>459.2</td>
<td>371.6</td>
<td>0.709</td>
</tr>
<tr>
<td>Total:</td>
<td>1610.3</td>
<td>1658.4</td>
<td>3.359</td>
</tr>
<tr>
<td>WP6</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1998</td>
<td>214.1</td>
<td>145.8</td>
<td>0.346</td>
</tr>
<tr>
<td>1999</td>
<td>154.2</td>
<td>92.8</td>
<td>0.304</td>
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<tr>
<td>2000</td>
<td>3.3</td>
<td>49.3</td>
<td>0.081</td>
</tr>
<tr>
<td>2001</td>
<td>303.8</td>
<td>302.9</td>
<td>0.847</td>
</tr>
<tr>
<td>2002</td>
<td>265.8</td>
<td>372.0</td>
<td>0.544</td>
</tr>
<tr>
<td>2003</td>
<td>468.4</td>
<td>263.7</td>
<td>0.581</td>
</tr>
<tr>
<td>Total:</td>
<td>1409.6</td>
<td>1226.5</td>
<td>2.703</td>
</tr>
<tr>
<td>WP5</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1998</td>
<td>250.6</td>
<td>155.2</td>
<td>0.381</td>
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<tr>
<td>1999</td>
<td>135.8</td>
<td>108.1</td>
<td>0.364</td>
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<td>2000</td>
<td>17.7</td>
<td>53.5</td>
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<tr>
<td>2001</td>
<td>399.9</td>
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<td>0.978</td>
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<tr>
<td>2002</td>
<td>370.3</td>
<td>412.4</td>
<td>0.809</td>
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<tr>
<td>2003</td>
<td>473.1</td>
<td>307.2</td>
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</tr>
<tr>
<td>Total:</td>
<td>1647.4</td>
<td>1360.5</td>
<td>3.686</td>
</tr>
</tbody>
</table>

[a] Simulated with the modifications made for shallow water-table and flatwoods conditions as described in Chapter 6.
[b] Simulated using original, unmodified equations as adapted from GLEAMS and described in Chapter 5.
Table 7-8. Annual observed and simulated runoff and N and P loads in runoff for summer pastures

<table>
<thead>
<tr>
<th>Pasture and Year</th>
<th>Runoff (mm)</th>
<th>N Load (kg/ha)</th>
<th>P Load (kg/ha)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SP1</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>1998</td>
<td>94.3</td>
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<tr>
<td>2003</td>
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<td>159.8</td>
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<td>124.4</td>
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<tr>
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<td>247.5</td>
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<tr>
<td>2002</td>
<td>293.1</td>
<td>246.5</td>
<td>0.704</td>
</tr>
<tr>
<td>2003</td>
<td>201.4</td>
<td>215.1</td>
<td>0.204</td>
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<tr>
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<td>1046.0</td>
<td>2.499</td>
</tr>
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</tr>
<tr>
<td>1998</td>
<td>145.8</td>
<td>124.2</td>
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<tr>
<td>1999</td>
<td>96.4</td>
<td>149.7</td>
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<td>2000</td>
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<td>62.9</td>
<td>0.0</td>
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<td>247.4</td>
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<td>2002</td>
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<td>2003</td>
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<td>Total:</td>
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<td>1045.2</td>
<td>2.559</td>
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</table>

[a] Simulated with the modifications made for shallow water-table and flatwoods conditions as described in Chapter 6.
[b] Simulated using original, unmodified equations as adapted from GLEAMS and described in Chapter 5.
Table 7-9. Observed and simulated average N and P concentrations in runoff for winter pastures

<table>
<thead>
<tr>
<th>Pasture and Year</th>
<th>N Concentration (mg/L)</th>
<th>P Concentration (mg/L)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Obs.</td>
<td>Sim.[a]</td>
</tr>
<tr>
<td><strong>WP7</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1998</td>
<td>0.23</td>
<td>0.22</td>
</tr>
<tr>
<td>1999</td>
<td>0.22</td>
<td>0.23</td>
</tr>
<tr>
<td>2000</td>
<td>0.66</td>
<td>0.23</td>
</tr>
<tr>
<td>2001</td>
<td>0.25</td>
<td>0.21</td>
</tr>
<tr>
<td>2002</td>
<td>0.20</td>
<td>0.21</td>
</tr>
<tr>
<td>2003</td>
<td>0.15</td>
<td>0.23</td>
</tr>
<tr>
<td>Mean:</td>
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<td>0.22</td>
</tr>
<tr>
<td><strong>WP6</strong></td>
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<td></td>
</tr>
<tr>
<td>1998</td>
<td>0.16</td>
<td>0.22</td>
</tr>
<tr>
<td>1999</td>
<td>0.20</td>
<td>0.22</td>
</tr>
<tr>
<td>2000</td>
<td>2.43</td>
<td>0.21</td>
</tr>
<tr>
<td>2001</td>
<td>0.28</td>
<td>0.21</td>
</tr>
<tr>
<td>2002</td>
<td>0.20</td>
<td>0.22</td>
</tr>
<tr>
<td>2003</td>
<td>0.12</td>
<td>0.24</td>
</tr>
<tr>
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<td>0.22</td>
</tr>
<tr>
<td><strong>WP5</strong></td>
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<td></td>
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<tr>
<td>1998</td>
<td>0.15</td>
<td>0.26</td>
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<tr>
<td>1999</td>
<td>0.27</td>
<td>0.23</td>
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<td>2000</td>
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<td>0.21</td>
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<td>2001</td>
<td>0.24</td>
<td>0.22</td>
</tr>
<tr>
<td>2002</td>
<td>0.22</td>
<td>0.26</td>
</tr>
<tr>
<td>2003</td>
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<td>0.28</td>
</tr>
<tr>
<td>Mean:</td>
<td>0.48</td>
<td>0.25</td>
</tr>
</tbody>
</table>

[a] Simulated with the modifications made for shallow water-table and flatwoods conditions as described in Chapter 6.
[b] Simulated using original, unmodified equations as adapted from GLEAMS and described in Chapter 5.
Table 7-10. Observed and simulated average N and P concentrations in runoff for summer pastures

<table>
<thead>
<tr>
<th>Pasture and Year</th>
<th>N Concentration (mg/L)</th>
<th>P Concentration (mg/L)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Obs.</td>
<td>Sim.[^{[a]}]</td>
</tr>
<tr>
<td>SP1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1998</td>
<td>0.12</td>
<td>0.16</td>
</tr>
<tr>
<td>1999</td>
<td>0.38</td>
<td>0.15</td>
</tr>
<tr>
<td>2000</td>
<td>0.0</td>
<td>0.16</td>
</tr>
<tr>
<td>2001</td>
<td>1.14</td>
<td>0.16</td>
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<tr>
<td>2002</td>
<td>0.22</td>
<td>0.16</td>
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<tr>
<td>2003</td>
<td>0.17</td>
<td>0.16</td>
</tr>
<tr>
<td>Mean:</td>
<td>0.34</td>
<td>0.16</td>
</tr>
<tr>
<td>SP4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1998</td>
<td>0.19</td>
<td>0.20</td>
</tr>
<tr>
<td>1999</td>
<td>0.25</td>
<td>0.41</td>
</tr>
<tr>
<td>2000</td>
<td>0.64</td>
<td>0.33</td>
</tr>
<tr>
<td>2001</td>
<td>0.35</td>
<td>0.41</td>
</tr>
<tr>
<td>2002</td>
<td>0.24</td>
<td>0.38</td>
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<td>2003</td>
<td>0.10</td>
<td>0.47</td>
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<td>Mean:</td>
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<td>0.37</td>
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<tr>
<td>1998</td>
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<td>1999</td>
<td>0.48</td>
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<td>2001</td>
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<tr>
<td>2002</td>
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<tr>
<td>2003</td>
<td>0.09</td>
<td>0.80</td>
</tr>
<tr>
<td>Mean:</td>
<td>0.20</td>
<td>0.62</td>
</tr>
</tbody>
</table>

\[^{[a]}\] Simulated with the modifications made for shallow water-table and flatwoods conditions as described in Chapter 6.
\[^{[b]}\] Simulated using original, unmodified equations as adapted from GLEAMS and described in Chapter 5.
Table 7-11. Mean absolute error (MAE), root mean square error (RMSE) and coefficient of efficiency (E) for annual runoff, N and P loads, and average N and P concentrations for all pastures

<table>
<thead>
<tr>
<th>Pasture and Parameter</th>
<th>MAE</th>
<th>RMSE</th>
<th>E (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>WP7</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Runoff (mm)</td>
<td>52.9</td>
<td>65.7</td>
<td>0.966</td>
</tr>
<tr>
<td>N Load (kg/ha)</td>
<td>0.138</td>
<td>0.159</td>
<td>0.905</td>
</tr>
<tr>
<td>P Load (kg/ha)</td>
<td>0.182</td>
<td>0.277</td>
<td>0.235</td>
</tr>
<tr>
<td>N Conc. (mg/L)</td>
<td>0.099</td>
<td>0.183</td>
<td>-0.288</td>
</tr>
<tr>
<td>P Conc. (mg/L)</td>
<td>0.111</td>
<td>0.149</td>
<td>-1.054</td>
</tr>
<tr>
<td><strong>WP6</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Runoff (mm)</td>
<td>81.2</td>
<td>103.1</td>
<td>0.722</td>
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<tr>
<td>N Load (kg/ha)</td>
<td>0.113</td>
<td>0.148</td>
<td>0.860</td>
</tr>
<tr>
<td>P Load (kg/ha)</td>
<td>0.127</td>
<td>0.202</td>
<td>0.326</td>
</tr>
<tr>
<td>N Conc. (mg/L)</td>
<td>0.416</td>
<td>0.906</td>
<td>-0.404</td>
</tr>
<tr>
<td>P Conc. (mg/L)</td>
<td>0.426</td>
<td>0.888</td>
<td>-0.490</td>
</tr>
<tr>
<td><strong>WP5</strong></td>
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<td></td>
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</tr>
<tr>
<td>Runoff (mm)</td>
<td>73.8</td>
<td>87.7</td>
<td>0.905</td>
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<tr>
<td>N Load (kg/ha)</td>
<td>0.146</td>
<td>0.175</td>
<td>0.813</td>
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<tr>
<td>P Load (kg/ha)</td>
<td>0.164</td>
<td>0.200</td>
<td>0.857</td>
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<tr>
<td>N Conc. (mg/L)</td>
<td>0.316</td>
<td>0.651</td>
<td>-0.745</td>
</tr>
<tr>
<td>P Conc. (mg/L)</td>
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<td>0.100</td>
<td>-0.899</td>
</tr>
<tr>
<td><strong>SP1</strong></td>
<td></td>
<td></td>
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<tr>
<td>Runoff (mm)</td>
<td>48.4</td>
<td>59.9</td>
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<td>N Load (kg/ha)</td>
<td>0.596</td>
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<tr>
<td>N Conc. (mg/L)</td>
<td>0.245</td>
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<td>P Conc. (mg/L)</td>
<td>0.415</td>
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<td>-0.995</td>
</tr>
<tr>
<td><strong>SP4</strong></td>
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<td></td>
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<tr>
<td>Runoff (mm)</td>
<td>28.2</td>
<td>33.4</td>
<td>0.987</td>
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<tr>
<td>N Load (kg/ha)</td>
<td>0.259</td>
<td>0.367</td>
<td>-0.710</td>
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<tr>
<td>P Load (kg/ha)</td>
<td>0.978</td>
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<td>0.649</td>
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<tr>
<td>N Conc. (mg/L)</td>
<td>0.176</td>
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<td>P Conc. (mg/L)</td>
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<tr>
<td>Runoff (mm)</td>
<td>68.3</td>
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<td>P Load (kg/ha)</td>
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<td>P Conc. (mg/L)</td>
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<td>0.675</td>
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Table 7-12. Mean absolute error (MAE), root mean square error (RMSE) and coefficient of efficiency (E) for annual N and P loads and average N and P concentrations for all pastures using original, unmodified ACRU2000 N and P algorithms

<table>
<thead>
<tr>
<th>Pasture and Parameter</th>
<th>MAE</th>
<th>RMSE</th>
<th>E (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>WP7</strong></td>
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<td></td>
</tr>
<tr>
<td>N Load (kg/ha)</td>
<td>0.472</td>
<td>0.534</td>
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<td>0.208</td>
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</tr>
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<td>0.148</td>
<td>0.200</td>
<td>-4.28</td>
</tr>
<tr>
<td><strong>WP6</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>N Load (kg/ha)</td>
<td>0.347</td>
<td>0.387</td>
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<td>P Load (kg/ha)</td>
<td>0.221</td>
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<td>-9.39</td>
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<td>0.608</td>
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<td>-1.09</td>
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<tr>
<td><strong>WP5</strong></td>
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<tr>
<td>N Load (kg/ha)</td>
<td>0.443</td>
<td>0.466</td>
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<tr>
<td>P Load (kg/ha)</td>
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<td>N Conc. (mg/L)</td>
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<td>P Conc. (mg/L)</td>
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<tr>
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<td></td>
<td></td>
</tr>
<tr>
<td>N Load (kg/ha)</td>
<td>0.273</td>
<td>0.364</td>
<td>-4.56</td>
</tr>
<tr>
<td>P Load (kg/ha)</td>
<td>1.04</td>
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<td>-3.65</td>
</tr>
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<td>N Conc. (mg/L)</td>
<td>0.118</td>
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<td>0.587</td>
<td>0.661</td>
<td>-18.6</td>
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<tr>
<td><strong>SP4</strong></td>
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<tr>
<td>N Load (kg/ha)</td>
<td>0.164</td>
<td>0.193</td>
<td>0.499</td>
</tr>
<tr>
<td>P Load (kg/ha)</td>
<td>1.18</td>
<td>1.59</td>
<td>-3.92</td>
</tr>
<tr>
<td>N Conc. (mg/L)</td>
<td>0.190</td>
<td>0.254</td>
<td>-2.54</td>
</tr>
<tr>
<td>P Conc. (mg/L)</td>
<td>0.605</td>
<td>0.677</td>
<td>-24.7</td>
</tr>
<tr>
<td><strong>SP3</strong></td>
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<tr>
<td>N Load (kg/ha)</td>
<td>0.326</td>
<td>0.446</td>
<td>-7.80</td>
</tr>
<tr>
<td>P Load (kg/ha)</td>
<td>0.717</td>
<td>1.02</td>
<td>-2.84</td>
</tr>
<tr>
<td>N Conc. (mg/L)</td>
<td>0.166</td>
<td>0.217</td>
<td>-1.67</td>
</tr>
<tr>
<td>P Conc. (mg/L)</td>
<td>0.351</td>
<td>0.403</td>
<td>-15.0</td>
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Table 7-13. Hydrologic input parameters included in the sensitivity analysis

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<th>Parameter</th>
<th>Unit</th>
<th>Base Value</th>
<th>Description</th>
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<td>$K_c$</td>
<td>–</td>
<td>0.875</td>
<td>Crop coefficient</td>
</tr>
<tr>
<td>L</td>
<td>m</td>
<td>0.8</td>
<td>Maximum depth of roots</td>
</tr>
<tr>
<td>C</td>
<td>–</td>
<td>-1</td>
<td>Root distribution shape parameter</td>
</tr>
<tr>
<td>$h_2$</td>
<td>cm</td>
<td>20</td>
<td>Transpiration reduction due to $O_2$ deficiency</td>
</tr>
<tr>
<td>$h_3$</td>
<td>cm</td>
<td>10000</td>
<td>Transpiration reduction due to water deficiency</td>
</tr>
<tr>
<td>$z_{Dep}$</td>
<td>mm</td>
<td>10</td>
<td>Depression storage</td>
</tr>
<tr>
<td>I</td>
<td>mm</td>
<td>1</td>
<td>Interception capacity</td>
</tr>
<tr>
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<td>1/day</td>
<td>150</td>
<td>Runoff resistance</td>
</tr>
<tr>
<td>$\beta$</td>
<td>–</td>
<td>1.67</td>
<td>Runoff exponent</td>
</tr>
<tr>
<td>$d_{w1}$</td>
<td>m</td>
<td>1.9</td>
<td>Initial depth to water-table</td>
</tr>
</tbody>
</table>

$\theta_s$: Saturated water content

| $A$ | cm$^3$/cm$^3$ | 0.45     |
| $E$ | cm$^3$/cm$^3$ | 0.35     |
| $Btg$ | cm$^3$/cm$^3$ | 0.42     |
| $Cg$ | cm$^3$/cm$^3$ | 0.36     |

$\alpha$: Soil moisture shape parameter of van Genuchten (1980)

| $A$  | 1/cm | 0.0158 |
| $E$  | 1/cm | 0.0150 |
| $Btg$  | 1/cm | 0.0055 |
| $Cg$ | 1/cm | 0.0114 |

$n$: Soil moisture shape parameter of van Genuchten (1980)

| $A$ | – | 3.52 |
| $E$ | – | 3.13 |
| $Btg$ | – | 1.26 |
| $Cg$ | – | 3.38 |

$K_{s,H}$: Horizontal saturated hydraulic conductivity

| $A$ | cm/h | 25.35 |
| $E$ | cm/h | 8.76  |
| $Btg$ | cm/h | 0.76  |
| $Cg$ | cm/h | 4.31  |

$K_{s,V}$: Vertical saturated hydraulic conductivity used in the upward flux relationship of Anat et al. (1965)

| $A$  | cm/h | 13.57 |
| $E$  | cm/h | 11.7  |
| $Btg$  | cm/h | 5.58  |

$h_b$: Bubbling pressure head used in the upward flux relationship of Anat et al. (1965)

| $A$ | cm | 46.8 |
| $E$ | cm | 40.1 |
| $Btg$ | cm | 23.0 |

$\eta$: Exponent used in the upward flux relationship of Anat et al. (1965)

| $A$  | – | 12.2 |
| $E$  | – | 8.80 |
| $Btg$ | – | 4.98 |
Table 7-14. Nutrient input parameters included in the sensitivity analysis

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<tr>
<th>Parameter</th>
<th>Unit</th>
<th>Base Value</th>
<th>Description</th>
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<td>BSAT</td>
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<tr>
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<td>pH</td>
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<td>4.52</td>
<td>pH</td>
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<tr>
<td>App</td>
<td>kg/ha/day</td>
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<td>EDI</td>
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<td>Soil depth of interaction with runoff</td>
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<td>TN</td>
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<td>875.0</td>
<td>Total nitrogen</td>
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<tr>
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<td>736.3</td>
<td>Stable organic nitrogen</td>
</tr>
<tr>
<td>Nₐ</td>
<td>kg/ha</td>
<td>131.3</td>
<td>Active organic nitrogen</td>
</tr>
<tr>
<td>NₐNH₄</td>
<td>kg/ha</td>
<td>2.5</td>
<td>Initial ammonium content</td>
</tr>
<tr>
<td>NₐNO₃</td>
<td>kg/ha</td>
<td>12.5</td>
<td>Initial nitrate content</td>
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<td>βNO₃</td>
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<td>Nitrate extraction coefficient</td>
</tr>
<tr>
<td>Kd,NH₄</td>
<td>L/kg</td>
<td></td>
<td>Ammonium partitioning coefficient</td>
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<tr>
<td>A</td>
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Table 7-15. Sensitivity of phosphorus and nitrogen loads to hydrologic parameters (reported as percent difference of base simulation result). Values in left-hand column are percent changes in input parameter values.

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<th>β</th>
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<th>K_s,H</th>
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<th>h_s</th>
<th>n</th>
<th>d_WT</th>
<th>η</th>
<th>I</th>
<th>K_s,V</th>
<th>L</th>
<th>c</th>
<th>h_3</th>
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Table 7-16. Sensitivity of phosphorus and nitrogen to soil parameters and manure application rate (reported as percent difference of base simulation result)

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<th>EDI</th>
<th>pH</th>
<th>BSAT</th>
<th>CL</th>
<th>OM</th>
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Table 7-17. Sensitivity of nitrogen loads to nitrogen parameters (reported as percent difference of base simulation result)

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<th>K_{d,NH4}</th>
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<th>N_{NH4}</th>
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<th>N_{\text{stream}}</th>
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Table 7-18. Sensitivity of phosphorus loads to phosphorus parameters (reported as percent difference of base simulation result)

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<th>PSP</th>
<th>Ps</th>
<th>Pa</th>
<th>Pl</th>
<th>P_{rain}</th>
<th>K_{d,P}</th>
<th>β_P</th>
<th>P_{stream}</th>
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<tr>
<td>+50%</td>
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<td>21</td>
<td>14</td>
<td>4</td>
<td>5</td>
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<td>30</td>
<td>21</td>
<td>6</td>
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Figure 7-1. Location of MacArthur Agro-Ecology Research Center (MAERC) at Buck Island Ranch

Figure 7-2. Semi-improved winter pasture array at MacArthur Agro-Ecology Research Center at Buck Island Ranch. Showing location of weather stations, groundwater wells, and flumes. Also shown are pasture ID and cow-calf pairs stocked in each pasture.
Figure 7-3. Improved summer pasture array at MacArthur Agro-Ecology Research Center at Buck Island Ranch. Showing location of weather stations, groundwater wells, and flumes. Also shown are pasture ID and cow-calf pairs stocked in each pasture.
Figure 7-4. Groundwater levels at the 4-inch well (near the flume) and the 2-inch well (center of pasture) in winter pasture 6 (WP6). Ground surface elevations at the 4-inch and 2-inch wells are 8.69 and 8.66 meters above sea level, respectively.

Figure 7-5. Groundwater levels at the 4-inch well (near the flume) and the 2-inch well (center of pasture) in summer pasture 1 (SP1). Ground surface elevations at the 4-inch and 2-inch wells are 8.14 and 8.42 meters above sea level, respectively.
Figure 7-6. Groundwater levels from the three winter pastures compared to the canal stage as measured at the S70 spillway. Groundwater levels are from the 4-inch well in Winter Pasture 6 (WP6) and the 2-inch well in Winter Pasture 5 (WP5) and Winter Pasture 7 (WP7).

Figure 7-7. Groundwater levels from the three summer pastures compared to the canal stage as measured at the S70 spillway. Groundwater levels are from the 4-inch well in Summer Pasture 1 (SP1) and the 2-inch well in Summer Pasture 3 (SP3) and Summer Pasture 4 (SP4).
Figure 7-8. Winter Pasture 6 (WP6) observed and simulated depth to water-table

Figure 7-9. Winter Pasture 7 (WP7) observed and simulated depth to water-table
Figure 7-10. Winter Pasture 5 (WP5) observed and simulated depth to water-table

Figure 7-11. Summer Pasture 1 (SP1) observed and simulated depth to water-table
Figure 7-12. Summer Pasture 4 (SP4) observed and simulated depth to water-table

Figure 7-13. Summer Pasture 3 (SP3) observed and simulated depth to water-table
Figure 7-14. Winter Pasture 6 (WP6) observed and simulated daily runoff
Figure 7-15. Winter Pasture 7 (WP7) observed and simulated daily runoff
Figure 7-16. Winter Pasture 5 (WP5) observed and simulated daily runoff
Figure 7-17. Summer Pasture 1 (SP1) observed and simulated daily runoff
Figure 7-18. Summer Pasture 4 (SP4) observed and simulated daily runoff
Figure 7-19. Summer Pasture 3 (SP3) observed and simulated daily runoff
Figure 7-20. Winter Pasture 6 (WP6) cumulative annual runoff

Figure 7-21. Winter Pasture 7 (WP7) cumulative annual runoff
Figure 7-22. Winter Pasture 5 (WP5) cumulative annual runoff

Figure 7-23. Summer Pasture 1 (SP1) cumulative annual runoff
Figure 7-24. Summer Pasture 4 (SP4) cumulative annual runoff

Figure 7-25. Summer Pasture 3 (SP3) cumulative annual runoff
Figure 7-26. Winter Pasture 7 (WP7) cumulative annual N load

Figure 7-27. Winter Pasture 7 (WP7) cumulative annual P load
Figure 7-28. Winter Pasture 6 (WP6) cumulative annual N load

Figure 7-29. Winter Pasture 6 (WP6) cumulative annual P load
Figure 7-30. Winter Pasture 5 (WP5) cumulative annual N load

Figure 7-31. Winter Pasture 5 (WP5) cumulative annual P load
Figure 7-32. Summer Pasture 1 (SP1) cumulative annual N load

Figure 7-33. Summer Pasture 1 (SP1) cumulative annual P load
Figure 7-34. Summer Pasture 4 (SP4) cumulative annual N load

Figure 7-35. Summer Pasture 4 (SP4) cumulative annual P load
Figure 7-36. Summer Pasture 3 (SP3) cumulative annual N load

Figure 7-37. Summer Pasture 3 (SP3) cumulative annual P load
Figure 7-38. Hydrologic parameters showing the greatest sensitivity on N loads in runoff.

Figure 7-39. Hydrologic parameters showing the greatest sensitivity on P loads in runoff.
Figure 7-40. Nutrient parameters showing the greatest sensitivity on N loads in runoff

Figure 7-41. Nutrient parameters showing the greatest sensitivity on P loads in runoff
CHAPTER 8
SUMMARY AND CONCLUSIONS

In this study an object-oriented hydrologic and water quality module within the framework of the ACRU2000 model was proposed, developed, and tested for high water-table regions such as the flatwoods of Florida. The hydrologic model uses approximate methods appropriate for sandy, yet poorly drained soils. The water quality component of the model, a prior model addition of the nitrogen and phosphorus algorithms of the Groundwater Loading Effects of Agricultural Management Systems model (Campbell et al. 2001), was adapted for use in poorly drained conditions.

The proposed approximate, daily time-step hydrologic module of the model was designed by

- assuming a hydrostatic moisture distribution within the soil profile using three different soil moisture retention models,
- approximating the upward flow of water in response to evapotranspiration as a steady-state process,
- adding the standardized Penman-Moneth equation of the Food and Agricultural Organization (Allen et al. 1998) to estimate reference potential evapotranspiration,
- estimating incoming solar radiation using extraterrestrial radiation and the difference between maximum and minimum air temperatures when observations are absent,
- representing plant water stress as a function of soil water pressure head,
- including a closed-form root distribution function,
- including the interaction of groundwater with adjacent water bodies or boundary conditions, and
- assuming that runoff occurs via saturation-excess only.
The accuracy of the proposed approximate hydrologic model was demonstrated by applying the model to three sets of experimental observations. The model was shown to adequately predict water-table depths, soil moisture contents, evapotranspiration, and runoff volumes with minimal field-collected data and when applied at a relatively coarse resolution. The model compared well to a more detailed, one-dimensional, finite-difference model in predicting water table depths, soil moisture contents, and daily evapotranspiration from a wet prairie in north-central Florida. The model was also compared to the Field Hydrologic And Nutrient Transport Model (FHANTM) and provided similar results. The original, unmodified field-scale hydrologic module of the ACRU2000 model was shown to greatly overpredict runoff unless parameterized with unrealistic values. Given the approximations made by the model, the model is only suitable for use in shallow water-table regions with highly permeable soils where runoff can be assumed to occur primarily by saturation-excess.

The nitrogen and phosphorus module of the model was modified for shallow water-table environments with flatwoods soils by

- Relating the prediction of phosphorus partitioning coefficients to aluminum, magnesium, organic carbon, and clay contents in the soil,

- defining the maximum rate of reaction of mineralization, immobilization, and nitrification processes as affected by soil moisture to occur over a range of moisture contents,

- allowing mineralization and immobilization processes to occur under relatively wet conditions, but at a depressed rate, and

- representing the movement of mobile soil nutrients with runoff water as a partial mixing process, accounting for concentration gradients, within a defined depth of soil.

The ability of the model to predict N and P loads from flatwoods fields was evaluated by applying the model to six experimental cattle pastures, two control (no
cattle), two low-stocked pastures, and two high-stocked pastures. One set of differently-stocked pastures was from a location with no fertilization history and the other set from a location with a history of fertilization which is reflected by relatively higher soil P contents. Applying the model with minimal field-specific data and with minimal calibration indicated that the model was overall a relatively poor predictor of N and P loads from these sites and provided insight on potential improvements to the model’s performance. However the model was in better agreement with observations compared to the original, unmodified model.

**Future research needs and recommendations.** The evaluation of the model proposed here indicated the need for site-specific data in order to accurately predict both hydrologic and water quality outputs from ungauged sites. Evaluation of the model also showed possible inadequacies in the representation of nutrient cycling processes and potential inflexibilities in the model algorithms.

Soil hydraulic parameters were shown to have a large effect on the prediction of runoff and nutrient loads in the sensitivity analyses performed. These results indicated that, while valuable, sources of soil hydraulic information such as Sodek et al. (1990) may only provide a first approximation of the soil properties found at a particular site. In future model development for the Florida flatwoods it is recommended that greater resources be devoted to collecting these parameters.

In lieu of detailed site-specific soil data, and recognizing the difficulty in defining effective parameters at the scale of model application compared to the scale of measurement methods, it is recommended that the number of soil hydraulic parameters required by the model be reduced. In such an approximate, daily time-step model, soils
could likely be just as well represented by a single set of soil moisture parameters. Horizontal hydraulic conductivities could likely be just as well represented using a single average value or as linearly or exponentially decreasing with depth. Upward flux relationships could be simplified to a single empirical, calibratable exponential relationship. These simplifications would be commensurate with the level of approximations made in the model, particularly when these parameters are to be calibrated.

Predictions of nitrogen and phosphorus loads in runoff indicate the need for more detailed, site-specific soil information. The site-specific needs include data of soil nitrogen, and the properties considered to have the greatest effect on P retention, namely soil clay, Al, and Fe contents. The model’s performance also indicated the need for a better representation of organic nutrient forms that should include organic soil accretion and mineralization, and associated nutrient retention and release, in response to hydrologic and environmental conditions. The model used hard-coded, constant optimal rates of reaction and coefficients of proportionality that determine the equilibrium states between nutrient pools. Such constants should be user-defined values where the most up to date and/or site- or region-specific data can be used. In the absence of such data these values should be calibratable in order to provide a best-fit to observations.
APPENDIX A
HYDROLOGIC PROCESS AND DATA OBJECTS AND DESCRIPTIONS

Process Objects

- PAcruHWTRitchieEvapoTranspiration
- PDeepSeepage
- PFAO56PenmanMonteithDailyEvap
- PFindNewWaterTableDepth
- PHWTCropCoefTrans
- PHWTPlantWaterStress
- PHWTRitchieSoilWaterEvap
- PHWTSimpleEvapoTranspiration
- PInitialiseSoilUFOptionHWT
- PMaximumUpwardFlux
- PPondedWaterEvaporation
- PRootDistributionFunction
- PSimpleRunoff
- PSoilStorageAvailable
- PSoilWaterCharacteristic
- PStorageLimitedInfiltration
- PStorageLimitedRedistribution
- PSuperSimpleEvapoTranspiration
- PUpwardFlux

Data Objects

- DActualUpwardFlux
- DActualVaporPressureOption
- DAirVolume
- DDeepSeepage
- DDeepSeepageHeadBoundary
- DDeepSeepageOption
- DDewPointTemperature
- DDrainedtoEquilMoistureContent
- DINstrumentPsychrometricCoefficient
- DMaximumUpwardFlux
- DMaxRelativeHumidity
- DMinRelativeHumidity
- DNetRainfall
• DNumberOfSoilHorizons
• DOldAirVolume
• DPoreSizeDistributionIndex
• DPressureHead1
• DPressureHead2
• DPressureHead3
• DPressureHead4
• DRDFRootDepth
• DRDFCoefficient
• DResidualMoistureContent
• DRestrictiveLayerHydraulicConductivity
• DRestrictiveLayerThickness
• DRootZoneDeficit
• DSoilBubblingPressureHead
• DSoilLayerLocation
• DSoilStorageAvailable
• DSoilWaterCharacteristicOption
• DSoilWaterEvapDepth
• DSWAPRunoffExponent
• DSWAPRunoffResistance
• DUFluxBubblingPressureHead
• DUFluxExponent
• DUFluxSaturatedHydraulicConductivity
• DUHighWaterTableOption
• DUUpwardFluxDistance
• DWaterTableDepth
• DWaterTableDepthvsAirVolume
• DWetBulbTemperature

Description of Process Objects

• PAcruHWTRitchieEvapoTranspiration. This process calls the PHWTRitchieSoilWaterEvap and PHWTCropCoefTrans which determine soil water evaporation and transpiration, respectively using the method of Ritchie (1972).

• PDeepSeepage. This process determines and transfers deep seepage between the soil profile and the groundwater store underneath a restrictive layer.

• PFAO56PenmanMonteithDailyEvap. This process calculates reference potential transpiration according to the FAO Penman–Monteith equation described in Irrigation and Drainage paper # 56. Additional features include multiple methods of determining actual vapor pressure and a new method for estimating incoming solar radiation as outlined by Hargreaves and Samani (1982) and Samani (2000).
• **PFindNewWaterTableDepth.** This process determines a new water table depth and new drained to equilibrium water contents for the new water table depth.

• **PHWTCropCoefTrans.** This process determines transpiration when Ritchie’s method is chosen for simulating ET.

• **PHWTPlantWaterStress.** This process determines a plant water stress factor that causes a decrease in transpiration or evapotranspiration below the maximum rate (EVTR = 1 or 2) using the reduction relationship of Feddes et al. (1978). This is not used when EVTR = 3.

• **PHWTRitchieSoilWaterEvap.** This process determines soil water evaporation when Ritchie’s method is used.

• **PHWTSimpleEvapoTranspiration.** This process determines evapotranspiration together, and accounts for plant water stress.

• **PInitialiseSoilUFOptionHWT.** This process initializes several properties of the soil. A data object is set for each soil layer that contains the soil layer’s distance from the ground surface, called DSoilLayerLocation, so it does not have to be calculated each time it is required. This process also initializes the relationship between water table depth and air volume (or void volume) of the soil, assuming hydrostatic conditions. This relationship, in the form of an array data object called DWaterTablevsAirVolume, is used later to calculate changes in the water table depth. This process also sets the initial water content of all of the soil layers for a given water table depth assuming hydrostatic conditions.

• **PMaximumUpwardFlux.** This process determines the maximum possible steady-state upward flux from the water table for a given water table depth.

• **PPondedWaterEvaporation.** This process determines evaporation from water ponded on the land surface.

• **PRootDistributionFunction.** This process determines the fraction of plant roots in each soil layer using the root density function proposed by Hoogland et al. (1981). This process is used when UFHWT is on and EVTR = 1 or 2. When this process is used the fraction of plant roots in each soil horizon does not have to be entered manually.

• **PSimpleRunoff.** This process determines and transfers runoff from a landsegment using a simple equation as is done in the SWAP and FHANTM models. This process is only used when the model is run in lumped mode (only one landsegment).

• **PSoilStorageAvailable.** This process determines the storage available, or air volume, for the entire soil profile. This process determines the actual air volume,
which includes any depleted root zone, not the idealized air volume assumed when imposing hydrostatic conditions.

- **PSoilWaterCharacteristic.** This process calculates the soil moisture of a soil layer for a given distance to the water table. This process determines the water content or drained to equilibrium moisture content. This process does not set the value of storage to the soil layer, it just determines what the upper limit is.

- **PStorageLimitedInfiltration.** This process determines infiltration for a storage-only limited soil profile. Infiltration is considered to never be rate-limited.

- **PStorageLimitedRedistribution.** This process redistributes water vertically to account for the changes in water contents and water table depths that have occurred on a given day.

- **PSuperSimpleEvapoTranspiration.** This process determines evapotranspiration in a very simple manner. ET proceeds at a rate equal to the maximum until the wilting point is reached. Water is removed from the root zone soil layers from top down. No plant water stress is accounted for.

- **PUpwardFlux.** This process determines and transfers the amount of upward flux from the water table to depleted soil layers.

**Description of Data Objects**

- **DActualUpwardFlux.** This DDailyData data object holds the value of the actual upward flux that occurred for a given day.

- **DActualVaporPressureOption.** This DInteger holds the user’s option for determining the actual vapor pressure in the FAO Penman-Monteith equation in PFAO56PenmanMonteithDailyEvap.

- **DAirVolume.** This DDouble holds the current value of the air volume, or void volume, of the soil. Every process that will raise or drop the water table depth adjusts this value accordingly. At the end of the day this value determines the new water table depth.

- **DDeepSeepage.** This DDailyData data object holds the amount of deep seepage that has occurred on a given day.

- **DDeepSeepageHeadBoundary.** This DDailyData data object holds the head of the groundwater below the restrictive layer.

- **DDeepSeepageOption.** This DInteger data object determines whether deep seepage is simulated. If it is not then the restrictive layer is considered to be completely impermeable.
• **DDewPointTemperature.** This DDailyData data object holds the dew point temperature for a given day.

• **DDrainedtoEquilMoistureContent.** This DDouble data object holds the upper limit water content in a given soil layer. This value is a function of only the water table depth since hydrostatic conditions are assumed.

• **DInstrumentPsychrometricCoefficient.** This DDouble data object holds the psychometric constant of the instrument (psychronometer) that records the wet bulb temperature.

• **DMaximumUpwardFlux.** This data object holds the maximum upward flux that is calculated for a given water table depth.

• **DMaxRelativeHumidity.** This DDailyData data object holds the maximum relative humidity for a given day.

• **DMinRelativeHumidity.** This DDailyData data object holds the minimum relative humidity for a given day.

• **DNetRainfall.** This DDailyData object holds the amount of rainfall minus loss to interception and belongs to the climate.

• **DNumberOfSoilHorizons.** This DInteger data object holds the number of soil horizons that are to be simulated.

• **DOldAirVolume.** This DDouble data object holds the air volume of the previous day.

• **DPoreSizeDistributionIndex.** This DDouble data object holds the pore size distribution parameter, or its equivalent, that is used in defining the soil water characteristic.

• **DPressureHead1.** This DDouble data object holds the pressure head (cm) below which transpiration (EVTR = 2) or lumped evapotranspiration (EVTR = 1) is reduced to zero due to water excess.

• **DPressureHead2.** This DDouble data object holds the pressure head (cm) below which transpiration (EVTR = 2) or lumped evapotranspiration (EVTR = 1) begins to be reduced below the maximum rate due to water excess.

• **DPressureHead3.** This DDouble data object holds the pressure head (cm) above which transpiration (EVTR = 2) or lumped evapotranspiration (EVTR = 1) begins to be reduced below the maximum rate due to water limitation.

• **DPressureHead4.** This DDouble data object holds the pressure head (cm) above which transpiration (EVTR = 2) or lumped evapotranspiration (EVTR = 1) is
reduced to zero due to water limitation. This is the pressure head at the wilting point.

- **DRDFRootDepth.** This DDailyData object holds the root depth that is used in the root distribution function proposed by Hoogland et al. (1981).

- **DRDFCoefficient.** This DDailyData object holds the shape parameter that is used in the root distribution function proposed by Hoogland et al. (1981).

- **DResidualMoistureContent.** This DDouble data object holds the residual moisture content that is used in defining the soil water characteristic.

- **DRestrictiveLayerHydraulicConductivity.** This DDouble data object holds the vertical saturated hydraulic conductivity of the restrictive layer below the soil.

- **DRestrictiveLayerThickness.** This DDouble data object holds the thickness of the restrictive layer below the soil.

- **DRootZoneDeficit.** This DDouble data object holds the amount that the root zone of the soil profile is depleted below drained to equilibrium.

- **DSoilBubblingPressureHead.** This DDouble data object holds the bubbling pressure head parameter, or its equivalent, that is used in defining the soil water characteristic.

- **DSoilLayerLocation.** This DDouble data object holds the soil layer location, the distance from the ground surface to the bottom of the soil layer.

- **DSoilStorageAvailable.** This DDouble data object holds the value of the total available storage available within the soil profile, including the root zone deficit.

- **DSoilWaterCharacteristicOption.** This DInteger data object holds the user’s choice of soil water characteristic to be used.

- **DSoilWaterEvapDepth.** This DDouble data object holds the maximum depth to which soil water evaporation can occur when using Ritchie’s method.

- **DSWAPRunoffResistance.** This DDouble data object holds the coefficient used in the SWAP runoff equation used in PSimpleRunoff process.

- **DSWAPRunoffExponent.** This DDouble data object holds the exponent used in the SWAP runoff equation used in PSimpleRunoff process.

- **DUFBubblingPressureHead.** This DDouble data object holds the bubbling pressure head used to calculate upward flux.
• **DUFExponent.** This DDouble data object holds the exponent used to calculate upward flux.

• **DUFsaturatedHydraulicConductivity.** This DDouble data object holds the vertical saturated hydraulic conductivity used to calculate upward flux.

• **DUFHighWaterTableOption.** This DInteger data object determines whether the UF High Water Table Option is used.

• **DUpwardFluxDistance.** This DDouble data object holds the distance between the water table depth and the bottom of the soil layer for which upward flux is to be calculated.

• **DWaterTableDepth.** This DDouble data object holds the depth of the water table from the ground surface.

• **DWaterTableDepthvsAirVolume.** This DDoubleArray holds the relationship between the water table depth and air volume (void volume) of the soil profile assuming hydrostatic conditions.

• **DWetBulbTemperature.** This DDailyData data object holds the wet bulb temperature.
Figure B-1. PAcruHWTRitchieEvapoTranspiration UML diagram
Data Objects Required:

CSoil:
- DAirVolume
- DWaterTableDepth

CSoilLayer:
- DWaterFluxRecord

CGroundwaterStore:
- DDeepSeepage
- DDeepSeepageHeadBoundary
- DRestrictiveLayerHydraulicConductivity
- DRestrictiveLayerThickness
- DWaterFluxRecord

Figure B-2. PDeepSeepage UML diagram
Figure B-3. PFAO56PenmanMonteithDailyEvap UML diagram
Figure B-4. PFindNewWaterTableDepth UML diagram
PHWTCropCoeffTrans

+transpireWater() : void
+flowWater() : void
+runProcess() : void

Data Objects Required:
CClimate:
DClimate
DNPotentialEvaporation
DGrossEvapoTranspiration

CEvaporationStore:
DWaterFluxRecord

CSoil:
DAirVolume
DEvapoTranspiration
DPotEvapoTranspiration
DPotTranspiration
DTranspiredWater
DRootZoneDeficit
DWaterTableDepth

CSoilLayer
DDepth
DEvapoTranspiration
DPorosity
DPotTranspiration
DRootFrac
DSoilLayerLocation
DTranspiredWater
DWaterFluxRecord
DWiltingPoint

CVegetation
DCropCoefficient
DPlantStressIndicator
DSoilStressFraction

CLeafCanopy
DLeafAreaIndex
DLeafAreaIndexAvailability

+getPlantWaterStress() : double
+runProcess() : void
-setRequiredData() : void

Figure B-5. PHWTCropCoeffTrans UML diagram
Figure B-6. PHWTPlantWaterStress UML diagram
**Data Objects Required:**

CClimate:
- DPotentialEvaporation
- DGrossEvapoTranspiration
- DReferencePotentialEvapMethod

CEvaporationStore
- DWaterFluxRecord

CLandSegment
- DPercentSurfaceCover
- DSurfaceInfiltration

CLeafCanopy
- DLeafAreaIndex
- DLeafAreaIndexAvailability

CSoil
- DAlphaSoil
- DEvapoTranspiration
- DPotEvapoTranspiration
- DPotSoilWaterEvap
- DSoyTexture
- DSoyWaterEvapDepth
- DSoyWaterEvaporation

CSoilLayer
- DActualUpwardFlux
- DDepth
- DEvapoTranspiration
- DPotSoilWaterEvap
- DSoyWaterEvaporation
- DWaterFluxRecord
- DWiltingPoint

Figure B-7. PHWTRitchieSoilWaterEvap UML diagram
Figure B-8. PHWTSimpleEvapoTranspiration UML diagram
Figure B-9. PInitialiseSoilUFOptionHWT UML diagram
Figure B-10. PMaximumUpwardFlux UML diagram
Figure B-11. PPondedWaterEvaporation UML diagram
Data Objects Required:

CSoil:
DRDFRootDepth
DRDFCoefficient

CSoilLayer:
DDepth
DSoilLayerLocation
DRootFrac

Figure B-12. PRootDistributionFunction UML diagram
**Data Objects Required:**

- CLandSegment
- DArea
- DMaximumSurfaceDepressionStorage
- DQuickflowDepth
- DSWAPRunoffCoefficient
- DSWAPRunoffExponent
- DWaterFluxRecord

Figure B-13. PSimpleRunoff UML diagram
Figure B-14. PSoilStorageAvailable UML diagram
Data Objects Required:

CSoil:
  DSoilWaterCharacteristicOption

CSoilLayer:
  DPorosity
  DResidualMoistureContent

Figure B-15. PSoilWaterCharacteristic UML diagram
Figure B-16. PStorageLimitedInfiltration UML diagram
Figure B-17. PStorageLimitedRedistribution UML diagram
Figure B-18. PSuperSimpleEvapoTranspiration UML diagram
Figure B-19. PUpwardFlux UML diagram
APPENDIX C
HYDROLOGIC MODEL INPUT/OUTPUT VARIABLE REFERENCE

Input Variable Reference

- **BCHB1**: Shape parameter used in the soil water characteristic models. This is the bubbling pressure head, $h_b$ (cm), when the Brooks and Corey (1964) or Hutson and Cass (1987) models are used. This is $1/\alpha$ when the van Genuchten (1980) model is used ($\alpha = 1/h_b$). Values should be entered for as many soil horizons as are specified and for the surface layer (SS) when NUTRI = 1.

- **BCL1**: Shape parameter used in the soil water characteristic models. This is the pore size distribution index, $\lambda$, when the Brooks and Corey (1964) or Hutson and Cass (1987) models are used. This is $n-1$ when the van Genuchten (1980) model is used ($n = \lambda + 1$). Values should be entered for as many soil horizons as are specified and for the surface layer (SS) when NUTRI = 1.

- **DEEPH**: Deep seepage head boundary (m). This is the hydraulic head of the groundwater below the restrictive layer and is measured downwards from the ground surface. This is only used when deep seepage is simulated. This is input as a daily time-series.

- **DEEPKV**: Vertical saturated hydraulic conductivity (m/s) for the restrictive layer. This is only used when deep seepage is simulated.

- **DEEPTHK**: Thickness (m) of the restrictive layer. This is only used when deep seepage is simulated.

- **EQPET**: Variable entered to specify which method of calculating reference potential evapotranspiration is used. This is not a new parameter, but new options have been added. EQPET = 116 FAO Penman-Monteith Equation daily input

- **EVDEP**: Soil evaporation depth used in Ritchie’s method for soil evaporation (EVTR = 2). This is the depth from which evaporation from soil occurs (units of m). Recommended values range from 0.1 to 0.15 m (Allen et al. 1998).

- **EVTR**: Option to designate the choice of actual evapotranspiration simulation. This is not a new parameter, but a new option has been added. EVTR = 3 Super Simple Evapotranspiration option.

- **HEAD1**: Pressure head (cm) below which transpiration (or evapotranspiration) is reduced to zero under soil water excess conditions (anaerobiosis point).
• **HEAD2**: Pressure head (cm) below which transpiration (or evapotranspiration) begins to be reduced below the potential rate due to soil water excess conditions.

• **HEAD3**: Pressure head (cm) above which transpiration (or evapotranspiration) begins to be reduced below the potential rate due to soil water limited conditions.

• **HEAD4**: Pressure head (cm) above which transpiration (or evapotranspiration) begins to be reduced to zero due to soil water limited conditions. This is the pressure head at the wilting point.

• **IAVP**: Option to calculate actual vapor pressure by different methods. This is currently used only with the FAO Penman-Monteith equation. IAVP = 0 Mean Relative Humidity is used; IAVP = 1 Maximum Relative Humidity is used; IAVP = 2 Maximum and Minimum Relative Humidity are used; IAVP = 3 Dew Point Temperature is used, IAVP = 4 Psychrometric data is used (Web Bulb Temperature).

• **IDEEP**: Option that specifies whether deep seepage is simulated. IDEEP = 0 no deep seepage; IDEEP = 1 deep seepage is simulated.

• **ISWAVE**: Variable entered to specify whether incoming solar radiation is available as an input, and if not which method will be used to estimate it. This is not a new parameter, but new options have been added. ISWAVE = 2 incoming solar radiation is estimated using the equation of Hargreaves and Samani (1982). KT coefficient is estimated using the equation of Samani (2000).

• **MAXRH**: Maximum relative humidity (%).

• **MINRH**: Minimum relative humidity (%).

• **NUMHOR**: Variable specifies the number of soil horizons for which parameters are entered.

• **PSYCOEF**: Psychrometric coefficient of the instrument. Values recommended by the FAO in Irrigation and Drainage Paper # 56. PSYCOEF = 0.000662 for ventilated (Asmann type) psychometers, with air movement of 5 m/s. PSYCOEF = 0.000800 for naturally ventilated pyschrometers (about 1 m/s).

• **PSYCOEF**: 0.001200 for non-ventilated psychrometers installed indoors.

• **RDFRD**: Root density function root depth (m). This is the root depth used when EVTR = 1 or 2 and the UFHWT option is used. This is input as a daily time-series.

• **RDFC**: Root density function coefficient in the RDF proposed by Hooland et al. (1981). This is used when EVTR = 1 or 2 and the UFHWT option is used. This is input as a daily time-series.
- **RSMI**: Residual soil moisture (cm³/cm³) used in the soil water characteristic models. Values should be entered for as many soil horizons as are specified and for the surface layer (SS) when NUTRI = 1.

- **SWAPRES**: Resistance coefficient in the runoff routing coefficient in the SWAP model. This is only used when a single land segment is simulated (lumped model).

- **SWAPEXP**: Exponent in the SWAP runoff equation. This is only used when a single land segment is simulated (lumped model).

- **SWCHAR**: Variable specifies the user’s choice of soil water characteristic model that will be used. SWCHAR = 1 Brooks and Corey (1964); = 2 Hutson and Cass (1987); = 3 van Genuchten (1980).

- **TDEW**: Dew point temperature (°C).

- **UFEXP1**: Exponent used to calculate upward flux. Values should be entered for as many soil horizons as are specified and for the surface layer (SS) when NUTRI = 1.

- **UFHB1**: Bubbling pressure head (cm) used to calculate upward flux. Values should be entered for as many soil horizons as are specified and for the surface layer (SS) when NUTRI = 1.

- **UFKSAT1**: Saturated hydraulic conductivity (m/s) used to calculate upward flux. Values should be entered for as many soil horizons as are specified and for the surface layer (SS) when NUTRI = 1.

- **UFHWT**: Option to specify mode of model simulation. When turned on the UF High Water Table processes are used. UFHWT = NO (0) Original ACRU; = YES (1) High water table processes are used.

- **WETBULB**: Wet bulb temperature (°C).

- **WTDEP**: Initial water table depth (meters). As measured from ground surface.

### Output Variable Reference

- **AUF**: Actual upward flux (m) calculated on a given day.

- **DEEP**: Amount of deep seepage (m) that occurs on a given day. Deep seepage is positive downwards (leaving the soil profile) and negative upwards (entering the soil profile).

- **DPOND**: Depth of ponded water (mm) on the land surface.
- **DTOE1**: Drained to equilibrium water content for a soil layer. This is the water content that the soil layer would contain by assuming a hydrostatic profile, no hysteresis, and no depletion by evapotranspiration.

- **WTDEP**: Water table depth (meters). As measured from ground surface.
Introduction

This manual describes the technical details of the High Water Table simulation component of the ACRU model developed at the University of Florida. The model uses physically based approximations which are suitable for simulating a dynamic water table on a daily time-step. Appropriate approximations are made specifically for highly conductive soils with a shallow water table. In order to accurately simulate a dynamic water table the importance of upward gradients within the unsaturated portion of the soil and the contribution of a high water table to evapotranspiration is accounted for. Parameters which are input or output variables are noted in italicized capital letters in parenthesis. In order to use the High Water table simulation the switch variable (UFHWT) must be on.

Simulation of the Water-Table and Soil Moisture Distribution

The water table and soil moisture distribution is simulated by assuming a steady-state condition within the soil profile. In addition, hysteresis is ignored. Thus, the soil profile is assumed to be hydrostatic, and the soil moisture distribution within the soil profile at a “drained to equilibrium” condition with the water table. Under these conditions the volume of water, or conversely the air volume, within the soil profile is a unique function of the water table depth and can be easily defined by a water characteristic function for each soil layer.
To make the model more realistic, the soil moisture distribution is allowed to deviate from steady-state due to the removal of water from the profile by evapotranspiration. This deviation from steady-state creates a depleted root zone. The amount to which the root zone is depleted is referred to here as the root zone deficit. This deviation from steady-state implies that an upward gradient is induced between the water table and the depleted root zone. Water may move upwards in the soil profile by upward flux in response to this gradient. This upward movement of water defines the connectivity between a shallow water table and evapotranspiration.

**Soil Water Characteristic Functions**

Three different water characteristic functions have been included in the model: Brooks and Corey (1964), Hutson and Cass (1987), and van Genuchten (1980). The Brooks and Corey (1964) water characteristic curve is defined as:

\[
\theta(h) = \theta_r + \left( \theta_s - \theta_r \right) \left( \frac{h_b}{h} \right)^{\lambda} \quad \text{for} \quad h > h_b \\
\theta(h) = \theta_s \quad \text{for} \quad h \leq h_b
\]  

\[\text{(D-1a)}\]

\[\text{(D-1b)}\]

where \(\theta(h)\) is the moisture content (\(\text{cm}^3/\text{cm}^3\)) as a function of capillary pressure head \(h\) (cm) (taken as positive in the unsaturated portion of the soil above the water table); \(\theta_r\) \((\text{RSM1})\) is the residual moisture content (\(\text{cm}^3/\text{cm}^3\)), which is taken as the moisture content at infinite capillary pressure head; \(\theta_s\) \((\text{PO1})\) is the saturated moisture content (\(\text{cm}^3/\text{cm}^3\)) and is taken to be equal to the porosity of the soil; \(h_b\) \((\text{BCHB1})\) is the bubbling pressure head, or air entry pressure head, of the soil (cm); and \(\lambda\) \((\text{BCLI})\) is the pore size distribution index (\(-\)).
For a soil profile that is assumed to be in hydrostatic equilibrium with a water table the capillary pressure head is equivalent to the distance above the water table and thus the water content can be expressed as a function of distance above the water table. Equation (D-1) then becomes:

\[ \theta(z) = \theta_r + (\theta_s - \theta_r) \left( \frac{h_b}{z} \right)^\lambda \quad \text{for} \quad z > h_b \]  

\[ \theta(z) = \theta_s \quad \text{for} \quad z \leq h_b \]  

where \( z \) is the distance above the water table (cm). An example of the Brooks and Corey water characteristic function can be seen in Figure D-1.

Hutson and Cass (1987) modified the water characteristic curve of Brooks and Corey (1964) by replacing the sharp discontinuity at the bubbling pressure with a parabolic equation without requiring any additional parameters to be specified. The Hutson and Cass (1987) water characteristic curve is defined as:

\[ \theta(h) = \theta_r + (\theta_s - \theta_r) \left( \frac{h_b}{h} \right)^\lambda \quad \text{for} \quad h > h_i \]  

\[ \theta(h) = \theta_r + (\theta_s - \theta_r) \left[ \frac{h^2 \left( 1 - \frac{\theta_i}{\theta_s} \right) \left( \frac{\theta_s}{\theta_i} \right)^{\frac{\lambda}{2}}}{h_i^2} \right] \quad \text{for} \quad h \leq h_i \]  

where \( \theta_i \) is the water content (cm\(^3\)/cm\(^3\)) at the inflection point in the water characteristic curve at capillary pressure head \( h_i \) (cm). \( \theta_i \) and \( h_i \) are defined as:

\[ \theta_i = \frac{2\theta_s}{\lambda + 2} \]  

(D-4)
\[ h_i = h_b \left( \frac{\lambda + 2}{2} \right)^{\frac{1}{\lambda}} \quad (D-5) \]

For a soil profile that is assumed to be in hydrostatic equilibrium with a water table the capillary pressure head is equal to the distance from the water table. Thus equation (D-3) can be expressed as:

\[ \theta(z) = \theta_r + (\theta_s - \theta_r) \left( \frac{h_b}{z} \right)^{\lambda} \quad \text{for} \quad z > h_i \quad (D-6a) \]

\[ \theta(z) = \theta_r + (\theta_s - \theta_r) \left[ 1 - \frac{z^2 \left( 1 - \frac{\theta_r}{\theta_s} \right) \left( \frac{\theta_s}{\theta_r} \right)^{\frac{1}{\lambda}}}{h_b^{\lambda}} \right] \quad \text{for} \quad z \leq h_i \quad (D-6b) \]

where \( z \) is the distance above the water table (cm). An example of the Hutson and Cass water characteristic function can be seen in Figure D-1.

The van Genuchten (1980) water characteristic curve is defined as:

\[ \theta(h) = \theta_r + (\theta_s - \theta_r) \left[ \frac{1}{1 + (\alpha h)^n} \right]^m \quad (D-7) \]

where \( \alpha, n, \) and \( m \) are empirical parameters. It is assumed that \( m = 1 - 1/n. \)

For a soil profile that is assumed to be in hydrostatic equilibrium with a water table the capillary pressure head is equal to the distance from the water table. Thus equation (D-7) can be expressed as:

\[ \theta(z) = \theta_r + (\theta_s - \theta_r) \left[ \frac{1}{1 + (\alpha z)^n} \right]^m \quad (D-8) \]

where \( z \) is the distance above the water table (cm). An example of the van Genuchten water characteristic function can be seen in Figure D-1.
Figure D-1. The three water characteristic functions with the same input parameters: $h_b = 30$ cm and $\lambda = 1$ (van Genuchten parameters are $\alpha = 0.0333$ cm$^{-1}$, $n = 2$, and $m = 0.5$). Note that the curves are similar but not the same. Specifically, the areas under the curves are not the same.

When entering parameters for the particular soil water characteristic function chosen the user enters the parameters as $h_b$ ($BCHB1$) and $\lambda$ ($BCL1$) regardless of the soil water characteristic function used. When these parameters are entered for the van Genuchten (1980) equation the following equalities are assumed: $\alpha = h_b^{-1}$, and $n = \lambda + 1$. This equivalency between model parameters was noted by van Genuchten (1980) but does not necessarily imply that the resulting soil water characteristics are equivalent.
This can be seen in Figure D-1. It should be noted that while the same values for the soil water characteristic parameters can be used with different soil water characteristic functions this does not mean that these soil water characteristics are the same!

The equivalence between parameters in the Brooks and Corey (1964) and van Genuchten (1980) soil water characteristic functions have been explored by Lenhard et al. (1989) in terms of preserving the shape of the water characteristic function and by Morel-Seytoux et al. (1996) in terms of preserving the “effective capillary drive” using the hydraulic conductivity functions of Burdine (1953) and Maulem (1976). This effective capillary drive is defined as (Morel-Seytoux et al. 1996):

\[
H_{cm} = \int_{0}^{\infty} k_{rw} dh
\]  

(D-9)

These two methods of determining equivalent parameters do not produce the same results in terms of both the shape of the water characteristic curve and the effective capillary drive. The desired result will dictate the method used.

**Determining Upper-Limit Water Contents of Soil Layers**

Assuming hydrostatic equilibrium, the average water contents for discrete soil layers (referred to here as either the “upper limit” or “drained to equilibrium” (DTOE1) water content), given the water table depth and soil layer parameters, can be found to be:

\[
\bar{\theta} = \frac{1}{z_2 - z_1} \int_{z_1}^{z_2} \theta(z)dz
\]  

(D-10)

where \( z_2 \) and \( z_1 \) are the distances from the water table to the top and bottom of the soil layer, respectively.

The Brooks and Corey (1964) water characteristic function may be integrated analytically. The integral of equation (D-2) is:
\[ \int \theta(z)dz = \frac{(\theta_s - \theta_r) h_b \lambda}{1 - \lambda} z \quad \text{for } \lambda \neq 1 \textbf{ and } z > h_b \quad (D-11a) \]

\[ \int \theta(z)dz = \frac{(\theta_s - \theta_r) h_b \ln(z)}{1 - \lambda} \quad \text{for } \lambda = 1 \textbf{ and } z > h_b \quad (D-11b) \]

\[ \int \theta(z)dz = \theta_r z \quad \text{for } z \leq h_b \quad (D-11c) \]

The Hutson and Cass (1987) water characteristic function may also be integrated analytically. The integral of equation (D-6) is:

\[ \int \theta(z)dz = \frac{(\theta_s - \theta_r) h_b \lambda}{1 - \lambda} z \quad \text{for } \lambda \neq 1 \textbf{ and } z > h_i \quad (D-12a) \]

\[ \int \theta(z)dz = \frac{(\theta_s - \theta_r) h_i \ln(z)}{1 - \lambda} \quad \text{for } \lambda = 1 \textbf{ and } z > h_i \quad (D-12b) \]

\[ \int \theta(z)dz = \theta_r z + \left( \frac{\theta_s - \theta_r}{\theta_i} \right) h_i \quad \text{for } z \leq h_i \quad (D-12c) \]

The van Genuchten (1980) water characteristic function can not be integrated analytically, so the integral of the function must be evaluated numerically. In order to retain computational efficiency, the water characteristic equation is integrated by Gauss-Legendre quadrature. Compared to basic numerical techniques such as the trapezoidal rule, Gauss-Legendre quadrature is not constrained to evaluating the functions at fixed locations and can achieve higher accuracy for a given number of integration points (Press 1986). Gauss-Legendre quadrature may only be used for a continuous function and not for tabulated data. Gauss-Legendre quadrature chooses the points, according to the method of undetermined coefficients, to evaluate the function to be integrated in such a
manner that positive and negative errors are balanced (Chapra and Canale 1998).

Specifically, a five-point Gauss-Legendre integration is performed for a single soil layer in order to integrate equation (D-8).

**Developing the Relationship between the Water-Table Depth and Soil Air Volume**

The unique relationship between the water table and air volume in the soil profile, assuming steady-state and neglecting, for now, any depletion of the root zone is developed by determining the sum of the air volume of each discrete soil layer for a given depth of the water table. This is determined as the difference between the porosity and the drained to equilibrium water content of the soil layer multiplied by the thickness of the soil layer.

This relationship is applied to determine a new depth of the water table given changes within the soil profile (i.e. upward flux or infiltration in excess of the root zone deficit, etc). Since this relationship is defined for the entire soil profile, as defined by the user, the water table can not be allowed to drop below the bottom of the soil profile.

It should be noted that the relationship between the water table depth and soil air volume does not account for the root zone depletion. If it did, the relationship would no longer be unique.

**Evapotranspiration and the Deviation from a Steady-State Profile**

Since a steady-state profile is unrealistic in a soil profile that is actively evapotranspiring, the root zone (where evaporation or transpiration from the soil is taking place) is allowed to deviate from steady-state by being reduced below the hydrostatic water content. The degree to which the root zone is below the “drained to equilibrium” water content is called the root zone deficit. As will be discussed below, the creation of a depleted root zone creates an upward gradient between the root zone and the soil below.
This gradient will induce an upward flux of water within the soil. This depleted root zone also must first be replenished before infiltrating water can cause a rise in the water table.

**New reference potential evapotranspiration methods**

The standardized Penman-Monteith grass reference potential evapotranspiration adopted by the Food and Agricultural Organization in the FAO Irrigation and Drainage Paper No. 56 (Allen et al. 1998) has been added to the model. The reference surface assumed is “a hypothetical grass reference crop with an assumed crop height of 0.12 m, a fixed surface resistance of 70 s/m and an albedo of 0.23”. This standardized Penman-Monteith equation is referred to here as FAO56PM (mm/day):  

\[
ET_0 = \frac{0.408\Delta(R_n - G) + \gamma \frac{900}{T_{mean} + 273}u_2(e_s - e_a)}{\Delta + \gamma(1 + 0.34u_2)}
\]

where \(R_n\) is the incoming net radiation (MJ m\(^{-2}\) day\(^{-1}\)), \(G\) is the soil heat flux density (MJ m\(^{-2}\) day\(^{-1}\)) as is assumed to be zero for daily calculations, \(T_{mean}\) (\(T\text{MEAN}\)) is the mean daily air temperature at 2 m height (\(^\circ\text{C}\)), \(u_2\) (\(W\text{IND}\) and is input to the model as km/day) is the wind speed at 2 m height (m/s), \(e_s\) is the saturated vapor pressure (kPa), \(e_a\) is the actual vapor pressure (kPa), the quantity \(e_s - e_a\) is the vapor pressure deficit (kPa), \(\Delta\) is the slope of the vapor pressure curve (kPa/\(^\circ\text{C}\)), and \(\gamma\) is the psychrometric constant (kPa/\(^\circ\text{C}\)).

The psychrometric constant, \(\gamma\) is given by:

\[
\gamma = \frac{c_p \rho}{\varepsilon \lambda} = 0.665 \times 10^{-3} \rho
\]

where \(c_p\) is the specific heat of moisture at constant pressure (MJ kg\(^{-1}\) \(^\circ\text{C}\)^\(^{-1}\)) (for average atmospheric conditions it is assumed constant at 1.013x10\(^{-3}\)), \(\lambda\) is the latent heat of
vaporization and is assumed to be constant at 2.45 MJ/kg, $\varepsilon$ is the ratio of the molecular weight of water vapor to dry air (0.622), and $P$ is the atmospheric pressure (kPa) and is estimated as:

$$P = 101.3 \left( \frac{293 - 0.0065z}{293} \right)^{5.26}$$ (D-15)

where $z$ ($ELEV$) is the elevation above sea level (m).

The saturated vapor pressure, $e_s$ is estimated from the mean of the saturated vapor pressure at the daily maximum and minimum temperatures (°C) ($T_{MAX}$ and $T_{MIN}$):

$$e_s = \frac{e^o(T_{max}) + e^o(T_{min})}{2}$$ (D-16)

where:

$$e^o(T) = 0.6108\exp\left[\frac{17.27T}{T + 237.3}\right]$$ (D-17)

The actual vapor pressure, $e_a$ can be determined (in order of desirability) from either the dew point temperature ($T_{DEW}$), pyschrometric data, maximum and minimum relative humidity ($MAXRH$ and $MINRH$), just maximum relative humidity, or from average relative humidity ($RH$). Using the dewpoint temperature (when $IAVP = 3$):

$$e_a = e^o(T_{dew})$$ (D-18)

Using psychrometric data ($IAVP = 4$):

$$e_a = e^o(T_{wet}) - \gamma_{psy} (T_{dry} - T_{wet})$$ (D-19)

where $T_{wet}$ is the wet bulb temperature °C ($WETBULB$), $T_{dry}$ is the dry bulb temperature (assumed to be equal to the mean daily temperature), and $\gamma_{psy}$ is the psychrometric constant of the instrument:
\[ \gamma_{psy} = a_{psy} P \]  

where \( a_{psy} \) is the psychrometric coefficient of the instrument (\(^{\circ}\mathrm{C}^{-1}\)) (\text{PSYCOEF}) and \( P \) is the atmospheric pressure. Using maximum and minimum relative humidity \( e_a \) can be estimated as \((\text{IAVP} = 2)\):

\[
e_a = \frac{e^\circ(T_{\min}) \frac{RH_{\max}}{100} + e^\circ(T_{\max}) \frac{RH_{\min}}{100}}{2}
\]  

\( e_a \) can be estimated as \((\text{IAVP} = 1)\):

\[
e_a = e^\circ(T_{\max}) \frac{RH_{\max}}{100}
\]  

When only mean relative humidity is available \( e_a \) can be estimated from \((\text{IAVP} = 0)\):

\[
e_a = \frac{RH_{\text{mean}}}{100} \left[ \frac{e^\circ(T_{\max}) + e^\circ(T_{\min})}{2} \right]
\]

The slope of the saturation vapor pressure curve, \( \Delta \) is given by:

\[
\Delta = \frac{4098e^\circ(T_{\text{mean}})}{(T_{\text{mean}} + 237.3)^{\frac{3}{2}}}
\]

The incoming net radiation \( R_n \) is the difference between the net incoming shortwave radiation, \( R_{ns} \) and the net incoming longwave radiation, \( R_{nl} \):

\[
R_n = R_{ns} - R_{nl}
\]

The net shortwave radiation, \( R_{ns} \) is determined from the albedo, \( \alpha \) (assumed to be 0.23) and the incoming shortwave radiation \( R_s \) (\text{RADMET}):

\[
R_{ns} = (1 - \alpha)R_s
\]

\( R_s \) can be supplied as daily input to the model \((\text{ISWAVE} = 0)\) or estimated from the Angstrom equation \((\text{ISWAVE} = 1)\):
\[
R_s = \left( a_s + b_s \frac{n}{N} \right) R_a
\]

where \( n \) is the number of sunshine hours, \( N \) is the maximum possible number of sunshine hours and can be determined from the sunset hour angle, \( \omega_s \) (defined below):

\[
N = \frac{24}{\pi} \omega_s
\]

\( R_a \) is the extraterrestrial radiation (MJ m\(^{-2}\) day\(^{-1}\)), \( a_s \) (ACONS) is a regression constant that expresses the fraction of extraterrestrial radiation reaching the earth on overcast days (\( n = 0 \)), and the quantity \( a_s + b_s \) (BCONS) is the fraction of extraterrestrial radiation reaching the earth on clear days (\( n = N \)). \( R_s \) may also be estimated using the equation of Hargreaves and Samani (1982) (ISWAVE = 2):

\[
R_s = (KT)(R_a)(TD)^{0.5}
\]

where \( TD \) is the difference between maximum and minimum air temperatures (°C) (\( TMAX \) and \( TMEAN \)) and \( KT \) is an empirical constant. Samani (2000) developed an equation to determine \( KT \) as a function of \( TD \):

\[
KT = 0.00185(TD)^2 - 0.0433TD + 0.4023
\]

using 25 years of data for the continental U.S. This method of estimating \( KT \) is included in the model.

Extraterrestrial radiation, \( R_a \) can be estimated from:

\[
R_a = \frac{24 \cdot 60}{\pi} G_{sc} d_r [\omega_s \sin(\phi) \sin(\delta) + \cos(\phi) \cos(\delta) \sin(\omega_s)]
\]

where \( G_{sc} \) is the solar constant (the amount of radiation striking a surface perpendicular to the sun’s rays at the top of the earth’s atmosphere), 0.0820 MJ m\(^{-2}\) min\(^{-1}\), \( d_r \) is the inverse relative distance between the earth and sun (corrects for eccentricity of the earth’s orbit):
\[ d_r = 1 + 0.033 \cos \left( \frac{2\pi}{365} J \right) \tag{D-32} \]

where \( J \) number of the day of year. \( \delta \) is the solar declination and is given by:

\[ \delta = 0.409 \sin \left( \frac{2\pi}{365} J - 1.39 \right) \tag{D-33} \]

\( \varphi \) is latitude of the location, expressed in radians, and \( \omega_s \) is the sunset hour angle and is given by:

\[ \omega_s = \arccos \left[ -\tan(\varphi) \tan(\delta) \right] \tag{D-34} \]

The net longwave radiation, \( R_{nl} \) can be estimated as (ISWAVE = 0):

\[ R_{nl} = \sigma \left[ \frac{T_{\text{max},K}^4 + T_{\text{min},K}^4}{2} \right] \left( 0.34 - 0.14 \sqrt{e_a} \left( 1.35 \frac{R_s}{R_{so}} - 0.35 \right) \right) \tag{D-35} \]

where the first term represents the theoretical maximum longwave radiation to leave the earth’s surface, the second term is a humidity correction, and the third term is a correction for cloudiness. \( \sigma \) is the Stefan-Boltzmann constant (4.903x10^{-9} MJ K^{-4} m^{-2} day^{-1}), \( T_{\text{max},K} \) and \( T_{\text{min},K} \) are the maximum and minimum temperatures in Kelvin, and \( R_{so} \) is the clear sky radiation (MJ m^{-2} day^{-1}) and can be estimated as (ISWAVE = 0 or 1):

\[ R_{so} = (a_s + b_s)R_a \tag{D-36} \]

or when values of \( a_s \) and \( b_s \) are not available (ISWAVE = 2):

\[ R_{so} = \left( 0.75 + 2 \times 10^{-5} z \right) R_a \tag{D-37} \]

The use of the FAO56PM for calculating reference potential evapotranspiration is not restricted to use with only the High Water Table simulation components discussed in this manual, it may also be used with the original version of the ACRU model.
It should be noted that the FAO56PM calculates reference evapotranspiration for a reference crop while the remaining methods of calculating evapotranspiration in ACRU use an open water surface as the reference. For further information about the FAO56PM the reader is referred to Allen et al. (1998).

**Actual evapotranspiration methods**

The two existing actual evapotranspiration calculation methods of ACRU have been adapted for use when using the High Water Table simulation option. In addition, a third simple actual evapotranspiration method has also been added. For all methods evapotranspiration is calculated in a top-down approach. First the evaporation demand is applied to intercepted water, next to ponded water on the ground surface, and then to the soil as evaporation and transpiration.

The ACRU model calculates actual evapotranspiration as either a lumped quantity ($EVTR = 1$) or by determining soil evaporation and transpiration separately ($EVTR = 2$) using Ritchie’s method (Ritchie 1972). In order to use the methods when simulating a high water table the removal of water from below the water table (when the water table is high) must be accounted for when determining an updated depth of the water table. Evaporation or transpiration from a soil layer containing or below the water table will change the air volume and hence water table depth directly. Additionally, the removal of water from above the water table (which creates or contributes to the root zone deficit) is accounted for.

When simulating evapotranspiration as a lumped quantity ($EVTR = 1$) the reference potential evapotranspiration (input or calculated by the model) is multiplied by the crop coefficient ($VICAY$). An approximate method is used to divide actual evapotranspiration
between soil water evaporation and plant transpiration for use for plant uptake of solutes and nutrients:

\[
T = 0.95T \left( \frac{V1CAY - 0.2}{0.8} \right) \quad \text{for} \quad V1CAY > 0.2 \tag{D-38a}
\]

\[
T = 0 \quad \text{for} \quad V1CAY <= 0.2 \tag{D-38b}
\]

When using Ritchie’s method to determine evapotranspiration the potential soil water evaporation and plant transpiration are separated as a function of the leaf area index (\(V1LAI\)):

\[
T_p = (0.7V1LAI^{0.5} - 0.21)E_o \quad \text{for} \quad V1LAI < 2.7 \tag{D-39a}
\]

\[
T_p = 0.95E_o \quad \text{for} \quad V1LAI >= 2.7 \tag{D-39b}
\]

where \(T_p\) is the potential transpiration (mm) with the remaining potential evapotranspiration being apportioned to soil evaporation. The potential transpiration is further adjusted by being multiplied by a crop coefficient.

Potential evaporation is multiplied by a factor of 1.15 to account for the difference in albedo between bare soil and a vegetated surface as recommended by Allen et al. (1998). Potential soil evaporation \(E_p\) is further adjusted for the percent surface cover by mulch, litter, \(C_s\) (PSUCO):

\[
E_p = E_p \frac{C_s}{100} \tag{D-40}
\]

Soil water evaporation takes place down to a user defined depth of the soil (\(EVDEP\)) with recommended values ranging from 0.1 to 0.15 m (Allen et al. 1998).

According to Ritchie’s method actual evaporation from the soil surface continues at a maximum rate equal to the potential rate (Stage 1 evaporation) until the accumulated soil water evaporation exceeds the stage 1 upper limit, \(U_i\) which is defined as (mm):
\[ U_1 = (\alpha_s - 3)^{0.42} \]  
(D-41)

where \( \alpha_s \) is a soil water transmission parameter that is related to the texture of the soil (TEXT). After \( U_1 \) is exceeded soil water evaporation proceeds at a reduced (Stage 2) rate:

\[ E = \alpha_s t_d^{0.5} - (t_d - 1)^{0.5} \]  
(D-42)

where \( t_d \) is the number of days since \( U_1 \) has been exceeded.

The contribution of rainfall and upward flux from the water table to the root zone for a given day reduces the accumulated soil water evaporation as it is accounted for to determine the transition between Stage 1 and 2 evaporation.

The new actual evapotranspiration method that has been added to the model (EVTR = 3) is a simplification to the lumped evapotranspiration (EVTR = 1) method described above. The simplifications include neglecting water stress (actual evapotranspiration occurs at its maximum rate until the wilting point is reached) and the apportionment of evapotranspiration to different layers. The original model applies evapotranspiration (or transpiration) to individual soil layers according to the fraction of plant roots within that layer. The new method described here applies evapotranspiration in a top-down approach. In this manner evapotranspiration occurs from the top-most soil layer until the wilting point is reached and then applies the remainder to the layer below until the bottom of the root zone (the last layer to contain roots) is reached.

In the High Water Table option the response of transpiration (when using Ritchie’s method) or lumped evapotranspiration to water-stress is different than in the original ACRU2000 model. Here the reduction below the potential rate is made using the method proposed by Feddes et al. (1978) as shown in Figure D-2 (and used only when \( EVTR = 1 \).
or 2). Transpiration (or evapotranspiration) is reduced to zero when soil pressure heads are below $h_1$ ($HEAD1$ representing oxygen depletion) and above $h_4$ ($HEAD4$ above the wilting point). Between $h_2$ ($HEAD2$) and $h_3$ ($HEAD3$) transpiration proceeds at its maximum rate. Between $h_1$ and $h_2$, and $h_3$ and $h_4$ a linear reduction of transpiration with pressure head is assumed.

![Figure D-2. Plant water stress as a function of soil water pressure head.](image)

Additionally, the root distribution function proposed by Hoogland et al. (1981) has been added to the model:

$$g(d) = \frac{c(2d - L) + L}{L^2}$$

where $c$ ($RDFC$) is a shape parameter (varies between -1, a linear decrease in density to 0, constant density with depth as seen in Figure D-3), $d$ is the depth from the ground surface, and $L$ ($RDFRD$) is the depth of plant roots.
Figure D-3.  Root distribution function of Hoogland et al. (1980).

**Upward flux of water in response to a depleted root zone**

In response to the deviation from a steady-state profile created by a depleted root zone an upward gradient exists within the soil profile. For simulating the upward flux of water between the water table and the depleted root zone a steady-state approximation is made. The steady-state upward flux is given by Darcy’s Law:

\[ q = K(h) \left( \frac{dh}{dz} - 1 \right) \]  

where \( q \) is the upward flux (m/s), \( K(h) \) is the hydraulic conductivity (m/s), and \( z \) is the height above the water table (m). Rearranging and integrating equation (D-12) yields:

\[ z = \int_0^h \frac{dh}{1 + q / K(h)} \]

\[(D-45)\]
Assuming a relationship between $K$ and $h$ allows eqn. (13) to be solved with a lower boundary condition of $h = 0$ at $z = 0$ (at the water table). The complexity of analytical or numerical solutions to eqn. (13) depends on the choice of $K(h)$ relationship. Once integrated values of $q$ can then be determined at a given height above the water table and soil suction. An upper limit of integration of $h = \infty$ is typically used for simplicity. Gardner (1958) showed that this upper limit is appropriate since upward flux quickly approaches a limiting value as soil suction increases. Thus solving equation (D-13) with this upper boundary condition gives the maximum steady-state upward flux.

Currently an approximate method developed by Anat et al. (1965) is used to calculate the maximum steady state upward flux:

$$q = K_s \left[ 1 + \frac{1.886}{\eta^2 + 1} \left( \frac{d}{h_b} \right)^\eta \right]$$

where $K_s$ is the saturated hydraulic conductivity of the soil (m/s) ($\text{UFKSAT1}$), $h_b$ is the bubbling pressure head ($\text{UFHB1}$), $d$ is the distance between the water table and the depleted root zone (m), and $\eta$ ($\text{UFEXP1}$) is related to the pore size distribution index of the Brooks and Corey (1964) model by:

$$\eta = 2 + 3\lambda$$

This relationship for upward flux assumes that the soil profile is homogeneous. For each soil layer a set of parameters are entered ($\text{UFHB1}$, $\text{UFEXP1}$, and $\text{UFKSAT1}$). These parameters represent all of the soil layers below it and can be best obtained by fitting the above equation to a steady-state solution of Richard’s equation for all water table depths below the layer in question.
The amount of upward flux that actually occurs on a given day is the minimum of the maximum value calculated and the amount to which the root zone is depleted (and is now fully replenished by upward flux). The amount of upward flux that has occurred increases the air volume of the soil profile even though there has been no net loss of water. The water has contributed to replenishing the root zone deficit.

**Infiltration and Redistribution of Infiltrated Water**

Water on the ground surface is allowed to infiltrate until the soil profile becomes completely saturated. Thus only saturated-excess ponding and runoff is simulated. Water that infiltrates into the soil profile will cause a rise in the water table after any depleted root zone is replenished. This depleted root zone is created by the removal of water from the soil profile by evapotranspiration. The amount of infiltrating water in excess of that required to replenish the root zone is the exact amount by which the air volume of the soil will be decreased. By knowing this change of air volume the new water table depth can be found easily from the water table vs. air volume relationship previously developed.

**Deep Seepage**

Deep seepage occurs through the restrictive layer located below the soil profile according to Darcy’s Law. The restrictive layer is defined by its thickness in meters \((\text{DEEPTHK})\) and saturated hydraulic conductivity in m/s \((\text{DEEPKV})\). Currently the hydraulic head below the restrictive layer is entered as a time-series \((\text{DEEPH})\).

**Determining a New Water-Table Depth, Upper-Limit Water Contents, and Redistribution of Soil Water**

Each process which adds or removes water from the soil profile changes the value of the air volume in the soil or the root zone deficit or both. As explained earlier, the root
zone deficit is created by evapotranspiration from soil layers within the root zone. It is replenished by upward flux, which simultaneously increases the air volume (causing a drop in the water table) or it is replenished by infiltration. Infiltration in excess of the root zone deficit will decrease the air volume (causing a rise in the water table). Evapotranspiration from a soil layer containing or below the water table increases the air volume directly.

Given the value of the air volume, after all of these processes have added or removed water, a new water table is found from the unique relationship discussed earlier. Once the new water table depth is found the new upper limit water contents can be defined for each soil layer. Once these upper limits are defined water is then redistributed (upwards or downwards) accordingly.

Other Phenomena that will change the Water Table Depth and Soil Moisture Distribution

Other processes that will cause a change in the water table depth must account for changes in the air volume in the soil profile. In addition care must be taken in determining when a portion of water must replenish a depleted root zone (and hence not affect the air volume). The value of the root zone deficit represents the degree to which the entire root zone (not just a single soil layer) is depleted. In order to determine which soil layers contain this root zone deficit the current water content and the drained to equilibrium water content must be known.

Runoff from a Lumped Model

When the model is used in lumped mode (only a single land segment) a simple, empirical equation is used to route runoff from the land segment. The equation used is
identical to that used in the SWAP model (Kroes and van Dam 2003) and is similar to that used in the FHANTM model (Tremwel 1992):

\[
q = \frac{1}{\gamma} \left( h_{\text{pond}} - z_{\text{dep}} \right)^\beta
\]  

where \( q \) is the runoff depth (mm/day), \( \gamma \) is the runoff resistance coefficient (day) to be calibrated, \( h_{\text{pond}} \) is the depth of ponded water, \( z_{\text{dep}} \) is the maximum depressional storage which must be filled prior to the occurrence of runoff, and \( \beta \) is an exponent to be calibrated but is usually taken as 1.67 for turbulent flow in the FHANTM model.
APPENDIX E
NUTRIENT PROCESS AND DATA OBJECTS AND DESCRIPTIONS

Process Objects

- PDetermineLayerPressureHeads
- PHWTAmmonification
- PHWTDenitrification
- PHWTImmobilization
- PHWTNitrification
- PHWTNutrientInputs
- PHWTPMineralization
- PHWTSurfaceTransport
- PHWTSubsurfaceTransport
- PMixingZoneExchangeModel
- PNutrientTransformationProcess
- PNutrientTransportProcess

Data Objects

- DDenitrificationThreshold
- DLayerPressureHead
- DNutrientFluxRecord
- DPondedAmmoniumNConc
- DPondedLabilePConc
- DPondedNitrateNConc

Description of Process Objects

- **PDetermineLayerPressureHeads.** This process determines the average pressure head in each soil layer on each day for use in soil moisture factors in nutrient transformations. This process also sets the denitrification threshold of each soil layer.

- **PHWTAmmonification.** This process determines the first step of the mineralization of organic N to ammonium. This process is identical to the original PAmmonification process with the exception of the different soil moisture reduction factors used.
• **PHWTDenitrification.** This process determines the denitrification of nitrate to the atmosphere. This process is the same as PDentification with the only difference being the soil moisture response factor.

• **PHWTImmobilization.** This process determines the immobilization of N and P by soil microbial biomass in the case that the decomposing organic matter is nutrient poor. This process is the same as PImmobilization with the only difference being the soil moisture response factor.

• **PHWTNitrification.** This process determines the nitrification of ammonium N to nitrate N. This process is the same as PNitrification with the only difference being the soil moisture response factor.

• **PHWTNutrientInputs.** This process determines the input of N and P to the land segment and to the soil by rainfall, fertilization, and infiltration.

• **PHWTTPMineralization.** This process determines the mineralization of organic P to labile P. This process is the same as PPMineralization with the only difference being the soil moisture response factor.

• **PHWTSubsurfaceTransport.** This process transports labile nutrient forms between soil layers both upwards (by upward flux) and downwards (by percolation).

• **PHWTSurfaceTransport.** This process transports labile nutrient forms in runoff.

• **PMixingZoneExchangeModel.** This process determines and transports nutrients between soil and ponded water.

• **PNutrientTransformationProcess.** This process is an abstract process that is extended by every process that transforms nutrients.

• **PNutrientTransformTransferProcess.** This process is an abstract process that extends PNutrientTransformationProcess and it is extended by every process that transforms and transfers nutrients simultaneously.

• **PNutrientTransportProcess.** This process is an abstract process that is extended by every process that transports nutrients.

**Description of Data Objects**

• **DDenitrificationThreshold.** This DDouble data object holds the value of the denitrification threshold (water content above which denitrification can occur) for each soil layer.

• **DLayerPressureHead.** This DDouble data object holds the value of the pressure head (cm) in each soil layer.
• **DNutrientFluxRecord.** This DDoubleFluxRecord data object holds the flux record which includes the current storage and all of the methods to transport or transform nutrients.

• **DPondedAmmoniumNConc.** This DDailyData data object holds the concentration of ammonium N (mg/L) in ponded water.

• **DPondedLabilePConc.** This DDailyData data object holds the concentration of labile P (mg/L) in ponded water.

• **DPondedNitrateNConc.** This DDailyData data object holds the concentration of nitrate N (mg/L) in ponded water.
APPENDIX F
NUTRIENT PROCESS AND DATA OBJECT UNIFIED MODELING LANGUAGE (UML) DIAGRAMS

**Figure F-1.** PDetermineLayerPressureHeads UML diagram
+cycleNutrient() : void
+runProcess() : void

PNutrientTransformationProcess

+calcAmmonification() : void
+cycleNutrient() : void
-setRequiredData() : void
+calcAmmonificationByLayer() : double
+calcSoilNAmmonificationByLayer() : void

PHWTAmmonification

Data Objects Required:

CPlantResidueLayer:
DPlantResidue
DPlantResidueInitial
DAnimalWasteOMInitial
DAnimalWasteDecayRate
DAnimalWasteOrganicMatter
DAnimalWasteDecayOrganicMatter
DNitrateNFluxRecord
DAmmoniumNFluxRecord
DAnimalWasteOrganicNFluxRecord
DAnimalWasteDecayOrganicNFluxRecord
DLabilePFluxRecord
DPlantResidueOrganicPFluxRecord
DAnimalWasteOrganicPFluxRecord
DAnimalDefecationOrganicPFluxRecord
DPlantResidueNFluxRecord
DPlantResiduePFluxRecord
DAnimalWasteOrganicPFluxRecord
DAnimalDefecationOrganicPFluxRecord

CSoilSurfaceLayer:
DActiveNFrac
DDepth
DResidualMoistureContent
DPorosity
DLayerPressureHead
DMeanSoilTemp
DPlantResidue
DPlantResidueInitial
DAnimalWasteOMInitial
DAnimalWasteDecayRate
DAnimalWasteOrganicMatter
DAnimalWasteDecayOrganicMatter
DNitrateNFluxRecord
DAmmoniumNFluxRecord
DAnimalWasteOrganicNFluxRecord
DAnimalWasteDecayOrganicNFluxRecord
DLabilePFluxRecord
DPlantResidueOrganicPFluxRecord
DAnimalWasteOrganicPFluxRecord
DAnimalDefecationOrganicPFluxRecord

CHorizon:
DActiveNFrac
DDepth
DResidualMoistureContent
DPorosity
DLayerPressureHead
DMeanSoilTemp
DPlantResidue
DPlantResidueInitial
DAnimalWasteOMInitial
DAnimalWasteDecayRate
DAnimalWasteOrganicMatter
DAnimalWasteDecayOrganicMatter
DNitrateNFluxRecord
DAmmoniumNFluxRecord
DAnimalWasteOrganicNFluxRecord
DAnimalWasteDecayOrganicNFluxRecord
DLabilePFluxRecord
DPlantResidueOrganicPFluxRecord
DAnimalWasteOrganicPFluxRecord
DAnimalDefecationOrganicPFluxRecord

Figure F-2. PHWTAmmonification UML diagram
Figure F-3. PHWTDenitrification UML diagram
Figure F-4. PHWTImmobilization UML diagram
Figure F-5. PHWTNitrification UML diagram
Figure F-6. PHWTNutrientInputs UML diagram
Figure F-7. PHWTPMineralization UML diagram
Figure F-8. PHWTSubsurfaceTransport UML diagram
Figure F-9. PHWTSurfaceTransport UML diagram
Figure F-10. PMixingZoneExchangeModel UML diagram
Figure F-11. PNutrientTransformationProcess UML diagram

Figure F-12. PNutrientTransfromTransferProcess UML diagram
Figure F-13. PNutrientTransportProcess UML diagram
Figure F-14. DNutrientFluxRecord UML diagram
APPENDIX G
NUTRIENT MODEL TECHNICAL MANUAL

Introduction

This manual describes the technical details of the nitrogen and phosphorus cycling algorithms used with the High Water Table simulation of the ACRU model developed at the University of Florida. The algorithms used are an adaption of those used in the GLEAMS model (Knisel et al. 1993). Only the appropriate changes are mentioned here. The model employs a simple accounting procedure to route nutrients between soil layers with the soil water. Transport occurs both upwards and downwards in response to soil evaporation, upward flux, infiltration, and percolation. The exchange of nutrients between soil and runoff/ponded water occurs by assuming a constant depth to which water is completely mixed. The effects of soil wetness on nutrient transformations is also outlined. Parameters which are input or output variables are noted in italicized capital letters in parenthesis. In order to use the nutrient simulation option the switch variables (NUTRI) and (UFHWT) must be on.

Nutrient Inputs

Nutrients may enter the system in rainfall, fertilizer, and animal wastes. Details of the various methods are outlined in Knisel et al. (1993).

Soil Moisture Effects on Nutrient Transformations

The nutrient transformations in the GLEAMS model are generally simulated as first order (or zero order in the case of nitrification) processes. The main components governing the rate of transformations are a maximum rate (sometimes constant and
sometimes a function of the rate of plant residue decomposition and C:N and C:P ratios, etc.) and factors that define the effect of temperature and soil moisture (these factors usually vary from 0 to 1). In general, the transformation rate coefficients take the form:

\[ k = k_{\text{Max}} \sqrt{f_{T} f_{\theta}} \]  

(G-1)

and in the case of nitrification:

\[ k = k_{\text{Max}} f_{T} f_{\theta} \]  

(G-2)

where \( k \) is the transformation rate coefficient (day\(^{-1} \)), \( k_{\text{Max}} \) is the maximum rate of transformation under optimal temperature and soil moisture conditions (day\(^{-1} \)), and \( f_{T} \) and \( f_{\theta} \) are the temperature and soil moisture response factors, respectively.

The temperature and soil moisture response factors, \( f_{T} \) and \( f_{\theta} \), are often determined empirically and can differ for each transformation. The use of soil moisture response factors in simulation models are inherently approximate as the response of microbial activity to soil moisture conditions is a function of a number of factors. The relationship between soil moisture and microbial activity has been shown to vary between soils, depending on the shape of the soil moisture curve, the abundance of organic matter, pH, and depth (Goncalves and Carlyle 1994; Rodrigo et al. 1997; Leiros et al. 1999). For the inclusion of the GLEAMS nutrient transformations in the ACRU high water table model it was determined that the response to soil moisture as simulated in GLEAMS was inadequate as GLEAMS was developed as an “upland” model where processes such as decomposition of organic matter is assumed to cease completely when the soil moisture is just above the field capacity of the soil.

There are several mechanisms that cause a decrease in microbial activity in dry soil. These include reduced mobility of both soluble substrate and microbes, and a direct effect
of dryness on microbial growth and survival. Under low soil moisture conditions a reduced rate of decomposition is caused by two factors; first, as the pores within the soil dry and water film coating the sediment surfaces becomes thinner, diffusion path lengths become more tortuous and the rate of both substrate and microbe diffusion declines, second, low water contents correspond to low water potentials that lower intracellular water potentials which in turn reduce hydration and enzymatic activity (Porporato et al. 2003).

Under wet conditions a decrease in aerobic microbial activity is caused by a reduction of oxygen diffusion (Grant and Rochette 1994). During periods of high soil moisture anoxic conditions prevent bacteria from aerobically oxidizing organic matter (decomposition).

In the GLEAMS model, there are three soil moisture functions employed for various processes. For ammonification (decomposition of organic nitrogen), phosphorus mineralization, and mineral nitrogen and phosphorus immobilization (the uptake of N and P by soil microbes when substrate is nutrient poor, generally a C:N ratio greater than 25 or a C:P ratio greater than 200) the soil moisture response function is of the form (shown in figure G-1):

\[
f_{o1} = \begin{cases} 
\frac{\theta - \theta_{wp}}{\theta_{fc} - \theta_{wp}} & \text{for } \theta \leq \theta_{fc} \\
0 & \text{for } \theta > \theta_{fc} 
\end{cases}
\]  

\hspace{1cm} (G-3a)

\hspace{1cm} (G-3b)

where \( \theta \), \( \theta_{wp} \), and \( \theta_{fc} \) are the moisture content of a soil layer, the moisture content at the wilting point, and the moisture content at field capacity, respectively.
As can be seen in figure G-1, the response to soil moisture rises from zero at the wilting point to an optimum value at field capacity. Immediately above field capacity the response reduces to zero, implying a complete cessation of microbial activity in the decomposition process.

For nitrification the soil moisture response function in the GLEAMS model is at an optimum value at field capacity and decreases linearly to zero at saturation and at the wilting point (figure G-1):

\[ f_{\theta n} = \frac{\theta - \theta_{wp}}{\theta_{fc} - \theta_{wp}} \quad \text{for} \quad \theta \leq \theta_{fc} \quad (G-4a) \]

\[ f_{\theta n} = 1 - \frac{\theta - \theta_{fc}}{\theta_{s} - \theta_{fc}} \quad \text{for} \quad \theta_{fc} < \theta < \theta_{s} \quad (G-4b) \]

where \( \theta_{s} \) is the water content at saturation.

For denitrification the soil moisture response function in GLEAMS begins when the water content is 10% above field capacity and increases linearly to saturation (figure G-1):

\[ f_{\theta d} = \frac{\theta - \left[ \theta_{fc} + 0.1(\theta_{s} - \theta_{fc}) \right]}{\theta_{s} - \left[ \theta_{fc} + 0.1(\theta_{s} - \theta_{fc}) \right]} \quad \text{for} \quad \theta \geq \theta_{fc} + 0.1(\theta_{s} - \theta_{fc}) \quad (G-5a) \]

\[ f_{\theta d} = 0 \quad \text{for} \quad \theta < \theta_{fc} + 0.1(\theta_{s} - \theta_{fc}) \quad (G-5b) \]

In a comparison of nine nitrogen simulation models Rodrigo et al. (1997) has shown significant differences in their response to both soil moisture and temperature. However Rodrigo et al. (1997) note that defining functions in terms of water pressure allows for comparison between soils of different textures, using soil water contents may be more useful in describing processes that can limit microbial activity in soils such as solute and oxygen diffusion, while expressing functions in terms of water filled pore
space, or relative saturation, appears to be the best indicator of aerobic/anaerobic microbial activity.

Figure G-1. Soil moisture response function of GLEAMS. The response of P mineralization, exchange of labile and stable P, and N and P immobilization to soil moisture follow the same curve as ammonification.

Previous studies have reported the highest rates of mineralization occurring near the field capacity of the soil, decreasing as the soil dries. The optimal soil moisture conditions that yield optimal decomposition and mineralization rates has been reported to occur at soil water pressure heads between 100 and 500 cm (Rodrigo et al. 1997). Contradictory results have been measured between the range of field capacity and saturation and there is no consensus on the moisture content or pressure at which microbial activity stops (Rodrigo et al. 1997). However the general form of the response functions to both temperature and soil moisture are often broadly similar in many models but the soil moisture functions often differ at the point and rate at which they decline from the optimum value (Wu and McGechan 1998). Differences in soil moisture
response functions between models can be the one of the main causes for differences in prediction (Rodrigo et al. 1997).

In order to better represent the response of transformation processes on soil moisture a range of soil moisture contents under which maximum transformation rates occur is desired. In order to do this soil moisture response functions for ammonification (and the other processes that use the same response) and nitrification are described as logarithmic functions of soil water pressure head as is done in several other models (Hansen et al. 1991; Rijetma and Kroes 1991; Vanclooster et al. 1996). Sommers et al. (1980) and Kladivko and Keeney (1987) have shown that mineralization rates of nitrogen could be well represented as a linear function of water content or a logarithmic function of soil water pressure head. The soil moisture response of denitrification is simulated as a function of relative saturation as proposed by Johnsson et al. (1987) and used by Vanclooster et al. (1996).

The soil moisture response function of ammonification, phosphorus mineralization, and mineral nitrogen and phosphorus immobilization used here is expressed in units of \( pF \) (\( \log_{10} \) of negative pressure head in units of cm) and is illustrated in figure G-2:

\[
f_{\theta i} = \frac{pF_{wp} - pF}{pF_{wp} - 2.7} \quad \text{for } pF > 2.7 \quad (G-6a)
\]

\[
f_{\theta i} = 0.6 + 0.4 \frac{pF - pF_s}{2 - pF_s} \quad \text{for } pF < 2 \quad (G-6b)
\]

\[
f_{\theta i} = 1 \quad \text{for } 2 \leq pF \leq 2.7 \quad (G-6c)
\]

where \( pF_{wp} \) is the \( pF \) value at the wilting point (15000 cm), \( pF_s \) is the \( pF \) near saturation (taken as 1 cm for mathematical reasons), and the optimal soil water response occurs...
between a $pF$ of 2 and 2.7 which correspond to 100 cm and 500 cm of soil water pressure head, respectively.

Figure G-2. Soil moisture response functions used. The response of P mineralization, exchange of labile and stable P, and N and P immobilization to soil moisture follow the same curve as ammonification.

The optimum soil moisture conditions of nitrification are similar to that of ammonification (decomposition of organic matter) with the exception that nitrification tends to zero under saturated conditions (Linn and Doran 1984; Skopp et al. 1990). For nitrification the soil moisture response function is of the form (figure G-2):

\[
f_{\text{en}} = \frac{pF_{wp} - pF}{pF_{wp} - 2.7} \quad \text{for } pF > 2.7 \quad \text{(G-7a)}
\]

\[
f_{\text{en}} = \frac{pF - pF_s}{2 - pF_s} \quad \text{for } pF < 2 \quad \text{(G-7b)}
\]

\[
f_{\text{en}} = 1 \quad \text{for } 2 \leq pF \leq 2.7 \quad \text{(G-7c)}
\]
The response of denitrification to soil wetness is defined using the form proposed by Johnsson et al. (1987) and similar to that developed by Rolston et al. (1984) and is shown in figure G-3:

\[
f_{\text{den}} = \left[ \frac{\theta - \theta_d}{\theta_s - \theta_d} \right]^d
\]

where \(\theta\) and \(\theta_s\) are the water content and saturated water content, respectively, \(\theta_d\) is a threshold water content which defines the water content above which denitrification occurs and is assumed to correspond to an effective saturation of 0.8, and \(d\) is an empirical exponent assumed to be equal to 2 (Vanclooster et al. 1996).

Figure G-3. Soil moisture response function to denitrification.

A comparison of the soil moisture response functions of GLEAMS and those proposed here can be seen in figure G-4 by applying them to a fictitious soil. It should be noted that there is considerable uncertainty in the soil moisture response functions proposed here, as there is with all soil moisture response functions. In particular, the
limits of the optimal ranges for ammonification and nitrification as well as the point at which denitrification begins to occur should be reevaluated for specific applications of the model.

Figure G-4. Comparison of the soil moisture response functions of GLEAMS and those proposed. Using a soil where \( \theta_s = 0.38 \), \( \theta_r = 0.05 \), \( \alpha = 0.02 \) cm\(^{-1}\), and \( n = 3 \) using the soil moisture model of van Genuchten (1980). Field capacity and wilting point are assumed to occur at 15,000 and 300 cm, respectively.

**Nutrient Transport in the Subsurface**

**Transport by Infiltration and Redistribution**

In order to determine the mass of nutrients that infiltrate on a given day, the volume of water that had been present at the ground surface is determined as:

\[
H_{\text{Max}} = H_{\text{End}} + E_{\text{pond}} + I + RO
\]  

(G-9)

where \( H_{\text{Max}} \) is the maximum amount of water that has been on the ground surface during the day (mm), \( H_{\text{End}} \) \( (\text{D}P\text{OND}) \) is the depth of ponded water at the end of the day (mm)
after all water movements have occurred, \( E_{pond} \) is the amount of evaporation that has occurred from ponded water (mm), \( I \) is the depth of water infiltrated (mm), and \( RO \) (mm) is the depth of runoff or quickflow (\( QUICKF \)). If ponded water is present at the end of the day (\( H_{end} > 0 \)) then the mass of nutrients that infiltrate \( M_{Infil} \) (kg/ha) is determined according to the ratio of the maximum depth of water at the ground surface and the depth of infiltrated water:

\[
M_{Infil} = \frac{I}{H_{Max}} M_{GS}
\]  

where \( M_{GS} \) is the mass of nutrient present at the ground surface (kg/ha). If no ponded water was present at the end of the day and rainfall occurred then the entire mass of nutrient at the ground surface is assumed to infiltrate.

Downward movement of the nutrient in the soil profile occurs according to the net percolation, \( q_{net} \) (cm):

\[
q_{net} = q_{Perc} - q_{UpwardFlux}
\]

where \( q_{Perc} \) (cm) is the amount of water that has moved downward out of the soil layer and \( q_{UpwardFlux} \) is the amount of water that has moved upward into the layer from below. The concentration in percolating water (\( C_{Perc} \)) is determined for nitrate:

\[
C_{Perc} = 10 \frac{M_{Layer}}{\theta \cdot d + q_{net} + q_{UpwardFluxOut}}
\]

where \( M_{Layer} \) is the mass of nutrient contained in a soil layer (kg/ha), \( \theta \) is the average water content of the layer (cm\(^3\) cm\(^{-3}\)), \( d \) is the thickness of the layer (cm), \( q_{UpwardFluxOut} \) is the amount of water that has moved upward out of the soil layer, and the value 10 is a unit conversion (mg ha cm kg\(^{-1}\) L\(^{-1}\)). For ammonium and labile phosphorus the concentrations are determined from:
\[ C_{\text{perc}} = 10 \frac{M_{\text{Layer}}}{K_d M_{\text{soil}} + \theta \cdot d + q_{\text{net}} + q_{\text{UpwardFluxOut}}} \]  

(G-13)

where \( K_d \) is the partitioning coefficient and \( M_{\text{soil}} \) is the soil mass (Mg/ha). The partitioning coefficient for ammonium is related to the clay content of the soil (CL in units of \%):

\[ K_d = 1.34 + 0.083 \cdot CL \]  

(G-14)

and the partitioning coefficient of labile phosphorus can be related to parameters such as the oxalate extractable aluminum and double acid extractable magnesium content of the soil. In the GLEAMS model it is assumed to be a function solely of the clay content:

\[ K_d = 100 + 2.5 \cdot CL \]  

(G-15)

The mass of nutrient that moves with the net percolation (\( M_{\text{perc}} \)) is calculated as:

\[ M_{\text{perc}} = 0.1 \cdot q_{\text{net}} C_{\text{perc}} \]  

(G-16)

where \( M_{\text{perc}} \) is in units of kg/ha and the value 0.1 is a unit conversion (kg L mg\(^{-1}\) cm\(^{-1}\) ha\(^{-1}\)). The mass of nutrient that moves upward out of the layer is:

\[ M_{\text{UpwardFlux}} = 0.1 \cdot q_{\text{UpwardFluxOut}} C_{\text{perc}} \]  

(G-17)

**Transport by Evaporation**

Upward migration caused by soil evaporation (not transpiration) is approximated by allowing the nutrients to move up a single layer. The mass that moves is determined from:

\[ M_{\text{Evap}} = 0.01 \cdot E_{\text{Layer}} C_{\text{perc}} \]  

(G-18)

where \( M_{\text{Evap}} \) is in units of kg/ha, \( E_{\text{Layer}} \) is the soil evaporation occurring from the layer and the value 0.01 is a unit conversion (kg L mg\(^{-1}\) mm\(^{-1}\) ha\(^{-1}\)).
Transport between Ponded and Soil Water

During runoff events water entrains some of the soil porewater, this entrainment was referred to as a process of accelerated diffusion by Ahuja and Lehman (1983) when infiltration is absent. The “extraction” of porewater solutes has been shown to occur principally near the soil surface and rapidly diminishing with depth (Ahuja et al. 1981). The exchange of solutes between the soil and ponded or runoff water has been simulated as a convective mass transfer or enhanced diffusion process as done by Parr et al. (1987), Wallach et al. (1989), Ahuja (1990), and Havis et al. (1992). However these models, both numerical and analytical, have to date only been applied on single controlled laboratory events and have yet to be integrated into a continuous model.

In works such as that of Wallach et al. (1989), Kesseler (1999), Boudreau (1997), and Boudreau and Jorgensen (2001) the mass transfer coefficient can be related to the diffusion coefficient and the thickness of a thin boundary layer, or diffusive sublayer. It is assumed that there is a thin film of water above the soil surface that is stagnant, through which chemical transport occurs by diffusion only. When runoff is flowing over this film the transport is enhanced. This thin film is assumed to contain all of the resistance to mass transfer (Kessler 1999).

In lieu of such a complex approach several approximate models have been adopted. Early modeling efforts assumed that soil water within a thin zone of surface soil mixes completely and instantaneously with runoff (Crawford and Donigian 1973; Steenhuis and Walter 1980). Some other modeling efforts considered a surface soil zone that mixes incompletely with runoff with the degree of mixing determined by an empirical extraction coefficient (Frere et al. 1980; Leonard et al. 1987). Ahuja and Lehman (1983)
suggested that the degree of soil water mixing should be an exponential function of depth, however with empirical parameters that must be known a-priori.

Since the depth of the soil that interacts with runoff has been shown to increase with reduced infiltration during the runoff event the thickness of the soil surface layer is allowed to vary and is a calibratable parameter. Extraction coefficients as determined in the GLEAMS model are retained.

Exchange of nutrients between the soil and ponded or runoff water is simulated by assuming a depth within the soil in which surface and soil water are completely mixed. The depth of ponded water that interacts with soil water is:

\[ H_{\text{Max}} = H_{\text{End}} + RO \]  \hspace{1cm} (G-19)

The top 1 cm of soil is assumed to mix with the ponded water. This portion of the soil is assumed to be incompletely mixed. The degree of mixing is defined by an extraction coefficient \( \beta \) as is done in the GLEAMS model, and ranges between 0.1 and 0.5 as a function of the partition coefficient of a particular nutrient (Knisel et al. 1993). Operationally, this extraction coefficient can be taken as being equivalent to the fraction of the top 1 cm layer that is completely mixed with the ponded water. The amount of water within this completely mixed soil zone, \( d_{\text{mixed}} \), is:

\[ d_{\text{mixed}} = \phi d_{ss} \beta \]  \hspace{1cm} (G-20)

where \( \phi \) is the porosity of the top, or soil surface, layer, and \( d_{ss} \) is the depth of the soil surface layer (1 cm).

Under ponded conditions, soluble nutrients within the plant residue layer are assumed to mix completely with the ponded water. Upon complete mixing between the
ponded water and the portion of the soil an equilibrium concentration of solute can be defined as (mg/L):

\[ C_{eq} = \frac{C_{ss}d_{mixed} + C_{w}H_{Max}}{d_{mixed} + H_{Max}} \tag{G-21} \]

where \( C_{ss} \) and \( C_{w} \) are the concentration of solute in the surface soil layer and ponded water (mg/L), respectively, and \( H_{Max} \) is expressed in units of cm.

Mass of solute is transported in the appropriate direction (upward out of the soil, or downward into the soil) in order to attain this equilibrium concentration.

**Surface Transport**

The transport of nutrients in runoff water is determined to be proportional to the fraction of ponded water that runs off on a given day:

\[ M_{runoff} = 0.01 \cdot RO \cdot C_{w} \tag{G-22} \]

where \( M_{runoff} \) is the mass of nutrient carried in runoff water, \( RO \) is the depth of runoff (mm), and the value 0.01 converts units.
APPENDIX H
CONSERVATIVE SOLUTE TRANSPORT PROCESS, INTERFACE, AND DATA OBJECTS AND DESCRIPTIONS

Process Objects

• PConservativeMixingZoneExchangeModel
• PConservativeSoluteEvaporationTransport
• PConservativeSoluteInputs
• PConservativeSoluteSubsurfaceTransport
• PConservativeSoluteSurfaceTransport
• PConservativeSoluteTransportProcess

Interface Objects

• IConservativeSoluteFlow

Data Objects

• DConservativeSoluteFluxRecord
• DConservativeSoluteOption
• DLeachateConservativeSoluteConc
• DLeachateConservativeSoluteLoad
• DPondedConservativeSoluteConc
• DPondedWaterEvaporation
• DRainfallSoluteConc
• DRainfallSolute
• DRunoffConservativeSoluteConc
• DRunoffConservativeSoluteLoad
• DSoluteApplicationMethod
• DSoluteAmountApplied

Description of Process Objects

• **PConservativeMixingZoneExchangeModel.** This process determines and transports solute between soil and ponded water.

• **PConservativeSoluteEvaporationTransport.** This process transports solute upwards in response to evaporation from the soil.
• **PConservativeSoluteInputs.** This process determines the input of solute to the land segment and to the soil by rainfall, land surface application, and infiltration.

• **PConservativeSoluteSubsurfaceTransport.** This process transports solute between soil layers both upwards (by upward flux) and downwards (by percolation).

• **PConservativeSoluteSurfaceTransport.** This process transports solute forms in runoff.

• **PConservativeSoluteTransportProcess.** This process is an abstract process that is extended by every process that transports solute.

**Description of Interface Objects**

• **IConservativeSoluteFlow.** This interface is implemented by PConservativeSoluteTransportProcess.

**Description of Data Objects**

• **DConservativeSoluteFluxRecord.** This DDoubleFluxRecord data object holds the flux record which includes the current storage and all of the methods to transport solute.

• **DConservativeSoluteOption.** This DInteger data object determines whether conservative solute transport is simulated.

• **DLeachateConservativeSoluteConc.** This DDailyData data object holds the concentration of leaching solute from a soil layer.

• **DLeachateConservativeSoluteLoad.** This DDailyData data object holds the load of leaching solute from a soil layer.

• **DPondedWaterEvaporation.** This DDailyData data object holds the amount of water that has been evaporated from the surface.

• **DRainfallSoluteConc.** This DDailyData data object holds the concentration of solute in rain.

• **DRainfallSolute.** This DDailyData data object holds the mass of solute in rain.

• **DRunoffConservativeSoluteConc.** This DDailyData data object holds the concentration of solute in runoff.

• **DLeachateConservativeSoluteLoad.** This DDailyData data object holds the load of solute in runoff.
• **DSoluteAmountApplied.** This DDailyData data object holds the mass of solute applied to the land surface.

• **DSoluteApplicationMethod.** This DDailyInteger data object holds the method of application of applied solute. Currently only one method is supported, the application to the ground surface.
APPENDIX I
CONSERVATIVE SOLUTE TRANSPORT PROCESS AND DATA OBJECT
UNIFIED MODELING LANGUAGE (UML) DIAGRAMS

PConservativeSoluteTransportProcess

+flowConservativeSolute() : void
+runProcess() : void

PConservativeMixingZoneExchangeModel

+calcExchange() : void
+flowConservativeSolute() : void
-setRequiredData() : void

CLandSegment

CSoilLayer

Data Objects Required:
CLandSegment:
DConservativeSoluteFluxRecord
DPondedConservativeSoluteConc
DWaterFluxRecord

CSoilLayer:
DDepth
DPorosity
DConservativeSoluteFluxRecord

Figure I-1. PConservativeMixingZoneExchangeModel UML diagram
Figure I-2. PConservativeSoluteEvaporationTransport UML diagram
Figure I-3. PConservativeSoluteInputs UML diagram
Figure I-4. PConservativeSoluteSubsurfaceTransport UML diagram
Figure I-5. PConservativeSoluteSurfaceTransport UML diagram

Figure I-6. PConservativeSoluteTransportProcess UML diagram
**Figure I-7. DConservativeSoluteFluxRecord UML diagram**
APPENDIX J
CONSERVATIVE SOLUTE TRANSPORT INPUT/OUTPUT VARIABLE REFERENCE

Input Variable Reference

• **CONSERV**: Option to specify the simulation of a conservative solute. CONSERV = NO(0) No solute simulated; = YES(1) Solute simulated

• **RAINSOL**: Mass of solute applied in rainfall (kg/ha).

• **RSOLCONC**: Concentration of solute in rainfall (mg/L).

• **SOLAMT**: Mass of solute applied on the land surface (kg/ha).

• **SOLMTD**: Solute application method.

• **SOLMTD**: No solute applied SOLMTD = 1 Solute applied to land surface.

Output Variable Reference
APPENDIX K
CONSERVATIVE SOLUTE TRANSPORT TECHNICAL MANUAL

Introduction

This manual describes the technical details of the conservative solute transport used with the High Water Table simulation of the ACRU model developed at the University of Florida. The model employs a simple accounting procedure to route solute between soil layers with the soil water. Transport occurs both upward and downward in response to soil evaporation, upward flux, infiltration, and percolation. The exchange of solute between soil and runoff/ponded water occurs by assuming a constant depth within the soil to which water is completely mixed. Parameters which are input or output variables are noted in italicized capital letters in parenthesis. In order to use this conservative solute option the switch variable (CONSERV) must be on.

Conservative Solute Inputs

Solute may enter the system in rainfall or by surface application. In rainfall the concentration, \( C_{\text{Rain}} \) is entered in units of mg/L (RSOLCONC). The total mass of solute input as rain on a given day is reported in units of kg/ha (RAINSOL). The mass of solute input by rain \( M_{\text{Rain}} \) (RAINSOL) is determined from:

\[
M_{\text{Rain}} = 0.01 \cdot R \cdot C_{\text{Rain}}
\]

(K-1)

where \( R \) is rainfall (mm) and the value 0.01 is a unit conversion (kg L mg\(^{-1}\) ha\(^{-1}\) mm\(^{-1}\)).

The solute application method (SOLMTD) and mass of solute applied (SOLAMT), in units of kg/ha, are input as daily time series. Currently there is only a single application method available, the application to the ground surface. The mass applied is
added to the ground surface of the land segment and may dissolve into water if ponded water is present or rainfall has occurred. Otherwise the solute will remain on the ground surface until a rainfall event occurs at which time it will infiltrate or runoff.

**Conservative Solute Transport in the Subsurface**

**Transport by Infiltration and Redistribution**

In order to determine the mass of solute that infiltrates on a given day, the volume of water that had been present at the ground surface is determined as:

\[
H_{\text{Max}} = H_{\text{End}} + E_{\text{pond}} + I + RO
\]  \hspace{1cm} (K-2)

where \(H_{\text{Max}}\) is the maximum amount of water that has been on the ground surface during the day (mm), \(H_{\text{End}}\) is the depth of ponded water at the end of the day (mm) after all water movements have occurred, \(E_{\text{pond}}\) is the amount of evaporation that has occurred from ponded water (mm), \(I\) is the depth of water infiltrated (mm), and \(RO\) (mm) is the depth of runoff or quickflow (\(\text{QUICKF}\)). If ponded water is present at the end of the day (\(H_{\text{end}} > 0\)) then the mass of solute that infiltrates \(M_{\text{Infil}}\) (kg/ha) is determined according to the ratio of the maximum depth of water at the ground surface and the depth of infiltrated water:

\[
M_{\text{Infil}} = \frac{I}{H_{\text{Max}}} M_{\text{GS}} \hspace{1cm} (K-3)
\]

where \(M_{\text{GS}}\) is the mass of solute present at the ground surface (kg/ha). If no ponded water was present at the end of the day and rainfall occurred then the entire mass of solute at the ground surface is assumed to infiltrate.

Downward movement of the solute in the soil profile occurs according to the net percolation, \(q_{\text{net}}\) (cm):

\[
q_{\text{net}} = q_{\text{Perc}} - q_{\text{UpwardFlux}} \hspace{1cm} (K-4)
\]
where $q_{perc}$ (cm) is the amount of water that has moved downward out of the soil layer and $q_{UpwardFlux}$ is the amount of water that has moved upward into the layer from below.

The concentration in percolating water ($C_{Perc}$) is determined from:

$$C_{Perc} = \frac{10}{\theta \cdot d + q_{net} + q_{UpwardFluxOut}}$$

(K-5)

where $M_{Layer}$ is the mass of solute contained in a soil layer (kg/ha), $\theta$ is the average water content of the layer (cm$^3$ cm$^{-3}$), $d$ is the thickness of the layer (cm), $q_{UpwardFluxOut}$ is the amount of water that has moved upward out of the soil layer, and the value 10 is a unit conversion (mg ha cm kg$^{-1}$ L$^{-1}$). The mass of solute that moves with the net percolation ($M_{Perc}$) is calculated as:

$$M_{Perc} = 0.1 \cdot q_{net} \cdot C_{Perc}$$

(K-6)

where $M_{Perc}$ is in units of kg/ha and the value 0.1 is a unit conversion (kg L mg$^{-1}$ cm$^{-1}$ ha$^{-1}$). The mass of solute that moves upward out of the layer is:

$$M_{UpwardFlux} = 0.1 \cdot q_{UpwardFluxOut} \cdot C_{Perc}$$

(K-7)

**Transport by Evaporation**

Upward migration caused by soil evaporation (not transpiration) is approximated by allowing the solute to move up a single layer. The mass that moves is determined from:

$$M_{Evap} = 0.01 \cdot E_{Layer} \cdot C_{perc}$$

(K-8)

where $M_{Evap}$ is in units of kg/ha, $E_{Layer}$ is the soil evaporation occurring from the layer and the value 0.01 is a unit conversion (kg L mg$^{-1}$ mm$^{-1}$ ha$^{-1}$).
Transport Between Ponded and Soil Water

During runoff events water entrains some of the soil porewater. This entrainment was referred to as a process of accelerated diffusion by Ahuja and Lehman (1983) when infiltration is absent. The “extraction” of porewater solutes has been shown to occur principally near the soil surface and rapidly diminishing with depth (Ahuja et al. 1981). The exchange of solutes between the soil and ponded or runoff water has been simulated as a convective mass transfer or enhanced diffusion process as done by Parr et al. (1987), Wallach et al. (1989), Ahuja (1990), and Havis et al. (1992). However these models, both numerical and analytical, have to date only been applied on single controlled laboratory events and have yet to be integrated into a continuous model.

In works such as that of Wallach et al. (1989), Kesseler (1999), Boudreau (1997), and Boudreau and Jorgensen (2001) the mass transfer coefficient can be related to the diffusion coefficient and the thickness of a thin boundary layer, or diffusive sublayer. It is assumed that there is a thin film of water above the soil surface that is stagnant, through which chemical transport occurs by diffusion only. When runoff is flowing over this film the transport is enhanced. This thin film is assumed to contain all of the resistance to mass transfer (Kessler 1999).

In lieu of such a complex approach several approximate models have been adopted. Early modeling efforts assumed that soil water within a thin zone of surface soil mixes completely and instantaneously with runoff (Crawford and Donigian 1973; Steenhuis and Walter 1980). Some other modeling efforts considered a surface soil zone that mixes incompletely with runoff with the degree of mixing determined by an empirical extraction coefficient (Frere et al. 1980; Leonard et al. 1987). Ahuja and Lehman (1983)
suggested that the degree of soil water mixing should be an exponential function of depth, however with empirical parameters that must be known a-priori.

Since the depth of the soil that interacts with runoff has been shown to increase with reduced infiltration during the runoff event the thickness of the soil surface layer is allowed to vary and is a calibratable parameter. The extraction coefficient for a conservative solute is assumed to be 0.5 as is done with nitrate in the GLEAMS model.

Exchange of solute between the soil and ponded or runoff water is simulated by assuming a depth within the soil in which surface and soil water are completely mixed. The depth of ponded water that interacts with soil water is:

\[ H_{Max} = H_{End} + RO \]

(K-9)

The top 1 cm of soil is assumed to mix with the ponded water. This portion of the soil is assumed to be incompletely mixed. The degree of mixing is defined by an extraction coefficient (\( \beta \)) as is done in the GLEAMS model. Operationally, this extraction coefficient can be taken as being equivalent to the fraction of the top 1 cm layer that is completely mixed with the ponded water. The amount of water within this completely mixed soil zone, \( d_{mixed} \), is:

\[ d_{mixed} = \phi d_{ss} \beta \]

(K-10)

where \( \phi \) is the porosity of the top, or soil surface, layer, and \( d_{ss} \) is the depth of the soil surface layer (1 cm).

Under ponded conditions, any solute within the plant residue layer is assumed to mix completely with the ponded water. Upon complete mixing between the ponded water and the portion of the soil an equilibrium concentration of solute can be defined as (mg/L):
\[
C_{eq} = \frac{C_{ss} d_{mixed} + C_w H_{Max}}{d_{mixed} + H_{Max}}
\]  

(K-11)

where \(C_{ss}\) and \(C_w\) are the concentration of solute in the surface soil layer and ponded water (mg/L), respectively, and \(H_{Max}\) is expressed in units of cm.

Mass of solute is transported in the appropriate direction (upward out of the soil, or downward into the soil) in order to attain this equilibrium concentration.

**Surface Transport**

The transport of solute in runoff water is determined to be proportional to the fraction of ponded water that runs off on a given day:

\[
M_{\text{Runoff}} = 0.01 \cdot RO \cdot C_w
\]  

(K-12)

where \(M_{\text{Runoff}}\) is the mass of solute carried in runoff water, \(RO\) is the depth of runoff (mm), and the value 0.01 converts units.
LIST OF REFERENCES


Horton, R.E. 1933. The role of infiltration in the hydrologic cycle. Transactions of the American Geophysical Union 14: 446-460.


BIOGRAPHICAL SKETCH

Christopher John Martinez was born in Thousand Oaks, California, in 1973. He graduated from the Richard Stockton College of New Jersey in 1996, with a B.S. from the Department of Environmental Studies. Shortly thereafter he moved to Florida and began graduate studies in the Department of Environmental Engineering Sciences at the University of Florida in 1998. There he studied the hydrology and hydraulics of constructed treatment wetlands for wastewater treatment and received his M.E. degree in 2001. He continued his education at the Department of Environmental Engineering Sciences at the University of Florida as a U.S.D.A. National Needs Fellow.