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Integrating Surface and Sub Surface Flow Models of Different Spatial and Temporal Scales Using Potential Coupling Interfaces

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INTEGRATING SURFACE AND SUB SURFACE FLOW MODELS

OF DIFFERENT SPATIAL AND TEMPORAL SCALES

USING POTENTIAL COUPLING INTERFACES

by

Alphonce C. Guzha

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ABSTRACT

Integrating Surface and Subsurface Flow Models of Different Spatial and Temporal Scales Using Potential Coupling Interfaces

by

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Utah State University, 2008

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The main objective of this research was to develop and utilize a coupled surface water groundwater model to simulate hydrological responses of watersheds. This was achieved by coupling the U.S. Geological Survey (USGS) groundwater flow model, MODFLOW, and the rainfall runoff model, TOPMODEL, in one case study and coupling MODFLOW with a networked version of TOPMODEL called TOPNET in another case study. The model coupling was achieved using the InCouple approach, which utilizes Potential Coupling Interfaces (PCIs) that are abstractions from model flow diagrams that expose only those aspects of a model relevant to coupling. Coupling the rainfall-runoff models to MODFLOW involved development of a routine relating the spatial discretization of MODFLOW to TOPMODEL and similarly MODFLOW to TOPNET and development of a feedback scheme where groundwater and surface water interact in the soil zone.

The key coupling concept was replacing the wetness index-based depth-to-water table concept of TOPMODEL with the groundwater heads simulated by MODFLOW. In the MODFLOW–TOPMODEL coupling, using data for the Tenmile Creek watershed, for the period, 1968 to 1972, it was concluded that the coupled model was able to continuously simulate the stream flow. However, the coupled model under predicted stream flow and did not agree well with observations in a point wise comparison. A mean coefficient of efficiency of 0.54 was obtained between simulated and measured stream flow. Only 24% of received precipitation was observed as baseflow and this shows that there is limited interaction between surface water and groundwater in the watershed. It was demonstrated using the coupled model that the lateral flow processes and the interactions between groundwater and surface water have a major importance for the water balance. For the Big Darby watershed, for the period 1992 to 2000, the coupled model adequately predicts the stream and groundwater flow distribution in the watershed.

After model calibration, simulated groundwater showed the greatest residual variance, attributed to model error and uncertainty in model parameters. Model fit efficiencies of 0.61 and 0.69 were obtained for simulating stream flow measured at two gaging stations. The overall watershed hydrologic budget also showed small mass balance errors using the coupled model. However, the study also shows the need for further research in regard to constraining the groundwater recharge parameter which links the models.

(241 pages)

DEDICATION

This dissertation is dedicated with so much love to the memory of my wonderful sister, Antonetta “Tonette” Kudzai. You bore tragedy and adversity with grace and undiminished faith. Once you said to me, “*The standard of excellence you achieve will never be an accident but it will always be a high intention of your most sincere effort, intelligent direction and skilled execution.*” I gave it my best shot, Sis. I know you are in a better place now and I know too well that you are celebrating with me. I miss you!

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Alphonce Chenjerayi Guzha

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CHAPTER 1

INTRODUCTION AND OBJECTIVES

Background and Motivation

As a result of population increases, water use for agricultural, municipal and industrial purposes has been rapidly increasing in the last few decades. Sustainable management and protection of water resources has become a key issue not only for the United States, but the entire world in the 21st century. As stated by Huyakorn (2000), effective management of watersheds requires a comprehensive knowledge of hydrologic processes, and impacts of point source and non-point source pollution on water quality. This has led to the advent of simulation models are being used increasingly to provide predictive capability in support of environmental and water resource assessment and restoration projects. However, the models used are often based on simplifications to complex hydrologic and transport processes, often without consideration of the entire hydrological flow of water from surface to subsurface systems. Such models incorporate restrictive assumptions pertaining to spatial variability, dimensionality and interaction of various components of flow and transport processes.

The United States Geological Survey (1998) department noted in its Strategic directions for the Water Resources Division report for the period 1998 to 2008 that there is an increasing research interest in the interactions between surface and groundwater systems. The main reason for this increased interest, especially in simulation and modeling research, is that it has become clear that the simulation of surface water flow of an area is not completed until the effects of the aquifers underneath the area are taken into consideration and the same can be said for the simulation of groundwater flow.

However, as stated by Nemeth, Wilcox and Solo-Gabriele (2000), linking groundwater and surface water models to each other is frequently problematic because the two models use different sets of governing equations. Additionally, the time scale of interest is usually longer for groundwater modeling than for surface water modeling. Linking surface and groundwater modeling provides a tool for a complete description of the spatio-temporal variability and organization of the underlying hydrologic processes (e.g. infiltration, surface runoff, deep percolation, evapotranspiration and groundwater flow). As noted by Winter et al. (1999), “Recent experience with resolution of difficult water management and allocation problems has shown that a capability to simulate the characteristics of the hydrologic system, at watershed scale, is critical.”

Considerable effort has been expended to characterize the physical, chemical, and biological processes affecting groundwater and surface water resources in river basins. This is because it has become apparent in hydrological studies that processes must inevitably be perceived in an integrated way. Many of the impacts of land use changes on surface water systems cannot be evaluated meaningfully without considering the dynamics in subsurface flow systems. As the development of fully integrated holistic model concepts for this purpose is still in its early stages, one means of integration is the coupling of existing disciplinary models. However, this has its own problems because disciplinary models were usually originally designed to solve specific problems in different domains of the water cycle. The processes and the process descriptions they include and the extent of their domain of interest was adapted to a typical class of problems (Barthel, 2006). Therefore the coupling of two or more disciplinary models can result in conceptual inconsistencies and incompatibilities as a result of the individual models which may describe the same process differently or ignore some important linking processes. There may also be overlaps and, in some cases, gaps between the model domains.

This research was inspired by this need to improve tools for simulating interactions between groundwater and surface water to quantify the effects of human activity and natural phenomena on watershed hydrological responses. The research developed and used an intermediate model coupling tool to link established subsurface water and surface water modeling systems. Managing and regulating water use in watersheds and aquifer systems can be aided by an understanding of surface–subsurface water interactions and overall annual hydrologic cycle dynamics as a result of these interactions.

The research was carried out using two case studies. The first case study involved coupling TOPMODEL and MODFLOW with application to Tenmile watershed, a 35 square mile watershed in the Lowlands of Water Resources Inventory Area 1 (WRIA1) in Washington State, USA. This watershed was selected to test the applicability of using Potential Coupling Interfaces as a model coupling tool. The watershed is a lowland watershed and thus it is an area where ground water surface water interactions are expected to play a leading role in the watershed hydrologic balance. The second case study involved coupling TOPNET and MODFLOW with application to the Big Darby watershed in Central Ohio. The watershed spans over six different counties and covers a total area of approximately 550 square miles. It is a watershed in which land use is rapidly changing from being predominantly an agricultural watershed with increases in urbanization especially the expansion of the city of Columbus on the eastern edge of the watershed. Thus it is a good case study in which to evaluate the effectiveness of the coupled model in evaluating the influence of land use changes on hydrologic balance of the watershed. Another reason for choosing this watershed was the availability of hydrologic data especially time series of ground water head data for model calibration which was often lacking in several watersheds originally considered for this research.

Research Objectives

The study was designed to address the issues outlined in the problem statement by meeting the following objectives. The general objective was is to develop a coupled simulation model that integrates a surface water flow model (TOPMODEL and TOPNET) to a sub surface hydrologic systems model (MODFLOW) to simulate flow over large space and time scales in a river basin.

The specific objectives are:

1. Test the use of Potential Coupling Interfaces (PCI) as a model coupling tool for the Tenmile watershed in Washington state, USA;
2. Use Potential Coupling Interfaces to couple TOPNET and MODFLOW models with application to the Big Darby watershed
3. Calibrate and validate the coupled model for the Big Darby watershed;
4. Test the functionality of the coupled model by evaluating its effectiveness in predicting the effects of land use changes and changes in ground water withdrawal rates on stream flows.

Relevance and Contributions of This Work

The major contribution of the research will be the development of a hydrological modeling system to predict the groundwater–surface water dynamics based on forcing functions and land use characteristics. For water resources managers, the question that needs to be addressed is: “How much water needs to be allocated from surface and ground sources to each of the various customers in a typical river basin in such a way that the overall benefit of the water resources of the watershed is maximized and ecological degradation is avoided or minimized?” Such information is crucial for workers attempting to 1) manage groundwater withdrawals, 2) prevent stream flow depletion, 3) quantify groundwater pollutant loading to streams and rivers, 4) define the role groundwater plays in maintaining stream flows, and 5) design and implement in stream-habitat protection and restoration programs.

This research will provide an initial tool that can be used to assess the influence of groundwater flow dynamics on surface water flow dynamics and vice versa. Knowledge of such information is critical in water resources allocation decision support systems.

Stresses such as increased or continued high levels of water use, change in land use, climatic change and the resulting response in terms of groundwater levels, base flow, stream yield, and water quality are also of interest. The results from this research can be part of an integrated GIS–based Decision Support System, which will allow the stakeholders in any river basin to compose and evaluate management scenarios in a point-and-click interface to represent possible future conditions and compare them to current or historical baseline conditions. The river flows, water quality, and resulting impacts on water availability and fish habitat determine the differences between scenarios. These differences provide valuable information in evaluating management options that can form watershed management plans. Integration of models increases the value and reliability of information by providing easy access to data and results and ensuring data integrity through a common data platform.

Currently there are no studies available to explain how the shallow groundwater–surface water hydrologic system works in the Tenmile watershed. For the Big Darby watershed there is only one report for a study by Yu and Schwartz (1998) that examines the use of an integrated model to understand surface water ground water interactions. To examine surface water ground water interactions, a process-based framework synthesizing what is known about the hydrology (both ground and surface water) is needed. The framework required would describe sources and sinks of water, general directions of groundwater flow, and estimates of travel times.

Central components in any Decision Support System (DSS) for groundwater management are models for surface and groundwater. Integration of the groundwater model, MODFLOW, with the watershed hydrological model, TOPNET, can be a scientific tool for assessment of possible preventive measures ensuring a sustainable groundwater supply, as well as to develop a comprehensive methodology to perform a proper assessment of groundwater management. Possible users could be local authorities (e.g. county councils, municipal water administrations, local environment and health authorities, consulting companies, and governmental agencies).

Because this study will improve the understanding of the hydrology of the Tenmile and the Big Darby watersheds, it will provide a basis to interpret previously collected water quality data. The sources and amount of baseflow contribution to each of the watersheds from their subbasins will be more accurately known and understood, and water resource managers will have a tool (groundwater–surface water flow model) to assess the effects of hydrologic stress on water resources

Organization of Dissertation

Chapter 2 gives an overview of related research work in groundwater–surface water interactions and integrated simulation models. Chapter 2 also outlines the background issues in studying groundwater–surface water interactions. This chapter explains the interaction between groundwater and surface water and also explains different model coupling techniques. Chapter 3 briefly describes the models that are used in this investigation namely MODFLOW, TOPMODEL, and TOPNET. Chapter 4 describes the first case study which involves coupling of TOPMODEL and MODFLOW using data for the Tenmile watershed in Washington State. This chapter describes the Tenmile study site, the data used as input and its sources, the development and calibration of the MODFLOW and TOPMODEL models for the watershed and the results of the calibration, the design of the coupled model using PCIs, and the results and discussion of the use of the coupled model to describe the hydrologic balance of the watershed.

Chapter 5 describes the coupling of TOPNET and MODFLOW. This chapter describes the Big Darby watershed, the data used as input, the rainfall runoff model development using TOPNET, the groundwater model development using MODFLOW, the design of the coupled model using PCIs, and the results and discussion of the simulations obtained using the coupled models. The results of using the coupled model developed in Chapter 5 as a tool to evaluate the effects of land use changes and withdrawal rates on stream flow are also described in this chapter. Chapter 6 outlines the dissertation summary, conclusions and recommendations.

CHAPTER 2

LITERATURE REVIEW

Hydrologic Modeling: An Overview

Over the past century, there has been numerous advances in the understanding of groundwater and stream hydraulics, runoff processes, and quantitative geomorphology, and improvements in data collection techniques, statistical applications to hydrologic data, and numerical methods used for modeling hydrologic processes. This knowledge base is critical for scientific development in physical hydrology. Two major areas of research as outlined by Dingman (2002) are 1) enhanced study and modeling of hydrologic processes at various scales (e.g. hill slope, basin, continental, global), and 2) the need for a more detailed empirical knowledge of the mechanisms involved in aquifer-stream interactions.

In this regard, simulation modeling has become an increasingly efficient and effective method of investigating hydrologic processes. This is mainly because of advances in data collection methods, storage and processing. Dooge (1986) defines a simulation model as a representation of the physical world "which is simpler than the prototype system and which can reproduce some but not all of the characteristics thereof." In this research , the term 'hydrologic model' refers to a mathematical tool that can be used to simulate any one or more of these watershed hydrological processes: runoff, stream flow, groundwater flow, infiltration, snowmelt, evapotranspiration, recharge, and ground water–surface water interactions. These processes are represented in Figure 2.1.

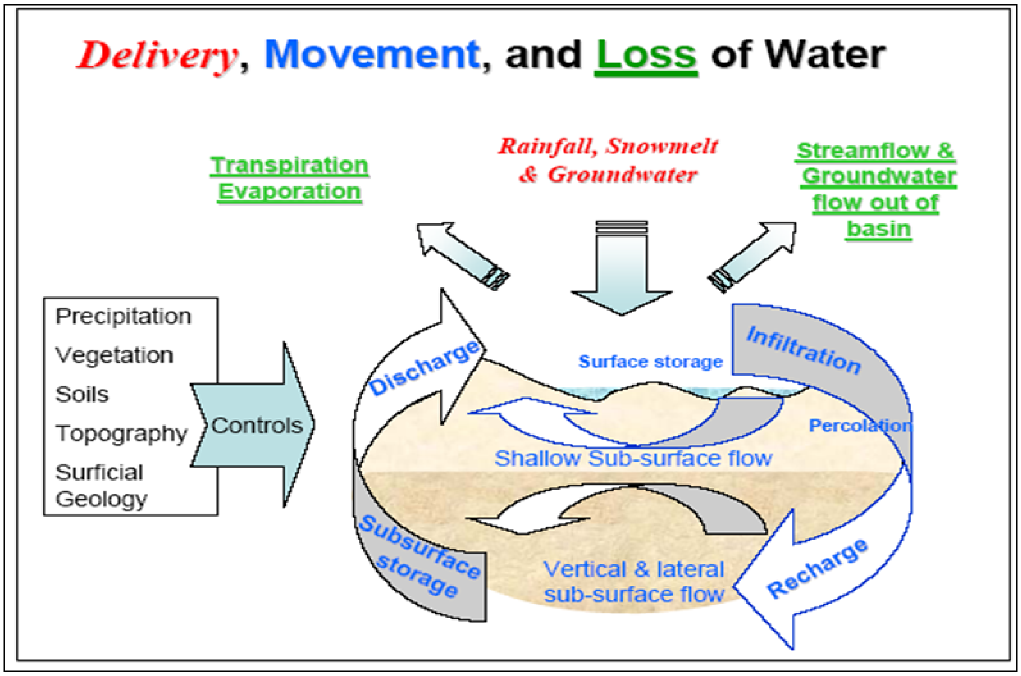


Figure 2.1. Hydrological flow mechanisms in a watershed (Stanley et al., 2005)

Hydrological models are generally described as, empirical, compartmental, or physical. Empirical models use mathematical relationships that have been developed from observational data to model the hydrologic process being studied. Such relationships are based on factors such as soil type and basin topography. Examples of such models include the Soil Conservations Service (SCS) curve number method for runoff determination.

Compartmental models represent hydrological systems as interconnected networks of "black boxes." Each "black box" represents a separate physical domain with uniform hydrologic characteristics (e.g. vadose zone, flood plain, and river). These compartments are linked through transfer functions to simulate the storage-discharge relationships between the compartments. Examples of this type of model include the

Stanford Model IV (Crawford and Linsley, 1966), the Sacramento model (Burnash, Ferral and McGuire, 1973), and the SWAT model (Arnold, Allen, and Bernhard, 1993).

There are also semi-distributed methods such as TOPMODEL (Beven and Kirkby, 1979) and ARNO (Todini, 1996) in which a hydrologic basin response is driven by a statistically representative set of grid cells determined by a topographical index. Physically based models use relationships derived from the basic concepts of physics such as conservation of mass, energy or momentum, diffusion and force balance to simulate flows and storage. Due to the nature of these relationships, physically based models are "distributed," the spatial domain is discretized into cells or elements in order to assign hydrologic parameters (valid for the entire cell or element) that can be observed or estimated.

Need for Hydrologic Models

There are two fundamental purposes of a hydrologic model: 1) scientific inquiry, and 2) resource management. A hydrologic model helps to develop a quantitative understanding of that data. Secondly it also provides a scientific basis for water management and for the administration of water resources. An example is that water resources managers and stream habitat modelers need tools that can be used to evaluate the influences of continuous ground water withdrawals on stream flows.

Models have also been used as tools to understand effects of past water management strategies. They can therefore also be used as prediction tools to evaluate or potentially even optimize future management options for water planning and management. In developing a model for water use planning, it is important to be sure that

1) the model will answer the relevant management questions and 2) sufficient data and understanding of the system are available to create a representative model.

Applications in both scientific inquiry and resource management require either a prediction: an estimate of the magnitude of some hydrologic quantity in response to a certain hypothetical event or stressor (e.g. estimating aquifer yield from average annual rainfall for water supply design); a forecast: an estimate of the hydrologic response to an anticipated event (e.g. estimating river flooding from a recent storm); or a hindcast: an estimate of an unmeasured hydrologic response to a previous event (e.g. estimating river stages in years prior to dam construction) (Dingman, 2002). The majority of models are designed for (and thus restricted to) a limited set of applications due to the basic assumptions that each model must make in order to represent the spatial and temporal scale of interest. For example, a flood prediction model that provides an estimate of a river's stage height given certain precipitation level (hourly/daily time scale) is generally not able to accurately estimate seasonal groundwater contributions to the stream (baseflow) for the same watershed (monthly/yearly time scale). This issue of scaling in hydrologic models is an ongoing debate in the literature (Beven and Kirkby, 1979; Grayson, Moore, and McMahon , 1992; Beven, 1995).

Studying Groundwater–Surface Water Interactions

In traditional hydrologic science, runoff modeling was mainly focused on quantifying water quantities and fluxes at the catchment scale only. However the development of holistic problem solving techniques in recent years has resulted in environmental problems providing an additional demand for hydrological models to serve as the foundation of biogeochemical models. This means that accurate quantification and simulation of internal variables, such as groundwater levels, and simulation of the interactions between the saturated and unsaturated zone have become important. In situations where the ground water table rises, e.g. after rain storms, there is a significant movement of water from the saturated zone to the unsaturated zone. The process is reversed when these groundwater tables fall. In order to satisfy the continuity conditions for the mass flow of water, this interaction between saturated and unsaturated storage has to be taken into account.

The significance of the interaction between the groundwater zone and soil water zone depends on the depth to the groundwater table. Three levels of interdependency can be identified:

1. If the groundwater table is comparatively deep (several meters), the connection is unidirectional, with groundwater recharge taking place during periods with high water content in the soil water zone. The soil water content is independent of the groundwater level.
2. With decreasing depth to the groundwater table there is an increasing interaction. As the groundwater table approaches the lower boundary of the root zone, the water in this zone moves to establish hydrostatic equilibrium with the groundwater table. A high groundwater table gives high soil water content, and only a small amount of infiltration is needed to give groundwater recharge. Still, the vertical extension of the soil water zone may be assumed to be constant over time, but the unsaturated storage at field capacity depends on the depth of the groundwater table.

3. With a very shallow groundwater table, the interaction becomes very strong. The groundwater table strongly influences the water content in the unsaturated part of the root zone and the groundwater table represents a moving boundary between saturated and unsaturated conditions. The latter results in a continuous transformation of root zone water between unsaturated and saturated conditions, with a rise in the groundwater table leading to a decrease in unsaturated soil water storage.

Simulation models must be available that can simulate these three different cases. However traditional conceptual models, such as the Hydrologiska Byråns Vattenbalansavdelning (HBV) model (Bergstrom, 1995) or TOPMODEL (Beven et al., 1995), are not capable of simulating the latter case, (i.e. a decreasing unsaturated storage with increasing saturated storage). The inability to distinguish correctly between the two different storages may hinder the use of the hydrological model as a foundation for the simulation of hydrochemical processes. Another problem with many conceptual models is caused by the simplified description of the relationship between groundwater storage and runoff. These models usually represent a catchment using a number of storages. One or more of these storages may represent groundwater storage and thus can be related to groundwater levels. In most conceptual runoff models, an unambiguous, monotonic function between the groundwater storage and runoff is implemented. Consequently, the dynamics of the simulated runoff from the groundwater zone always follow the simulated rise and fall in groundwater levels.

Thus this study was proposed to develop a modeling tool that can effectively describe these interactions by coupling surface water and ground water models that have already been developed and are know to effectively describe surface and subsurface flow systems.

Stream aquifer interactions

Winter (1995) made an analysis on recent analysis in understanding surface water groundwater interactions. Stream-aquifer relations have practical importance in the evaluation of surface water resources in a basin. The amount of available surface water along a river for irrigation, drinking or any other use is defined by the flows in the river and depends on water transfers with subsurface aquifers. The evaluation of flows, whether from or to the aquifer, is difficult because of several factors that influence the water movement. However, it is possible to evaluate transfer flows in irrigated areas where inter annual water movement cycles are relatively constant. Stream-aquifer water movement is strongly affected during floods. Water level fluctuations in the river produce important changes in hydraulic transversal gradients of groundwater near the stream and, consequently, variations in groundwater velocities and flows.

Before the advent of numerical modeling, interaction of surface water with groundwater in alluvial aquifers was concerned with analytical solutions to 1-dimensional flow of groundwater to fully penetrating streams (Rorabaugh, 1964). This approach is still being used today to estimate groundwater recharge from stream flow hydrographs (Beven, 1986b), and automated computer-based techniques for using these analytical methods were recently developed (Rutledge, 1992). Determination of groundwater baseflow from recession analysis of stream flow hydrographs, hydrograph separation, continue to be used e.g., methods for determining the baseflow component of stream flow graphically have recently been developed by the Institute of Hydrology (United Kingdom), by Wahl and Wahl (1988), and by Rutledge (1992). Mathematical digital filtering has also been used recently to determine the baseflow component of stream flow (Nathan and McMahon, 1990). Although hydrograph analysis continues to be used, most recent studies have used other analytical techniques and numerical modeling.

Heij (1989) used analytical methods to determine travel times of water seeping from surface water into contiguous aquifers in The Netherlands. He found a linear relation between the surface-water level and the infiltration rate and an inverse relation between the surface-water level and the average time it takes water particles to flow into the stream bank. Bank storage, the movement of surface water into groundwater at times of high river stage, was the focus of a study by Hunt (1990) who developed an approximate flood-routing solution for coupled groundwater and open-channel flow equations. Neglecting seepage initially, solution of a linearized kinematic wave equation was used to obtain a solution for the groundwater movement. This in turn was used to obtain a second-order solution for the flood routing. An example of using this approach indicated that changes created in the downstream hydrograph by bank storage could be as large as changes created by retaining all terms in the open-channel flow equations and routing the flood down the channel with no bank storage.

Zhang (1992) developed solutions for transient flow in an aquifer-aquitard system that considers storativity in a confined layer in response to abrupt changes in water level, uniform changes in water level, and steady rates of seepage from a river. From these equations, he determined groundwater levels for the aquifer and the aquitard, as well as rates and total volume of seepage from the river. Transient conditions were also of interest to Rastogi (1991), who determined seasonal groundwater flow to a river reach bounded by two reservoirs, where the water-table aquifer was underlain by an impermeable bed. The objective was to determine the amount of groundwater that could be developed from this aquifer system that was receiving seepage from the upstream reservoir, losing seepage to the downstream reservoir, and receiving seepage from the river.

The depletion of stream flow by pumping groundwater from the contiguous alluvial aquifer has been a major impetus to studies of the interaction of groundwater and surface water. A recent paper on this problem (Wallace, Darama, and Annable, 1990) was concerned with comparing a dimensionless volume of stream depletion over a pumping cycle with maximum rate of stream depletion at a practical state of dynamic equilibrium. Dimensionless plots of equations developed by applying superposition principles to analytical solutions for steady continuous pumping were used in the study. Although the plots provided a way to quickly determine the time at which a practical state of dynamic equilibrium is reached, the study also indicated that under some conditions approximating cyclic pumping using steady continuous pumping at the equivalent cycle-average rate is inadequate. In another study involving the effects of pumping groundwater on stream flow, Spalding and Khaleel (1991) compared the results of several analytical solutions to a two-dimensional groundwater flow model. They found that simplifying assumptions needed for use of the analytical methods resulted in differences in stream flow depletion from the numerical model that ranged from 20 percent, due to neglect of partial penetration, to 45 percent, due to neglect of clogging layer resistance, after 58 days of pumping.

Use of the analytic element method (Strack, 1989) has recently been expanded for modeling the interaction of groundwater and surface water. Mitchell-Bruker (1993) used this method to investigate the hydrologic effects of changing recharge and boundary conditions on groundwater flow in the Pere Marquette River Basin, Michigan. She found that, especially on the local scale, as recharge varies areally, the contributing area to a surface-water body changes. The boundaries of regional groundwater systems are more stable because the local variations tend to be averaged.

Statistical methods have also been used recently to study problems related to the interaction of groundwater and surface water. For example, Adamowski and Feluch (1991) proposed a new nonparametric regression model to investigate the relation between fluctuations in groundwater levels and time series of stream flow. They determined that the nonparametric method resulted in more accurate predictions than those obtained from parametric regression.

In another study involving time-series analysis, Niestle and Reusing (1990) compared Autoregressive Moving Average and Fractional Gaussian Noise models to assess their reliability for the analysis of drought risk of the Nile River at Aswan, Egypt. River discharges were converted to water levels, which were then used as input to a simulation model of the interaction of the Nile River with groundwater. Although statistics on low stream flow have been used for many years in studies of the interaction of groundwater and surface water, Vogel and Kroll (1992) found that in western Massachusetts baseflow recession constants could be used as a surrogate for basin hydraulic conductivity and drainable soil porosity.

Modeling Groundwater–Surface Water Interactions

Although early work in hydrology emphasized the linkages between surface water and groundwater (Theis, 1941; Rorabaugh, 1964), water managers have always looked at groundwater and surface water as two separate entities. With increasing development of land and water resources, however, the understanding that development of either of these resources will affect the quantity and quality of the other has gained importance (Winter et al., 1999). Watershed hydrology represents a complex interconnection of flow paths coupling reservoirs in the atmosphere and biosphere, surface water bodies, streams, the soil profile, vadose zone and groundwater. Understanding of these interconnections has resulted in a large body of literature on groundwater–surface water interactions and their ecological, economic, and legal implications. Comprehensive reviews of that literature given by Winter (1995) and Sophocleous (2002). Bouwer and Maddock (1997) outline some of the legal ramifications of groundwater–surface water interactions; Glennon (2002) describes a series of case studies where groundwater use has negatively affected surface water; and theoretical considerations of river–aquifer interactions and their mathematical formulation are discussed in Kaleris (1998) and Rushton and Tomlinson (1979).

Groundwater discharge to streams, or baseflow, often constitutes the major source of stream flow during dry periods. During these periods groundwater use is usually highest. Kondolf et al. (1987) described the impacts of groundwater pumping on stream flows in a case study of the Carmel River in California. Groundwater withdrawal locally decreased or even eliminated base flows. Quantity and timing of base flows were identified as important for fish migration. Along the Mojave River in California, increasing groundwater pumping has caused seasonal and long-term stream flow depletion (Lines, 1996). Chen and Soulsby (1997) used a numerical model to assess the impacts of proposed groundwater development on stream flow in a nearby stream that was important for salmonids. In their study, changes in stream stage caused by the proposed development were small and were found to have only minimal impacts on fish habitat. Ramireddygari et al. (2000) used a numerical groundwater and surface water model to investigate the effects of irrigation practices and stream diversions on river flows and water levels in an environmentally important wetland in Kansas. They found that stream flows were most sensitive to changes in groundwater pumping for irrigation. Under increasing pressure to meet water demands and yet comply with environmental standards, numerical models that include stream-aquifer interactions have become indispensable tools for water management in many parts of the world (Pucci and Pope, 1995; Perkins and Sophocleous, 1999a; Sophocleous and Perkins, 2000).

The complex interconnection of watershed flow paths mentioned above can be reduced to a tractable form for a particular problem by limiting its scope to a domain within the watershed, and specifying boundary conditions to represent hydrologic connections to the remainder of the watershed. A watershed is typically partitioned into unsaturated and saturated zones of porous media, surface reservoirs and a drainage network of streams. Models are available which simulate hydrologic processes in these domains.

Due to increased economic requirements, water resources managers now use numerical models to simulate the behavior of water systems and implement resource planning to meet industrial, domestic, agricultural and municipal water needs. Models that can simulate all the significant conceptual flow paths in a watershed and provide an overall hydrologic balance are called integrated watershed models. A number of integrated basin scale hydrologic models to simulate surface–subsurface interactions have been developed in recent years. Pucci et al. (1995) used a quasi-three-dimensional finite difference model that incorporated the effect of surface–groundwater interaction to describe the groundwater flow of an aquifer in New Jersey. Sophocleous et al. (1999) linked a watershed model with a groundwater model to simulate the stream-aquifer system in a river basin. The surface water flow model, Potential Yield Revised (POTYLDR), and the groundwater flow model, MODFLOW, were combined into an integrated, watershed scale, continuous simulation model. Enhancements were made to the POTYLDR and MODFLOW models for simulating the detailed hydrologic budget for the Wet Walnut Creek Watershed in Kansas. The computer simulation model was calibrated and verified using historical stream flow records (at Albert and Nekoma gaging stations), reported irrigation water use, observed water level elevations in watershed structure pools, and groundwater levels in the alluvial aquifer system. The interface module links the two models by taking a spatially weighted average of the surface response at each time step and distributing the resulting interface fluxes to the corresponding spatial and temporal location in the aquifer model. The resulting "integrated" model has significantly lower input data requirements than other fully distributed watershed models (e.g. MIKE-SHE, Abbott et al., 1986) while enhancing the ability to examine stream–aquifer interactions, distributed well withdrawals and land use impacts. These studies show that the versatile applicability and medium complexity of this approach facilitate its use in resource management applications.

BRANCH (Schaffranek, Baltzer, and Goldberg, 1981), a physically based model for open channel dynamics, has been combined with the Modular 3-D Finite Difference Groundwater Flow Model, MODFLOW (McDonald and Harbaugh, 1988) to create the integrated model known as MODBRANCH (Swain and Wexler, 1993). In MODBRANCH, terms that describe leakage between a stream and an aquifer are added to the continuity equation in BRANCH and a package was added to MODFLOW to interface with the modified BRANCH. Leakage between the aquifer and the stream can be calculated separately in each model, or leakage calculated in BRANCH can be used in MODFLOW. MODBRANCH calculates new stream stages for each time interval in a transient simulation based on upstream boundary conditions, stream properties and initial estimates of aquifer heads. Aquifer heads are then calculated in MODFLOW based on stream stages calculated in MODBRANCH and aquifer properties. Because time steps used in groundwater modeling can be much longer than time steps used in surface water simulation MODBRANCH can handle multiple surface water flow time steps contained in one groundwater flow time step.

MIKE-SHE ("System Hydrologique European," Abbott et al., 1986) is a physically based watershed model that can simulate surface and groundwater movement, the interactions between the surface water and groundwater systems, and the associated point and non-point source water quality problems. The system has no limitations regarding watershed size. MIKE-SHE subdivides the modeling area into polygons based on land use, soil type, and precipitation region, and the polygons are then assigned identification numbers. Model input files can be generated by overlaying the model input parameters with a grid network. Data preparation and model setup can be completed using Geographic Information System (GIS) software, Arc View, or MIKE-SHE's built-in graphic pre-processor. The MIKE-SHE modeling system simulates hydrology components, including the movement of surface water, unsaturated subsurface water, saturated groundwater, and exchanges between surface water and groundwater. With regard to water quality, the system simulates sediment, nutrient, and pesticide transport in the model area. The model also simulates water use and management operations, including irrigation systems, pumping wells, and various water control structures. The system has a built-in graphic and digital post-processor for model calibration and evaluation of both current conditions and management alternatives.

In regions with high groundwater flow transmissivity and shallow water tables much of the groundwater behavior has been linked or correlated to elevation and fluctuation of nearby streams and rivers. Meyer and Turcan (1953) correlated water elevations on the Mississippi River to a well located about two miles from the river. The groundwater elevations of the well followed that of the river very closely.

Definition and Conceptualization of Model Coupling

Hydrological Model coupling requires a thorough conceptualization of the coupling strategy which includes a definition of the individual model domains, the “transboundary” processes and the exchange parameters. In coupling a groundwater flow model and a surface water flow model, it is very important to find a common definition and scale-appropriate process description of groundwater recharge and baseflow in order to achieve a meaningful representation of the processes that link the unsaturated and saturated zones and the river network. As such, integration by means of coupling established disciplinary models is problematic given that in such models, processes are defined from a purpose-oriented, disciplinary perspective and are therefore not necessarily consistent with definitions of the same process in the model concepts of other disciplines.

Model coupling here means coupling of distinct existing models or model concepts that were developed to simulate processes in one “system”. Coupling in the present context mainly means coupling via exchange variables rather than directly coupling process equations and code.

The first step in attempting to couple two models that describe different but interdependent systems should be the consideration of some basic questions. They include the questions that should generally be asked before starting to model. These questions relate to the problems the coupled model complex will be used to solve, the output variables that are required, the relevant scales, the required accuracy of the results, the data availability, etc. As consideration of these general questions should be a standard procedure in model conceptualization, the topic is not elaborated on. In addition to these general issues, there are a number of questions that relate specifically to model coupling of the two systems:

1. How are the individual systems defined, and what are the (dominant) processes that take place in each system?
2. Where and what is the boundary between the systems? Is it sharp and stable or just a virtual, time-dependent boundary?
3. Which processes connect the systems to each other? Are the connecting processes clearly related to processes that take place within the individual systems?

4. Which process descriptions are needed, which are available, which are applicable in view of discretization and data availability?
5. What are the dynamic relations between the two systems (uni- or bi-directional, feedback, different dynamics)?
6. Which measurable quantities are available to determine the effect of changes in inputs to the individual system and how do these quantities relate to the connecting processes?
7. What are the relevant process scales (time and space) and how are they related to the scale of the problem? Are the relevant scales equal on either side?

Answers to the questions listed above lead to the definition of system boundaries, connecting processes, exchange variables and appropriate scales, and finally, to a first conceptual description of at least one possible coupling approach. Once this conceptualization has been achieved, the next step is to choose (or to develop) the appropriate individual models for each system and the mechanism for their coupling.

Methods for Coupling Hydrological Models

There are several ways in which distinct hydrologic models can be coupled. Various authors categorize the methodologies of coupling in different ways. For example, Jewitt (1998) categorizes them into two categories: series and parallel, whereas Chou and Ding (1992) offer a cubic perspective using a data sharing method, a modeling method, and a user interface as the dimensions of the cube. Brandmeyer and Karimi (2000) give a detailed description of the different types of coupling, establishing five possible levels of integration: one-way data transfer, loose coupling (two-way data transfer), shared coupling (sharing one component, Graphic User Interface [GUI] or data storage), joined coupling (sharing both components), and tool coupling (models are coupled using an overall modeling framework). A brief description of these coupling methodologies will be summarized here; a further discussion of their advantage and disadvantage can be found in Brandmeyer and Karimi (2000).

One-way data transfer coupling

This is the most basic level in model coupling hierarchy. In this approach of coupling, the two models to be coupled (Model 1 and Model 2) remain completely separate; they are coupled only in the sense that the output produced by one model becomes input to the other (Figure 2.2). Data is passed from one model to the other through non-automated (manual) extraction, transfer, and a conversion process. The coupling process doesn't have an impact on individual models; therefore, each model may continue along its own development path without influencing the other.

Loose coupling

Loose coupling, also known as series coupling in the literature (e.g. Jewitt, 1998), is similar to the one-way data transfer except the manual data transfer is replaced by an automated system and the interaction can be two-way (Figure 2.3). Converting the output of one model's data structure to the format of the other model often involves the use of some sort of data transformation program.

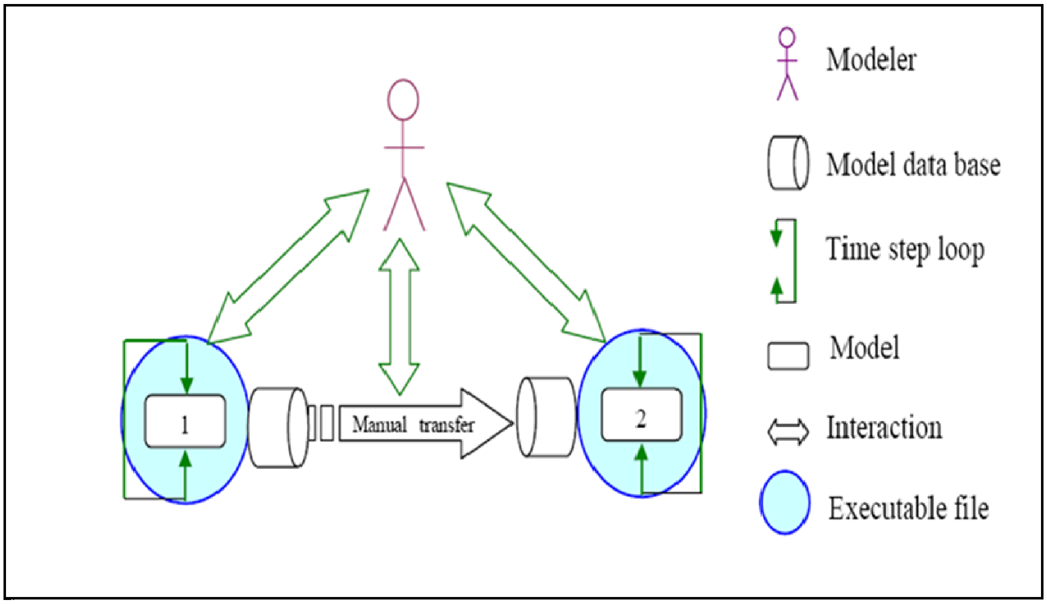


Figure 2.2. One-way data transfer coupling (Adapted from Jewitt, 1998 and Brandmeyer and Karimi, 2000).

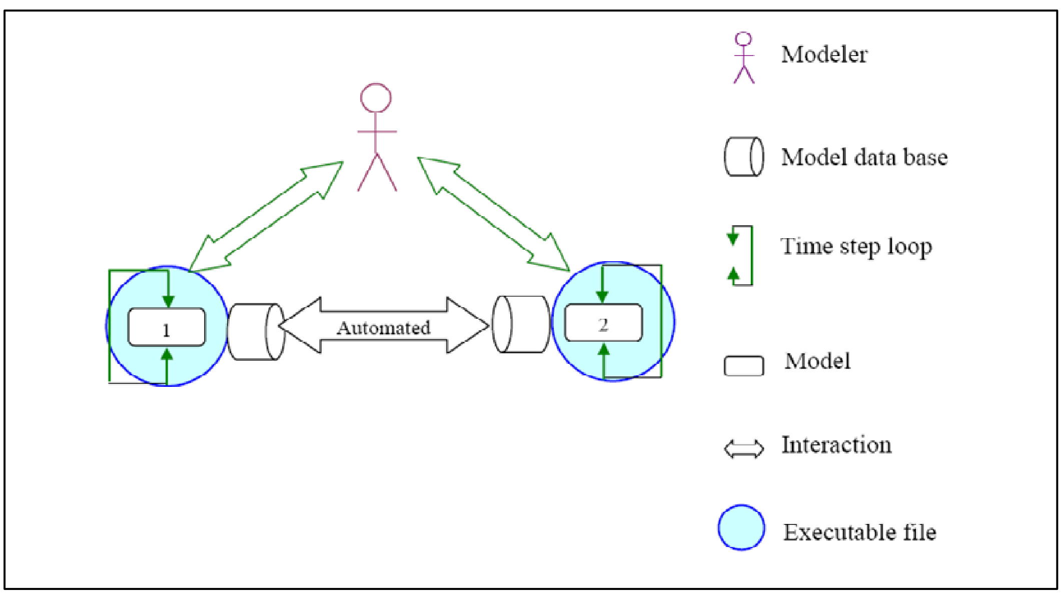


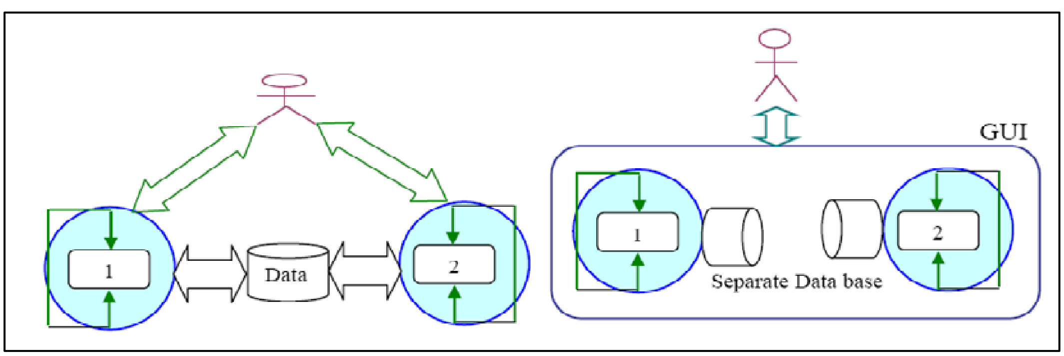
Figure 2.3. Loose coupling (Adapted from Bandmeyer and Karimi, 2000).

Shared coupling

In shared coupling (Figure 2.4), the models share a major component of the data base system or the Graphic User Interface (GUI). In data coupling, the user interacts directly with each model's user interface, but the models share data files from a common database. On the other hand, in GUI coupling, a single user interface provides a user-friendly method of coupling the models, but data are stored separately for each model.

Joined coupling

Joined coupling utilizes both the GUI and common data base from shared coupling; however, the structure of the model relationship is different. The model relationship can be embedded or integrated. In embedded methodology, one model contains another in a master–slave relationship, and they can also be compiled in one executable program (Figure 2.5a). The user interacts only with the master model through its user interface. In contrast, in integrated coupling the two models have a peer relationship and the user interacts with any of the models using the common GUI (Figure 2.5b). Besides, the two models can have a shared library of functions and subroutines.



a b

Figure 2.4. Shared coupling a) Data coupling b) GUI coupling (Modified from Jewitt, 1998 and Brandmeyer and Karimi, 2000).

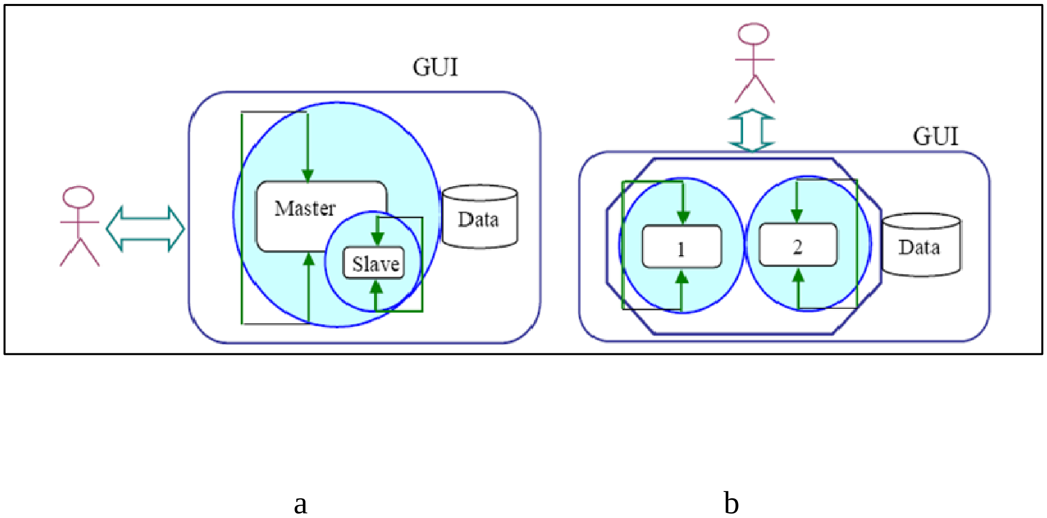


Figure 2.5. Joined coupling (a) Embedded coupling and (b) Integrated coupling (Adapted from Jewitt, 1998 and Brandmeyer and Karimi, 2000).

Tool coupling

Tool coupling is the most sophisticated coupling methodology using a modeling framework. This approach has a framework that has integral subsystems wrapped within a common user interface. Within the framework can be both joined and shared coupling within sand between the subsystems. The framework consists of five subsystems and a GUI capable of operating in a networked computing environment (Figure 2.6).

The framework also provides functions and tools common to multiple models, while managing data and computing resources (Brandmeyer and Karimi, 2000). In this approach the subsystem can be resident on one computer or could be distributed over a network. Subsystems can be data management, spatial data processing, model building and management, model execution and quality management.

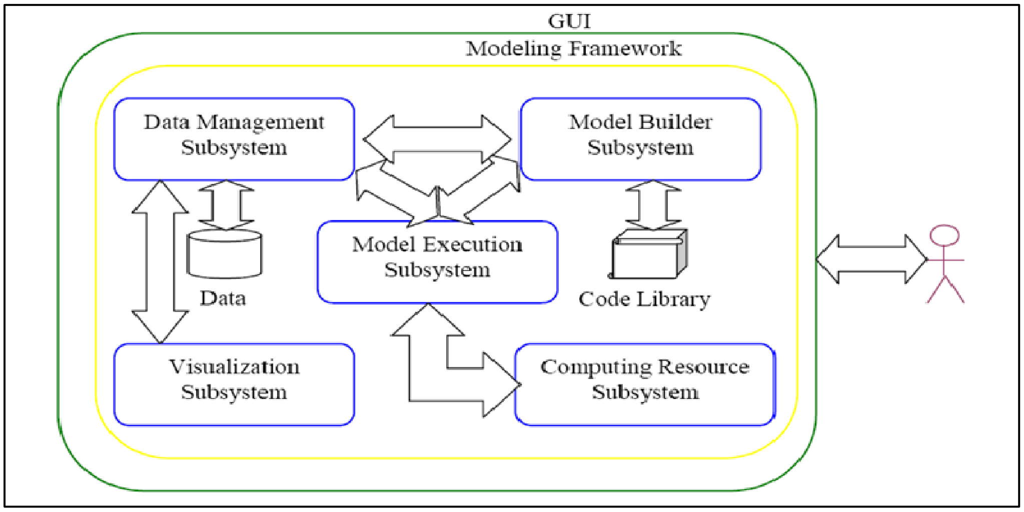


Figure 2.6. Tool coupling (Adapted from Brandmeyer and Karimi, 2000).

Overview of the Potential Coupling Interface Approach

This section will give a description of the design of the Potential Coupling Interface (PCI) for a model and the process for creating it will be highlighted. The basic idea of model coupling using PCIs, developed by Bulatewicz (2006), is for the state of one model affect and be affected by the state of another model. The state of one model in this case is the combined effects of the state variable sin the model. Therefore the important initial step in the design of coupling interfaces is the identification of model state variables. The designer also has to identify areas in model code where the state variables have meaningful values, usually ate the beginning or the end of a time step. The number of times that model state variable can be accessed in each simulation depend on the number of times a particular location in a model is reached. This therefore affects how many times the state variables are accessible to be influenced or influence another model. Therefore, control structures such as loops dictate the accessibility of state variables. In summary the three basic elements of model coupling are:

(i) Identification of state variables

(ii) Identification of places/locations where state variable have meaningful values and

(iii) Control structures surround these locations.

Model coupling Interfaces are best illustrated using Control Flow Graphs, CFG, in which blocks or nodes represent sections of model codes and directed edges represent flow of control between them. Each node may represent several statements in model code. This representation is similar to flow charts commonly used in model documentation. An example is MODFLOW which consists of blocks representation thousands of model code performing different tasks such as equation solving and input/output functions. In PCIs some of the flow chart blocks are combined and thus reducing the CFG as shown in Figure 2.7. Thus only those aspects of the model code relevant to model coupling are exposed.

In the CFG the dark arrows represent coupling points where state variables can be accessible. In Figure 2.7 the pop up window shown shows the location (coupling point) where a state variable such as, hydraulic head, Hnew, is accessible. The loops around the coupling points are the control structures.

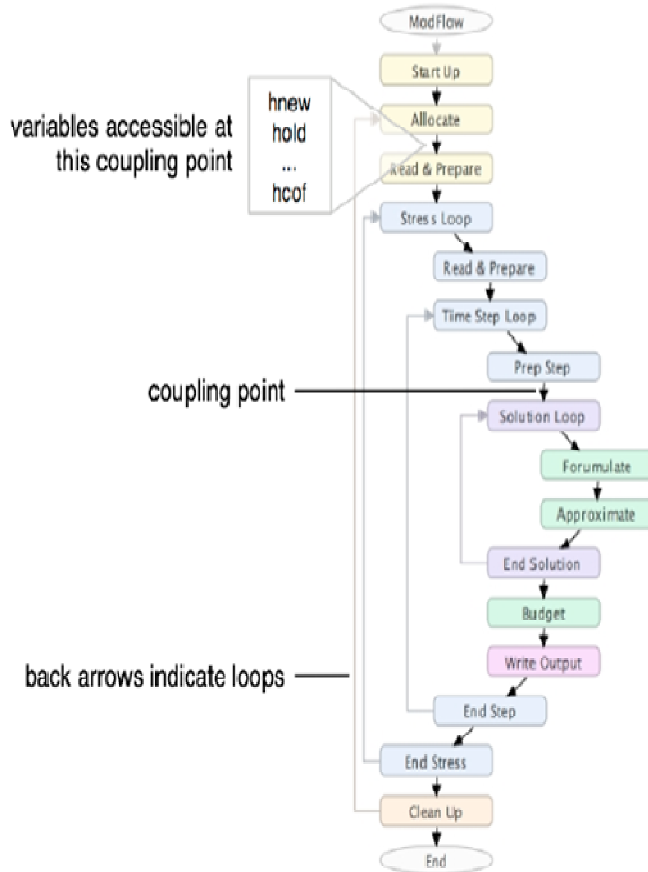


Figure 2.7. A PCI for MODFLOW (Bulatewicz, 2006).

PCICreate Software

This is Java written software developed by Bulatewicz (2006) to automatically convert model code into CGFs. It uses the open source JGraph component to display CFGs. Figure 2.8 is a screen shot of this application for MODFLOW model.

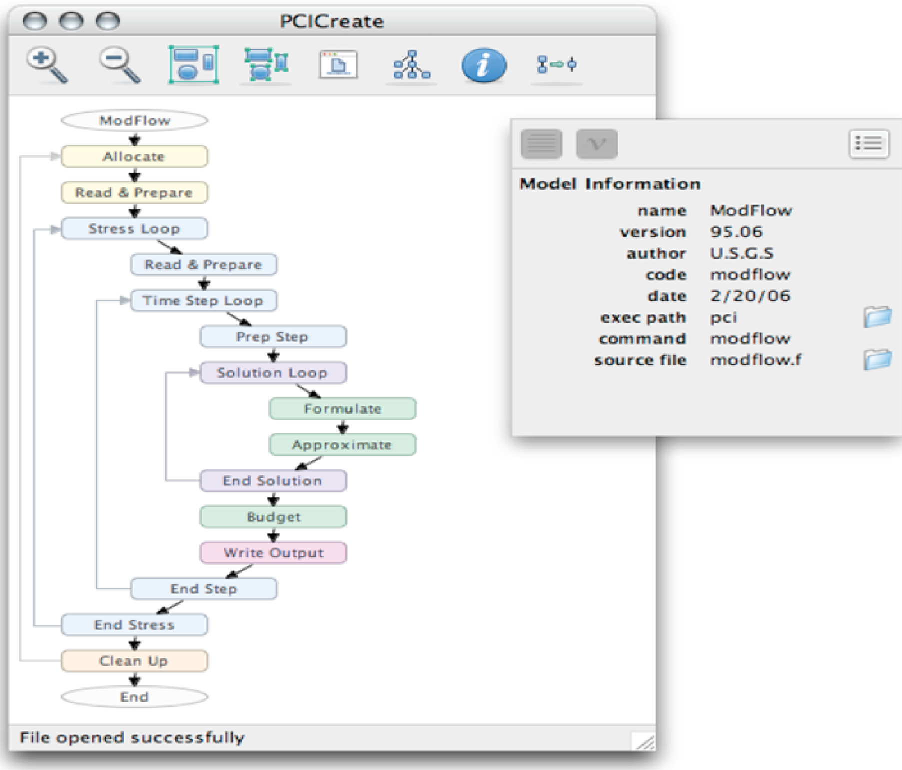


Figure 2.8. PCICreate application (Bulatewicz, 2006).

In Figure 2.8 the main editor window shows the PCI, while the smaller window on the right, also called the Inspector window shows information about the PCI, in this case MODFLOW information entered by the PCI creator. Figure 2.9 shows another screenshot of PCICreate. In this figure the state variable list of the selected coupling point is shown in the inspector window. This information about the variables is added by the user in the final step of the PCI creation process.

The various steps in the creation of model PCIs are shown in Figure 2.10. In this figure, dark arrows represent tasks performed by the modeler while light colored arrows represent automatically generated steps using PCICreate software.

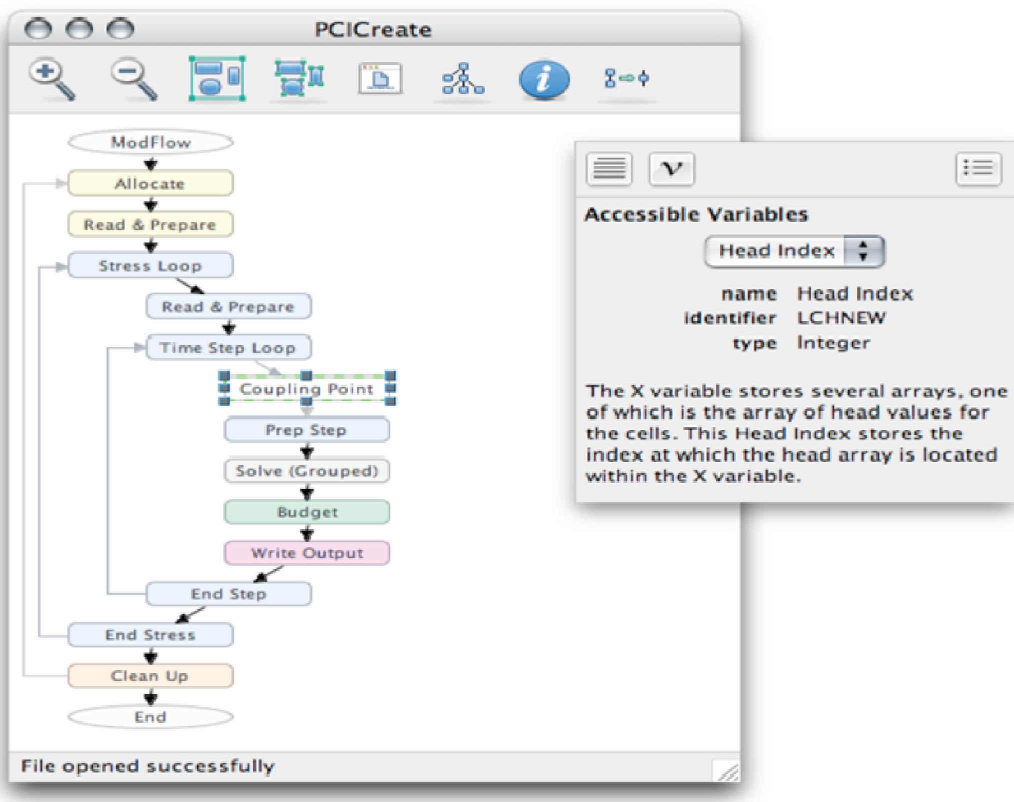


Figure 2.9. Inspector window showing available state variables at a coupling point (Bulatewicz, 2006).

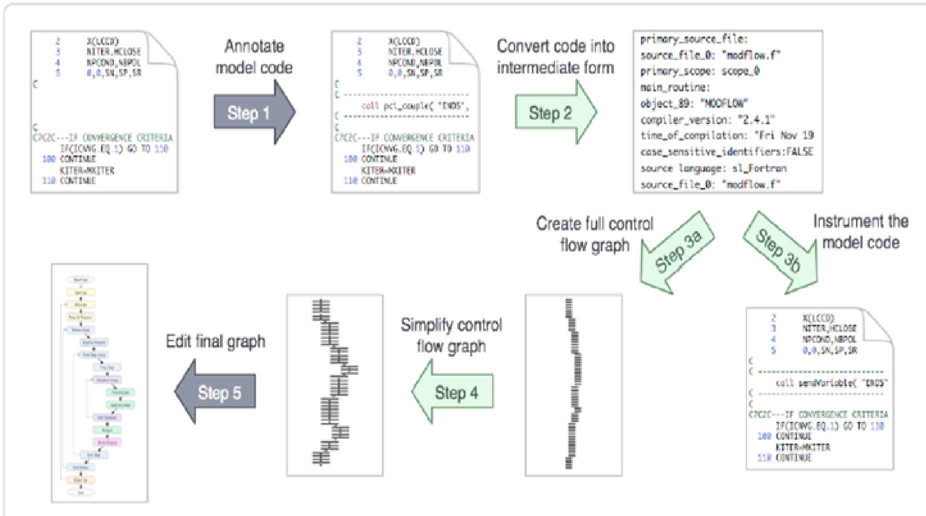


Figure 2.10. Steps in the creation of model PCI.

Step 1.

In this step the modeler annotates the model code, indicating which state variables should be accessible, and at what points in the code.

Step 2.

The annotated model code is automatically converted into an intermediary (analyzable) form.

Step 3a.

A complete control flow graph is derived from the annotated code and model code is instrumented

Step 4.

The complete graph is reduced around the coupling points.

Step 5.

The modeler customizes the simplified graph and incorporates any necessary domain-level information. Model annotations are program statements added to a model code that indicate where a coupling point should be added to the model's PCI and they are function calls as shown in Figure 2.11.

In the example annotation, a coupling point should be created at the specified point in the model code, uniquely identified by the name *asbi*, and that the four variables t*ime_delay*, *tdp*, *tdi*, and *cdi* should be accessible at the coupling point. Each variable in the annotation is followed by a number that indicates the number of elements in the variable (for scalars, the number of elements is 1).

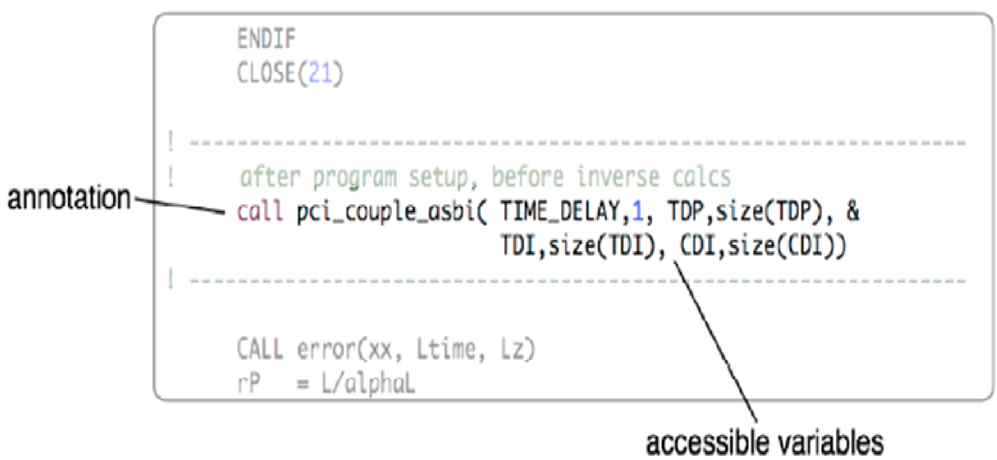


Figure 2.11. Example model annotation.

The annotated code is imported into PCICreate via the *Import* menu item, and is then translated (Step 2) from its source language into a structured intermediate form using the Program Database Toolkit (PDT) (Mohr et al., 2000). The intermediate form is then parsed by PCICreate in Step 3 to generate a complete control flow graph. With one block in the graph for each statement in the code, these graphs are generally huge, far too large to be comprehensible. The CFGs are reduced by combining or collapsing adjacent blocks of code using the Interval Analysis (Aho and Ullman, 1972) algorithm while preserving annotations and the control structures in the CFG. This reduction process significantly reduces the number of nodes of a CFG. An example is that in the ModFlow model, the complete CFG consists of 10,158 nodes, while the reduced graph consisted of only 35 nodes. In code instrumentation (Step 3b) of the PCI creation process, coupling independent communication code is added to the model source code, enabling the model to send and receive the value of any variable at any coupling point.

Model Coupling Environment

The PCICreate software is used to create model PCIs while software called the PCICcouple is used to describe and execute coupled models. Figure 2.12 shows the coupling environment within PCICouple. It shows two model PCIs. As in the PCICreate, the Inspector window is used to inspect different aspects of a PCI, and in the PCICouple it also includes information about couplings.

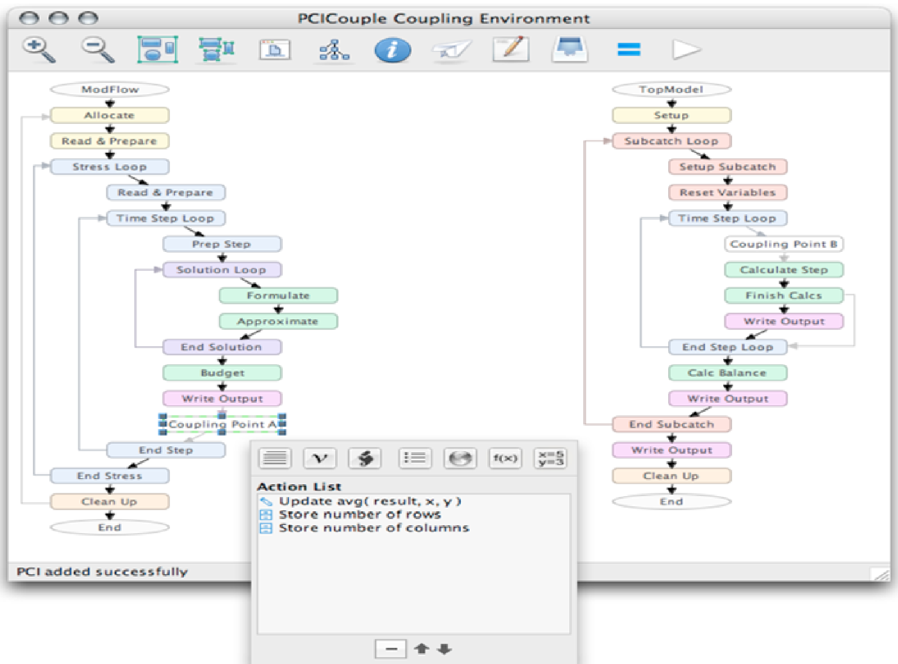


Figure 2.12. Coupling environment within PCICcouple.

The PCICouple consists of a set of operations called actions. Describing a coupled model is specifying a list of actions, action lists. The inspector window in Figure 2.12 shows an action list at a coupling point. During execution of the coupled model, when a model reaches a coupling or set point, the actions in its action list are carried out. There are four kinds of actions: Set, Send, Store, and Update. Basically, these actions allow the values of a model's variables to be changed based on the values of other variables from that model, or from other, coupled models. A full description of each of these actions is given by Bulatewicz (2006).

Executing coupled models consists of interchange of data between the models using the action lists mentioned above. When a coupling is executed in PCICouple, the description is compiled into scripts, and the controller is started. The controller oversees the execution of the coupled model and has control over starting and stopping the instances of each model. When a coupling point is reached by an instance, the coupling point's accessory function is invoked. The function iterates through the instance's script, sending and receiving values for all pertinent state variables.

Problems in Development of Coupled Models

Though coupled models are highly desirable for quantitative analyses, scientists have faced many problems in developing such models. As highlighted by Camp Dresser and McKee Inc. (2001), a model that integrates simulations of surface water and groundwater processes must account for the different scales of spatial and temporal variability of the two systems. Typical groundwater models that implement finite-element or finite-difference solution techniques discretize the model area into relatively small nodal elements or grid cells because the independent variables (head) computed by the model and aquifer characteristics can vary over relatively short distances. Although it is difficult to generalize over the entire class of surface water models, some treat the model area as a set of large subbasins transecting several nodal elements or grid cells in groundwater models. Computed variables including stage, flow rate and runoff and specified parameters such as topography, bottom elevation and roughness often have a different spatial scale of variation than those of the groundwater system. The need for detailed spatial variability is characteristic of groundwater models and an integrated model might need to utilize groundwater nodal elements or grid cells to adequately simulate water movement between the surface and subsurface.

On a temporal scale, surface water models often use small time increments (minutes to hours) to depict changes in the system such as large storm events or releases of water in rivers or canals. Groundwater models, because of the naturally slower groundwater flow (laminar flow), require longer time periods (weeks to months or years) to simulate groundwater movement and solute transport.

The second context in which multiple-model integration projects have occurred is in conjunctive stream-aquifer management, especially in the western U.S. where water appropriations from both surface and groundwater are a critical policy issue. One system created to support these policy decisions is described by Fredericks, Labadie, and Altenhofen (1998). It combines Modular Simulation (MODSIM), a generalized river basin network flow model, with MODRSP which is a modification of MODFLOW that includes the calculation of response functions for stream-aquifer interactions (Maddock and Lacher, 1991).

The modular design of MODFLOW encourages alterations and enhancements, and there have been various modifications to MODFLOW and surface water models to make them interact better. The MODBRANCH model (Swain and Wexler, 1993) is a modification of the BRANCH stream flow model that enables it to effectively function as a module within MODFLOW, thus providing more sophisticated surface water interaction than is provided by MODFLOW's Stream package.

Similarly, Perkins and Koussis (1996) replaced the Stream component with a model of their own that performs diffusive wave routing to model flood wave propagation in the surface water system. Their approach also includes a useful feature that allows for two different time scales to be used. The flood wave routing portion of the system can operate on shorter time steps, which are then integrated into longer time steps in the groundwater flow portion of the system. There have been some projects that include the integration of a storm water model and a groundwater model. Ross and Tara (1993) created a GIS-based system that connects HSPF to MODFLOW; however, the connection between the models only functions in one direction. Infiltrated water as calculated by HSPF is moved into the groundwater system represented by MODFLOW, but there is no capability for feedback to allow the MODFLOW results to influence the HSPF simulation.

A more ambitious integration is one that combines the SWAT model with MODFLOW (Sophocleous et al., 1999; Perkins and Sophocleous, 1999a). SWAT is a watershed model designed for use in rural basins that focuses primarily on storm water runoff but also includes simple routing of groundwater and stream flow. The SWATMOD combination combines SWAT's simulation of surface water processes with MODFLOW's much more powerful capabilities for modeling groundwater flow. This was accomplished by modifying MODFLOW so that it could be called as a subroutine by the SWAT model to run single time increments.

In summary, multiple-model integration projects have produced several useful examples. The storm water–receiving water combinations are the result of a conceptually simple operation: that of routing the output of the storm water model into the receiving water model as input. In such a configuration, no direct interaction between the models is necessary. The storm water model can run its simulation through to completion and write its results to a file, because it needs no information from the receiving water model. After that output data is appropriately reformatted, it then becomes input for the receiving water model, which can run through its entire simulation uninterrupted. Integration of groundwater and stream flow models involves more complex interactions between the two models, since each system affects the other continuously.

The examples cited above all involve restructuring the surface water model so that it can operate within the context of the groundwater model. The applicability of this technique is limited, because in some cases it would be extremely difficult to do such a restructuring without fundamentally changing the characteristics of the original surface water model. Accurate representation of the interactions between storm water runoff, infiltration, and groundwater flow requires that the storm water and groundwater models be more closely integrated than is required in storm water–receiving water model integration. The key to the integration is a method that allows for each model to interact with the other after each time step. But rather than encapsulating one model within the other, a more flexible approach is to alter both models so they can communicate with each other and still operate separately.

CHAPTER 3

MODELING TOOLS

Introduction

The research for this dissertation was undertaken with a goal of demonstrating a modeling system that can be used to evaluate and explain how surface and subsurface flow systems affect each other in an integrated way. Because it is a demonstration project and should have wide applicability in different watersheds it was imperative to select well-known, widely-tested models. In addition, because modifications to the models will be necessary, the selected models had to have source code in the public domain or code that can be made available. This research therefore makes use of TOPMODEL, TOPNET and MODFLOW models, and these are described in the following sections.

Topography-based Hydrologic MODEL (TOPMODEL)

The physically based hydrological model, TOPMODEL (Beven and Kirkby, 1979; Wolock, 1993; Beven et al., 1995) was selected for this study because of its relative simplicity and limited data requirements. The model consists of linear and exponential equations that are solved quickly and directly. Model efficiency allows a large number of simulations to be run, so a broad range of physical conditions can be explored. Notably, topography is distilled into a single topographic index that can serve as a first-order surrogate for the distribution of soil moisture. The required input data include a Digital Elevation Model (DEM) of the study area and time-series of precipitation and evapotranspiration. Observed discharge is typically used to evaluate model efficacy. The theoretical basis of TOPMODEL is fully reported in Beven and Kirkby (1979). Therefore only a brief description of the model is provided here. Figure 3.1 shows a representation of the TOPMODEL concepts.

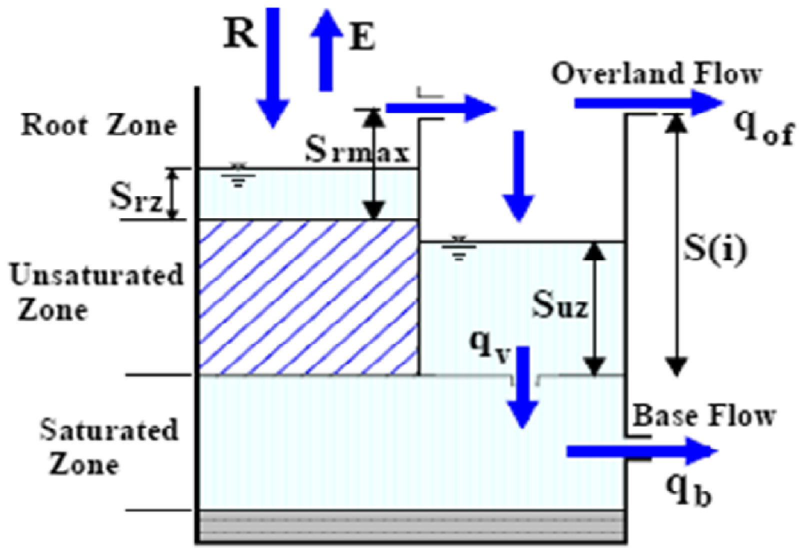


Figure 3.1. TOPMODEL concepts.

TOPMODEL governing equations

TOPMODEL is a combination of lumped and distributed sub models using soil-topographic characteristics and makes use of a topographic index of hydrological similarity based on an analysis of the topographic data which is described as :

$$\ln(a/\tan\beta) \tag{3.1}$$

where a is the cumulative area drained through a unit length of contour line, and $\tan\beta$ is the slope.

The subsurface flow rate per unit width of contour length, q_i, at any point on the hill slope is approximated by:

$$q_i = T_o \tan\beta_i e^{-S_i/m} \tag{3.2}$$

where T_o is the average soil transmissivity, β_i is the slope angle, S_i is the local storage deficit and m describes the change in transmissivity with depth. Based on this exponential approximation, the local deficit is derived as:

$$S_i = \bar{S} + m[\Lambda - \ln(a/\tan\beta)_i] \tag{3.3}$$

where $\bar{S}$ is the average storage deficit and Λ is the areal average of topographic index.

Equation 3.3 states that the saturation deficit at any point in a catchment is equal to the average saturation deficit for the catchment plus a soil parameter, *m*, times the difference between the average topographic index and the local topographic index. TOPMODEL does water balance accounting by keeping track of the "saturation deficit": the amount of water that one would have to add to the soil at a given point to bring the water table to the surface. This equation is used to predict the saturated contributing areas at each time step. A negative value of S_i indicates that the area is saturated and saturation overland flow is generated, while a positive value of S_i indicates that the area is unsaturated; and any water will infiltrate in to the ground and no overland flow will occur until the saturation store is satisfied.

Unsaturated zone calculations are made for each ln (a/tanβ) increment. The increments are topographic index classes. The calculations use two storage elements, SUZ and SRZ. SRZ represents a root zone storage, the deficit of which is zero at field capacity and becomes more positive as the soil dries out; and SUZ represents an unsaturated zone storage that is zero at field capacity and becomes more positive as storage increases. SUZ_i represents storage subject to drainage for the *i*th increment of topographic index. When $SUZ_i > 0$, vertical flow to the unsaturated zone is calculated as:

$$q_v = \frac{SUZ_i}{S_i t_d} \tag{3.4}$$

where the parameter t_d is a time constant.

The maximum value of storage in this zone is SR_{MAX}. The rate of evapotranspiration loss *E* is assumed be proportional to a specified potential rate Ep and the root zone storage SRZ as

$$E = Ep \times SRZ/SR_{MAX} \tag{3.5}$$

The sum of vertical flows weighted by the area associated with each ln(a/tanβ) increment is added to reduce the average saturated deficit $\bar{S}$. An outflow from the saturated zone, qb, is calculated as:

$$q_b = e^{-\Lambda} e^{S_i/m} \tag{3.6}$$

A water balance calculation for $\bar{S}$ produces a new end-of-time step value that is used to calculate the new value of S_i at the start of the next time step. There should be no water balance error involved since the incremental change in $\bar{S}$ is equal to the areally weighted sum of changes in the S_i.

The modeling processes are made for areal subdivisions of the catchment based on the ln(a/tan β) subdivisions. The generated runoff is routed to the outlet based on the assumption of constant kinematic wave velocity. All of the water balance accounting parts of the model is simple applications of the conservation of mass. A fuller description of TOPMODEL is available in Wolock (1993).

Over the past two decades, TOPMODEL concepts have been implemented with various computer languages upon different computer platforms (Beven, 1995). The tools developed have been used widely in the application of hydrological modeling in various catchments in the world (Beven, 1995). Various attempts have also been made to improve the concepts, such as uncertainty evaluation when using TOPMODEL for hydrological modeling.

TOPMODEL assumptions

The fundamental TOPMODEL equations are the continuity equation and Darcy's law. The basic assumptions that govern TOPMODEL are:

-The dynamics of the saturated zone can be approximated by successive steady state representations.

-The hydraulic gradient of the saturated zone can be approximated by the local surface topographic slope, $\tan\beta$.

-The groundwater table and saturated flow are assumed to be parallel to the local surface slope.

-The distribution of down slope transmissivity with depth is an exponential function of storage deficit or depth to the water table.

-Grid cells with the same topographic index are hydraulically similar.

TOPMODEL was chosen among a number of hydrologic models to illustrate the integrated simulation of surface water and groundwater. This is because TOPMODEL, a physically based watershed model that simulates the variable source area concept of stream flow generation requires few parameter files and is thus fairly easy to parameterize and run. This model requires a DEM and a sequence of rainfall and potential evapotranspiration data to predict the pattern of soil water deficit, as well as the resulting stream discharges. TOPMODEL is popular, as it provides computationally efficient prediction of distributed hydrological responses with a relatively simple model.

TOPNET Model

Many hydrologists have been working to develop new hydrologic models or to try improving the existing ones. Consequently, a plethora of hydrologic models are in existence today, with many more likely to emerge in the future (Singh, 1995; Singh and Frevert, 2002a, 2002b). As noted by Vieux (2001), with the advancement of the Geographic Information System (GIS), a class of models, known as distributed hydrologic models, has become popular. These models explicitly account for spatial variations in topography, meteorological inputs and water movement.

The hydrological model called TOPNET (Ibbitt et al., 2001; Bandaragoda, Tarboton, and Ross, 2004), selected for use in this research, is one such model. It is a semi-distributed rainfall runoff routing model based on TOPMODEL and kinematic wave routing in a river network. It keeps daily accounts of the following water balance component s of a catchment – precipitation, evapotranspiration, discharge to rivers, and change in soil water storage. The model monitors this in two parts: root zone water, which can be evaporated, and groundwater, which can only be evaporated if it is close to the ground surface. The model does not include deep aquifers. The discharge to rivers from each catchment is passed into a model of the river network. Water is routed along the network, and accumulates discharge from other catchments. The water eventually flows out of the end point of the spatial domain; there are no losses modeled from the river into groundwater systems. Figure 3.2 summarizes some of the basic concepts in the TOPNET model.

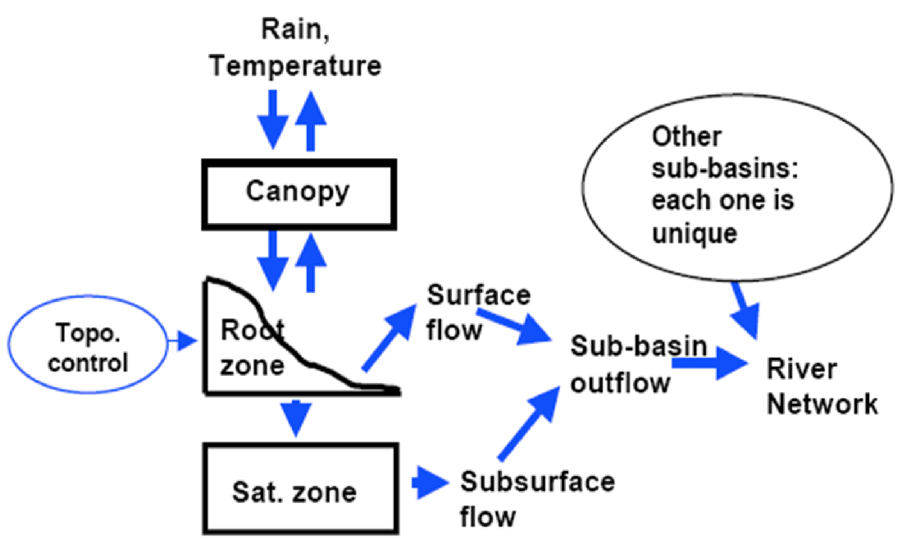


Figure 3.2. Runoff generation processes in TOPNET (Woods and Henderson, 2003).

The catchment precipitation is calculated from the precipitation at the grid points in and around the catchment. All precipitation becomes either surface runoff or infiltration, according to infiltration calculations. If the groundwater storage levels are high, then the catchment is saturated to the surface, and more surface runoff is generated. The proportion of this “saturated area” surface runoff varies with seasons. In addition, if the soil in the root zone is dry then more water can infiltrate. The model also takes account of the fact that as the groundwater levels rise closer to the surface, the soil which is near the saturated area is also getting wetter.

Evapotranspiration is calculated by first estimating a potential evapotranspiration given the temperature and day-length using the Priestley-Taylor approach, and then adjusting for the increase or decrease in evaporation due to vegetation and canopy cover characteristics. If the soil in the root zone is wet enough (water holding capacities are estimated from a soils database), then the actual value of evapotranspiration is the “potential evapotranspiration,” and if the soil moisture in the root zone is below “field capacity” then actual value is proportionately less than the potential value. If the soil is wet (above field capacity) then water drains to the shallow groundwater system. Water flows from the groundwater zone into streams. The more water there is in the groundwater system, the faster it flows into the streams. The flow in streams is routed through the river network using kinematic wave modeling illustrated in Figure 3.3.

Runoff from catchments is represented by blue arrows and it enters the river network either at the head of small streams or at the node points within the river network.

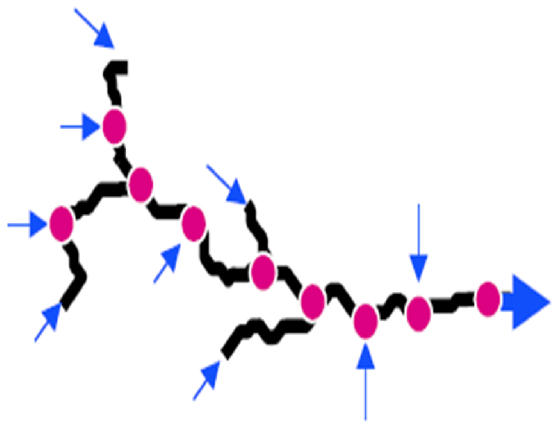


Figure 3.3. A typical river network in TOPNET (Woods and Henderson, 2003).

TOPNET model inputs

The main model driving inputs are precipitation and meteorological quantities (e.g. wind speed, minimum and maximum temperature) which are required to model the energy balance. These parameters are obtained from nearby climate stations and interpolated for each model element. Precipitation is separated into rain or snow based upon a temperature threshold. The snowpack is modeled using the Utah Energy Balance (UEB) model (Tarboton et al., 1995).

Rainfall is first input to the canopy interception model. Interception storage, S_i, is obtained by:

$$dS_i/dt = (P+I_r)\,(1-f(S_i)) - E\,C_r\,f(S_i) \qquad (3.7)$$

where P is the precipitation rate, I_r the irrigation application rate, Cr is an interception adjustment factor, E the reference evapotranspiration rate and $f(S_i)$ a function giving through fall depending on the amount of water held in interception storage (Ibbitt, 1971)

$$f(S_i) = S_i/CC(2–S_i/CC) \tag{3.8}$$

where CC is canopy capacity. The through fall is

$$T = P\ f(S_i) \tag{3.9}$$

Reference evapotranspiration demand not satisfied by evaporation of intercepted water is

$$E_p = E\ (1–f(S_i)) \tag{3.10}$$

These, together with snowmelt outflow,M, serve as the forcing for the root zone store, which represents the upper layer of soil depth where roots can extract water.

Root zone store component

Parameters describing the root zone store processes are the depth of the root zone store (d), saturated hydraulic conductivity (K), Green-Ampt wetting front suction (ψ_f), soil drainage parameter (c), drainable moisture ($\Delta\theta_1$), plant available moisture ($\Delta\theta_2$), and impervious fraction (f_i). Infiltration excess runoff and drainage to the saturated zone are influenced by the root zone store. In the root zone store the moisture content range is divided into the drainable moisture between saturation and field capacity, and plant available moisture. The soil parameters, except for the impervious fraction, are estimated using the Clapp and Hornberger (1978) soil textural relationships. The impervious fraction is determined from land cover and changes due to land use changes. Over the impervious areas infiltration is zero so surface runoff is maximum. In the pervious areas the state variable S_r quantifies the amount of water held in the root zone store and this is obtained from:

$$dSr/dt = I - E_s - R \tag{3.11}$$

where I is the infiltration rate, E_s is soil evapotranspiration rate and R the soil zone drainage rate or recharge to saturated zone. I, the infiltration rate is limited to be less than the infiltration capacity modeled with a Green-Ampt formulation

$$I_c = K\frac{z_f + \psi_f}{z_f} \tag{3.12}$$

The depth to the wetting front is estimated assuming all water in the root zone store occupies a saturated zone above the wetting front giving

$$z_f = \frac{S_r}{\Delta\theta_1 + \Delta\theta_2} \tag{3.13}$$

Unsatisfied evapotranspiration demand is given first priority when there is available surface water, so infiltration only occurs when there is excess surface water after evaporative demand has been met, i.e. $M+T-E_p$ is positive. When this excess water exceeds I_c, infiltration excess surface runoff is generated. Drainage from the soil zone is modeled when the moisture content is greater than the field capacity. The relative drainable saturation, S_{rd}, is defined as

$$S_{rd} = \mathrm{Max}(0, S_r - d\Delta\theta_2)/d\Delta\theta_1 \tag{3.14}$$

Unsaturated hydraulic conductivity is estimated as , $K\,S_{rd}^{c}$ and recharge to the saturated zone is obtained as:

$$R = K\,S_{rd}^{c} \tag{3.15}$$

Soil evapotranspiration is unlimited when soil moisture content is in excess of field capacity. Between field capacity and permanent wilting point, evapotranspiration reduces linearly to zero as the wilting point is reached. The relative plant available saturation is therefore defined as

$$S_{re} = \mathrm{Min}(1, S_r / d\Delta\theta_2) \qquad (3.16)$$

Evapotranspiration from soil moisture is called to fulfill evapotranspiration demand not met by interception and evaporation of available surface water. This is expressed as

$$E_s = S_{re}\,\mathrm{Max}(0, E_p - M - T) \qquad (3.17)$$

<u>Saturated zone component</u>

The saturated zone component is modeled using the TOPMODEL assumptions of saturated hydraulic conductivity decreasing exponentially with depth and saturated lateral flow driven by topographic gradients (Beven and Kirkby, 1979; Beven et al., 1995). Two important parameters are soil profile lateral transmissivity, T_o, and the sensitivity parameter, *f*, characterizing the decrease of hydraulic conductivity with depth.

Using these TOPMODEL assumptions, a state variable called the average depth to the water table or average soil moisture deficit, $\bar{z}\Delta\theta_1$, is introduced and this state variable is evolved according to:

$$\frac{d(\bar{z}\Delta\theta)}{dt} = -R + T_o e^{-\lambda} e^{-f\bar{z}} \qquad (3.18)$$

where λ is the spatial average of the topographic wetness index ln(*a/tanβ*). *a* is specific catchment area and *tanβ* the topographic slope. The parameters T_o and f are estimated based on relationships to soil texture from GIS soils data that represents texture at different depths (e.g. Clapp and Hornberger, 1978). The topographic variables, *a* and *tanβ* are evaluated using the Terrain Analysis using Digital Elevation Models (TauDEM) method developed by (1997).

As in TOPMODEL, the local depth to the water table is given in terms of the topographic wetness index as:

$$z = \bar{z} + (\lambda - \ln(a / \tan\beta)) / f \qquad (3.19)$$

The distribution of wetness index is represented using a histogram of wetness index classes with the proportion of area falling within each class recorded and depth to the water table calculated for each class. The depth to the water table is used to areas of surface saturation and the excess surface water input becomes saturation excess surface runoff. The depth to the water table in each class is also used to determine the parts of the model element where the saturated zone upwells into the soil zone which represents loss of water from the groundwater saturated zone.

MODFLOW Model

The code selected for this research is MODFLOW-96 (Harbaugh and McDonald, 1996). MODFLOW-96 is a multi-dimensional, finite-difference, block-centered, saturated groundwater flow code which is supported by enhanced boundary condition routines to handle recharge, evapotranspiration and streams (Prudic, 1988). The single most important advantage of using MODFLOW is its modular structure which means that specific packages can be used for specific problems of interest. As stated by Harbaugh and McDonald (1996), other benefits of using MODFLOW include: 1) MODFLOW incorporates the necessary physics represented in the conceptual model for flow for the study area, 2) MODFLOW is the most widely accepted groundwater flow code in use today, 3) MODFLOW was written and is supported by the USGS and is public domain, and 4) MODFLOW is well documented

A brief description of the model is given here. In MODFLOW, Darcy's law and the continuity equation are solved numerically using a finite difference technique. The three-dimensional movement of ground water of constant density through porous earth material may be described by the partial-differential equation:

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right)+W=S_s\frac{\partial h}{\partial t} \qquad (3.20)$$

where Kxx, Kyy, and Kzz are values of hydraulic conductivity along the x, y, and z coordinate axes, which are assumed to be parallel to the major axes of hydraulic conductivity; h is the potentiometric head ;W is a volumetric flux per unit volume representing sources and/or sinks of water, with $W<0.0$ for flow out of the ground-water system, and $W>0.0$ for flow into the system;SS is the specific storage of the porous material ; and t is time.

In general, S_s, K_{xx}, K_{yy}, and K_{zz} may be functions of space ($S_s = S_s(x,y,z)$, $K_{xx} = K_{xx}(x,y,z)$, and so forth) and W may be a function of space and time ($W = W(x,y,z,t)$). Equation 3.20 describes groundwater flow under non equilibrium conditions in a heterogeneous and anisotropic medium, provided the principal axes of hydraulic conductivity are aligned with the coordinate directions. Equation 3.20, together with specification of flow and/or head conditions at the boundaries of an aquifer system and specification of initial-head conditions, constitutes a mathematical representation of a ground-water flow system. A solution of equation 3.20, in an analytical sense, is an algebraic expression giving h(x,y,z,t) such that, when the derivatives of *h* with respect to space and time are substituted into equation 3.20, the equation and its initial and boundary conditions are satisfied. A time-varying head distribution of this nature characterizes the flow system, in that it measures both the energy of flow and the volume of water in storage, and can be used to calculate directions and rates of movement.

Except for very simple systems, analytical solutions of Equation 3.20 are rarely possible, so various numerical methods must be employed to obtain approximate solutions. One such approach is the finite-difference method, wherein the continuous system described by Equation 3.20 is replaced by a finite set of discrete points in space and time, and the partial derivatives are replaced by terms calculated from the differences in head values at these points. The process leads to systems of simultaneous linear algebraic difference equations; their solution yields values of head at specific points and times. These values constitute an approximation to the time-varying head distribution that would be given by an analytical solution of the partial-differential equation of flow.

Preparation of a groundwater model based on the MODFLOW code requires:

1. Creation of three-dimensional model grid (x, y,z) cells of (**x**, **y**, **z**) size.
2. Creation of surface elevation model from DEM.
3. Description of hydraulic conductivity, porosity, specific yield and specific storage

4. Description of flow boundary conditions such as Constant head and River head

5. Placement of pumping or recharge wells with pumping rates and time-series.

6. Description of recharge, evaporation and no flux boundaries.

Steady state and transient models

MODFLOW can be run for “steady state” or “transient” simulations. A steady state simulation represents a cross section in time, and produces one array of hydraulic head values for every cell. The model does not run for a given length of time, but it runs until the system reaches equilibrium, and the residuals have converged given the criterion specified in the Solver Package (McDonald and Harbaugh, 1988). A steady state simulation is usually performed during the calibration procedure to develop an optimal parameter set. The optimal parameter set is then used for a transient simulation to solve a time-dependent problem, and the data are verified with available data for that time period. Sometimes the transient model will need to be recalibrated to better match the observed data available for the transient simulation; however, if the steady state model is adequately calibrated, usually the recalibration procedure only requires fine-tuning of the calibrated parameters, if necessary.

Limitations of MODFLOW

There are limitations inherent in MODFLOW, just as in any simulation model. MODFLOW is strictly a saturated zone model, and does not model the unsaturated zone. The recharge value is assumed to be the actual volume of water that recharges the aquifer directly.

Numerous problems involving ground water flow modeling of real-world situations exist because the data necessary for the direct or inverse solutions is often not available. Head distribution is never known exactly because measurements do not exist at all points and in some cases, not correct. Estimates of parameters are obtained in cases by spot measurements, which in some cases are too few for use in regional ground water flow models. Modeling problems in ground water hydrology mostly involve an incomplete combination of data and error, and error propagation is an important consideration. Some important sources of random errors in water level data with respect to ground water models are:

(i) Areal ground water assume that the head used is the average over the vertical but wells may not be screened over the entire interval modeled.

(ii) Hydraulic conductivity varies from point to point, which causes water levels to vary from values where hydraulic conductivity is uniform.

(iii) Measurement of well head elevation may be in error. Errors of several feet are common in water level surveys and in some cases due to interpolation.

In some cases, groundwater flow modeling is limited by errors in parameter data. These errors include:

(i) Too few data estimates of parameters to compute stable estimates

(ii) Results of point sampling are often biased because a large amount if data does not necessarily allow computation of nearly true or effective values of parameter and its variance. An example is permeability values from core analysis often are not representative of regional values because flow through large fractures is not reproduced by core analyses.

(iii) Transmissivities estimated from specific capacity data collected can have errors such as mismeasured water levels and clogging of screens. Wells are also normally drilled in favorable locations and screened only at the most productive zones

However, despite these limitations, MODFLOW is the most commonly used ground water simulation model. It has been applied to many situations with reliable results. Thus it was selected for this research

CHAPTER 4

TESTING A COUPLED TOPMODEL-MODFLOW MODEL IN A SMALL LOWLAND WATERSHED IN WASHINGTON STATE, UNITED STATES OF AMERICA

Introduction and Objectives

In recent years, floodplain and lowland catchments have been subject to fast-changing conditions, altering between agricultural land use (including the installation of widespread drainage systems and intensive fertilization) on the one hand and nature conservation areas on the other hand (Sophocleous, 2002; Mohrlock, 2003). It has thus become more important to focus on the sustainable use of lowlands and wetlands and to promote the natural regulation functions for the water balance of lowland catchments (Krause and Bronstert, 2004). To improve and manage the water quality of a lowland river system and to promote its natural regulation functions, it is necessary to investigate the water balance controlling functions of the strongly connected surface water and groundwater systems in the lowlands and its changes as a result of management practices (Winter et al., 1999; Sophocleous, 2002; Acreman, King, and Brown, 2003).

In several previous studies, the coupled interactions between surface waters and the groundwater of the adjacent lowland watersheds have been mentioned (Hayashi and Rosenberry, 2002; Sophocleous, 2002). The importance of the interactions between the shallow groundwater and surface waters for water balance processes of floodplains and wetlands in lowland areas (Waddington, Roulet and Hill, 1993; Devito, Hill, and Roulet, 1996; Andersen, 2004), and subsequently for floodplain ecology (Brunke and Gonser, 1997; Gasca-Tucker and Acreman, 2000; Hayashi and Rosenberry, 2002), have been investigated for numerous differently scaled streams and catchments. The characteristics, intensity and direction of groundwater– surface water interactions are controlled by pressure head gradients, hydraulic permeability of the hyporheic zone, and by the riverbed geometry (Winter et al., 1998; Winter, 1999; Woessner, 2000; Sophocleous, 2002).

As a result of the spatial heterogeneity of the controlling factors of groundwater–surface water interactions and the subsequent variability of the impact of these interaction processes, the watershed water balance is also characterized by highly variable spatial patterns and temporal dynamics (Cey et al., 1998; Langhoff Heidemann, Christensen, and Rasmussen, 2001; Sophocleous, 2002). However, spatially detailed studies concerning the temporally and spatially variable effects of controlling functions on the characteristics and intensity of groundwater–surface water interactions have been limited to the investigation of cross sections or small stream reaches (Langhoff Heidemann, Christensen, and Rasmussen, 2001). Although groundwater–surface water interactions have been qualitatively described for different scales, analyses of their temporally and spatially variable impact on water balance are rare.

In this study, we formulated a model in which the interaction between saturated and unsaturated storage is taken into account. The rainfall runoff model, TOPMODEL, is coupled to the MODFLOW groundwater model and used to simulate the interactions between the surface and subsurface water systems in a watershed. In most cases hydrological models are tailor-made to simulate either surface water dynamics or groundwater dynamics. Surface water models have rudimentary groundwater simulation routines while groundwater models have rudimentary surface water simulation procedures.

TOPMODEL was not designed to accurately simulate groundwater dynamics and makes simplifying assumptions regarding it: that the saturated zone is in equilibrium with a steady recharge rate over an upslope contributing area, and the water table is almost parallel to the surface such that the effective hydraulic gradient is equal to the local surface slope. Therefore, a more accurate simulation can be achieved by incorporating the simulation of the groundwater dynamics performed by MODFLOW into the simulation of surface water runoff performed by TOPMODEL.

The study objective was to test the effectiveness of using the InCouple approach to couple surface and subsurface water models. It is a further objective of this study to investigate the temporally and spatially variable groundwater–surface water interactions and to analyze their variable impacts on the watershed water balance. Hydrologic data for the period 1968 to 1972 for the Tenmile watershed, in the Water Resources Inventory Area 1, of Washington State, USA was used in this study. The effectiveness of the coupled model is describing hydrological processes in the watershed was done by comparing it with results from the calibrated stand alone TOPMODEL and MODFLOW models.

<u>Study area description</u>

WRIA1 has 28 major watersheds. Figure 4.1 shows the location of WRIA 1 and its associated watersheds. The figure also shows the outline of Tenmile watershed within WRIA 1.

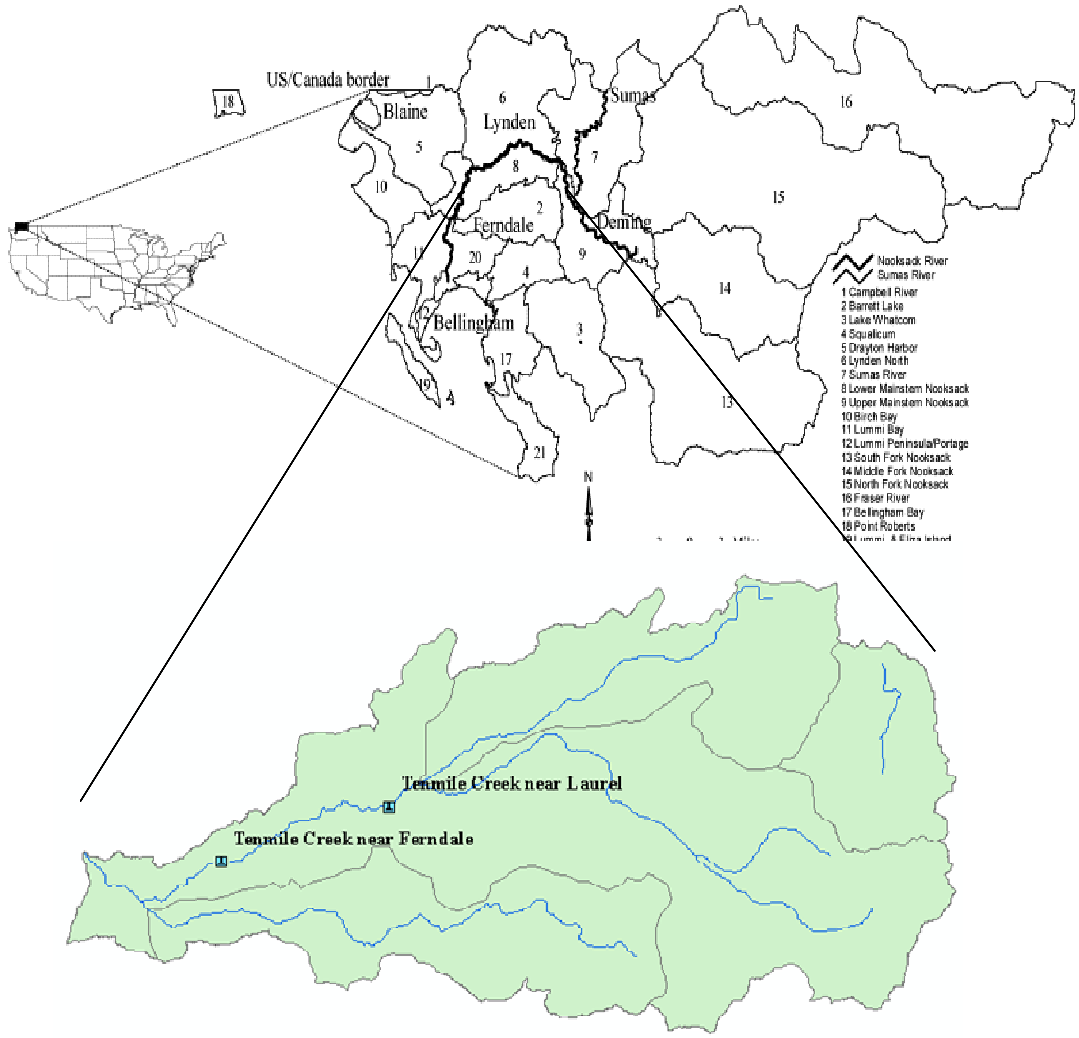


Figure 4.1. Location of Tenmile watershed in WRIA 1 (Adapted from Almasri and Kaluarachchi, 2004).

Tenmile Creek is a tributary of the lower Nooksack River, entering the Nooksack River near the town of Ferndale. Figure 4.2 shows a DEM of Tenmile watershed. Elevation in the watershed varies from 2.4 feet above means sea level in the west to 113 feet above means sea level on the south east of the watershed. Tenmile and its two tributaries, Fourmile and Deer Creek, drain a major portion of the Whatcom Basin lying south of the Nooksack River between the settlements of Strendell and Goshen to the east and Ferndale to the west.

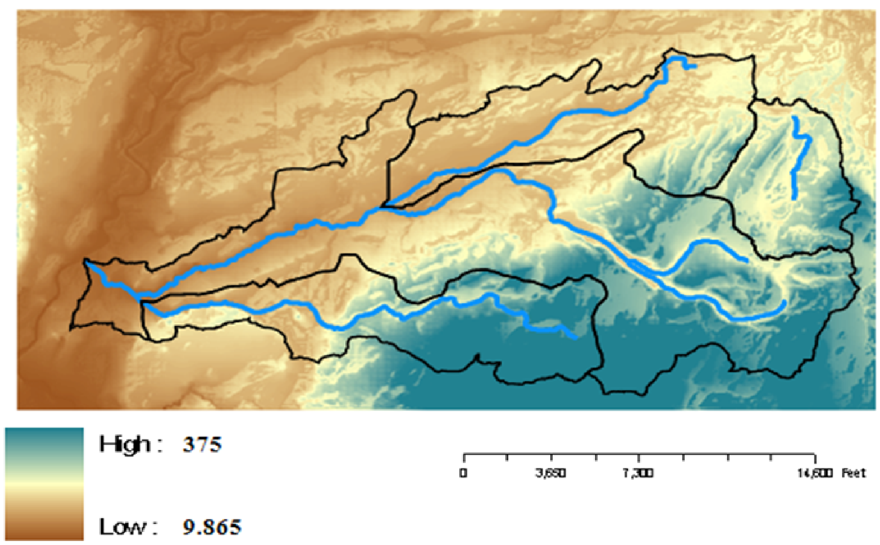


Figure 4.2. Digital Elevation Model for Tenmile watershed.

The watershed covers 35 square miles in area and consists of predominantly flat terrain with rolling hills along Deer Creek and along the upper portion of Tenmile Creek. Stream gradients are less than 0.5%, except for the headwater areas of the Deer and Tenmile drainages (WCCD, 1986).

Precipitation in the watershed ranges between 35 inches in the western end to 45 inches in the eastern part of the area. Seventy percent of the precipitation falls as rain between the months of October and March. April and September are the transition months between the wet and dry seasons. June, July, and August receive about 12% of the yearly average (WCCD, 1986).

Groundwater–surface water interactions in Tenmile watershed

Before embarking on modeling interactions between groundwater and surface water in a watershed it is important to first ascertain if the watershed displays that physical characteristic. Evidence of groundwater–surface water interactions can be observed from soil moisture data. Cox et al., (2005) studied groundwater–surface water interactions in the Fourmile Creek subwatershed of the WRIA 1 lowlands. Fourmile Creek subwatershed lies within the Tenmile watershed. As shown in Figure 4.3, Fourmile Creek originates from Green Lake and surrounding wetlands where it flows generally west to join Tenmile Creek just west of the Guide Meridian, which in turn flows into the Nooksack River near Ferndale. Thus results from this study by Cox et al. (2005) were assumed to be assumed to be representative of the hydrological conditions within the entire Tenmile watershed.

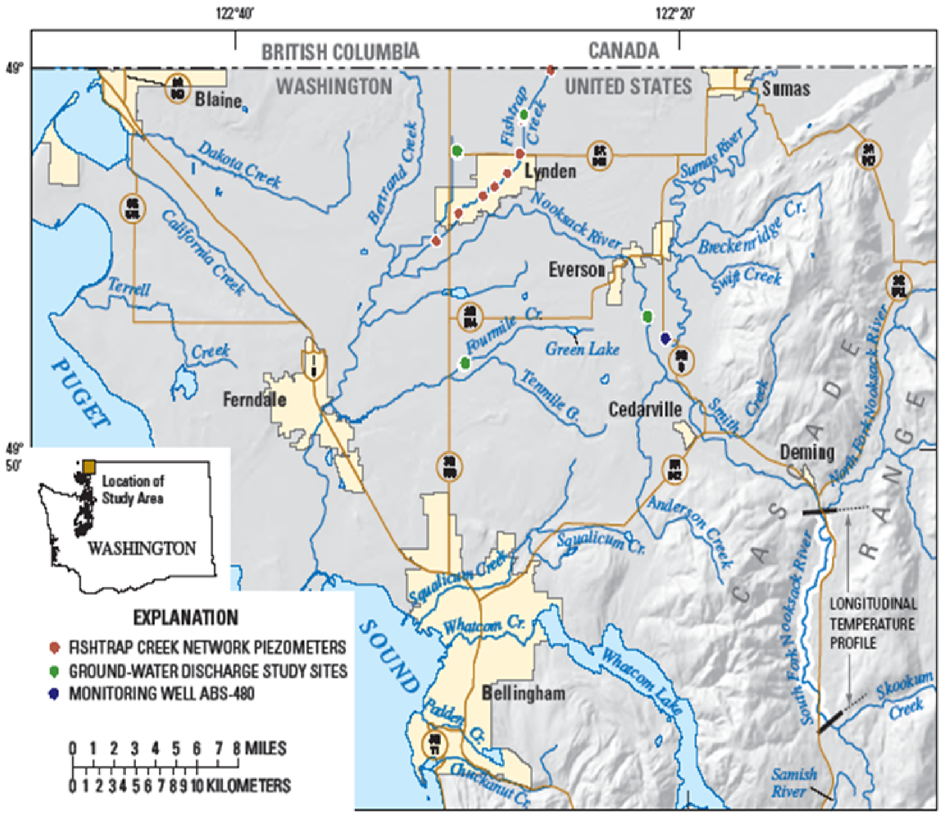


Figure 4.3. Location of Fourmile creek in Tenmile watershed (Cox et al., 2005).

An analysis of the temporal dynamics of the observed soil moisture for the period of December 2003–October 2004, showed that that the soil moisture dynamics were strongly linked to precipitation for most of the time; however, for certain events, the soil moisture dynamics cannot be determined by precipitation dynamics alone. In some cases, as in August 2004, one could observe a soil water increase without the occurrence of any rainfall. As stated by Cox et al. (2005), such an increase in the soil moisture is not explainable by any vertical infiltration of precipitation. This increase in soil moisture is due to an increase of the river water level. This shows that the river water dynamics also influence the unsaturated zone of the catchment, in particular in regard to the subsoil. In the same study of the Fourmile creek, two transects of three piezometers each were installed along the creek. The piezometers were installed on each bank and the middle of the creek for each transect. Vertical hydraulic gradients were measured in each piezometer with a manometer board four times from February to May 2004

Seasonal conditions may control the groundwater elevation and thus the direction of flow between the stream and aquifer. When the hydraulic gradient of the aquifer is towards the stream, groundwater discharges to the stream, and the stream is a gaining or effluent stream. When the hydraulic gradient of the aquifer is away from the stream, the stream is losing or influent. The rate of this water loss is a function of the depth of water, the hydraulic gradient towards the groundwater, and the hydraulic conductivity of the underlying alluvium. The channel system can be hydraulically connected to the aquifer, or have a leaking bed through which water can infiltrate to the subsurface. The extent of this interaction depends on physical characteristics of the channel system such as cross section and bed composition. Streams commonly contain a silt layer in their beds which reduces conductance between the stream and the aquifer.

In the Fourmile study, vertical hydraulic gradients were consistently positive. However the positive gradient tended to become smaller from February through early May 2004. Vertical hydraulic gradients also varied from one bank of the creek to the other suggesting the presence of groundwater fluxes. These fluxes are higher in areas with higher hydraulic gradients. This variation in hydraulic gradients also shows both the spatial and temporal variability in surface water–groundwater interactions at a field scale. Such variations are likely due to heterogeneities within aquifer materials underlying the streambed, local topography, the placement of tile drains, or other factors. Precipitation also may influence localized groundwater discharge.

A temporary gaging station was established at the Fourmile Creek site and continuous water level data were collected from December 2003 to October 2004 (Figure 4.4). Groundwater levels recorded in a piezometer installed to a depth of 5.4 ft below the streambed were consistently higher than surface water levels throughout the study period indicating that throughout the year there is movement of water from the saturated zone to the unsaturated zone manifested as stream flow. Vertical hydraulic gradients were upward, and generally were larger during winter and autumn than during the summer months. This evidence supports the presence of strong interactions between surface water and groundwater in the area.

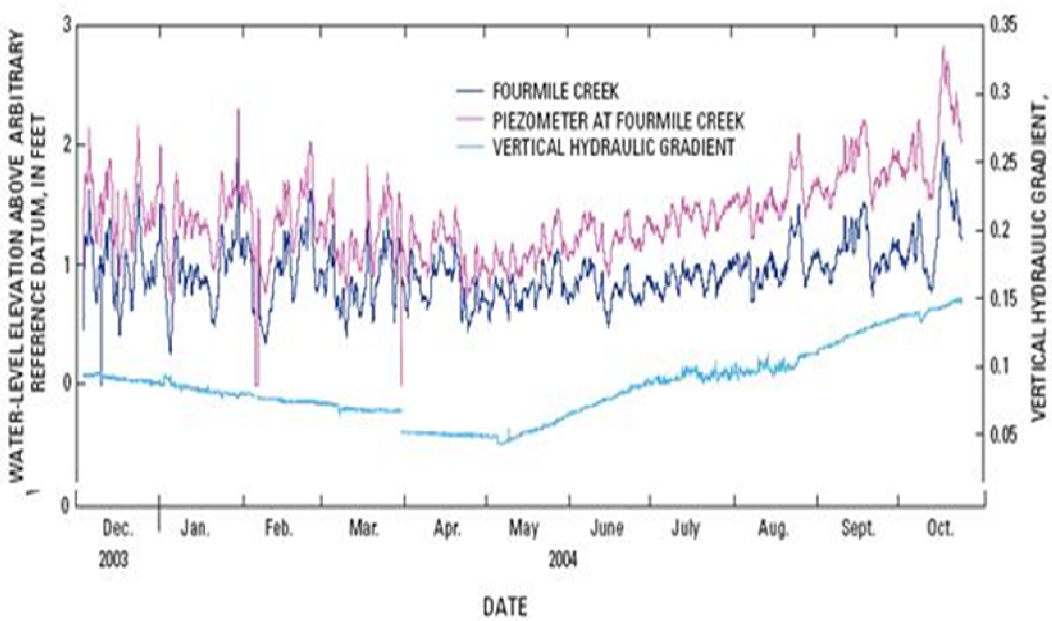


Figure 4.4. Water level and vertical hydraulic gradient data at Fourmile Creek near Guide Meridian, lower Nooksack River basin for period December 2003 to October 2004 (Cox et al. 2005).

Model Coupling Procedure

The model developed has two main components that have been coupled to allow a two-way interaction of water flow (i.e. feedback effects are taken into account in both directions). The two components are: runoff generation and vertical soil water dynamics are simulated by using the relevant routines of TOPMODEL and the flow in the saturated zone is modeled using MODFLOW.

The overall model coupling strategy is based on the understanding that the vertical groundwater recharge is derived by the simulation results of TOPMODEL and linked via a transfer function as percolation in the case of positive values or uptake in the case of negative values to the groundwater model MODFLOW. The coupling of vertical fluxes in and out of the unsaturated zone with the groundwater module is performed by transmitting the fluxes into/from TOPMODEL soil storage as groundwater recharge or uptake to MODFLOW, and vice versa (Figure 4.5).

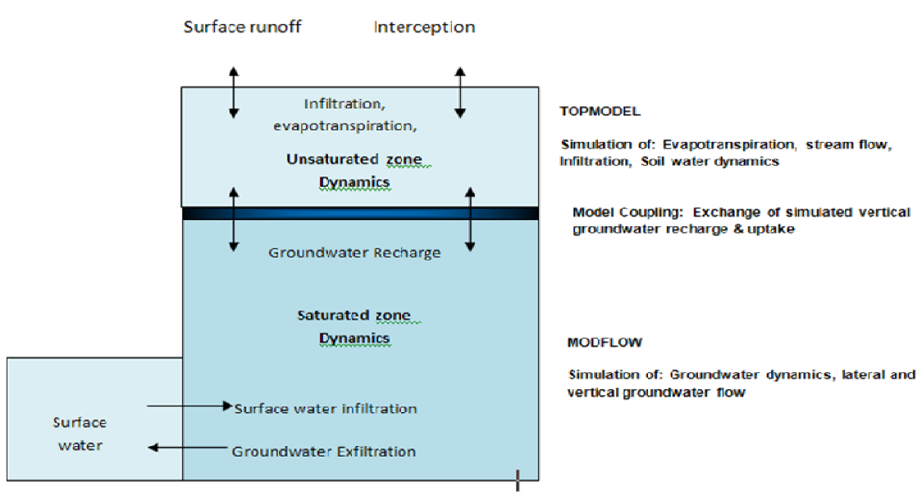


Figure 4.5. Concept of coupling the surface water and groundwater dynamics routines.

Tenmile Watershed Groundwater Model Using MODFLOW

Conceptual model

The purpose of a conceptual model is to simplify field problems and organize the associated field data so that the system can be analyzed readily (Anderson and Woessner, 1992). It also reflects our understanding of existing conditions. A simplified conceptual model of steady state recharge, movement, and discharge of groundwater was used to guide development of the numerical groundwater flow model of the study area. The groundwater system was conceptualized as a water table aquifer recharged by infiltration of precipitation and seepage of stream flow into stream reaches. Groundwater discharge was simplified in the conceptual model by considering only discharge to streams. Some groundwater also discharges to wells and other quarries and as evapotranspiration along the riparian zone; however, these sinks were not considered in the model.

The ground water model was constructed to provide an understanding of both the groundwater flow system (hydraulic head distribution and water budgets) and the major controls on the flow system. Groundwater flow was simulated using MODFLOW and operated under steady sate conditions and transient conditions. The steady state calculated heads and water budget components represent long term mean annual values.

Hydrogeological and aquifer characterization

The Tenmile Creek watershed is located primarily on the lower drainage basin of the Nooksack River and consists mainly of floodplains and hills. According to Vaccaro et. al (1998), there are four major hydrogeologic units delineated in WRIA 1. These are the Sumas–Blaine Aquifer, Everson-Vasion semi-confining unit, Vasion semi-confining unit, and the bedrock confining unit. Tenmile watershed falls within the Sumas-Blaine aquifer with the Everson Vashon fine grained layer outcropping as shown in Figure 4.6.

The Sumas Blaine aquifer is mainly composed of Sumas stratified sand, and gravel outwash of the Sumas and Noocksack rivers. It is mainly phreatic (unconfined) but has some locally confined areas in places overlain by lacustrine silt and clay. The aquifer is mainly stratified and has a minimum thickness of about 140ft. The outcropping Everson Vasion aquifer will mean different geological characteristics.

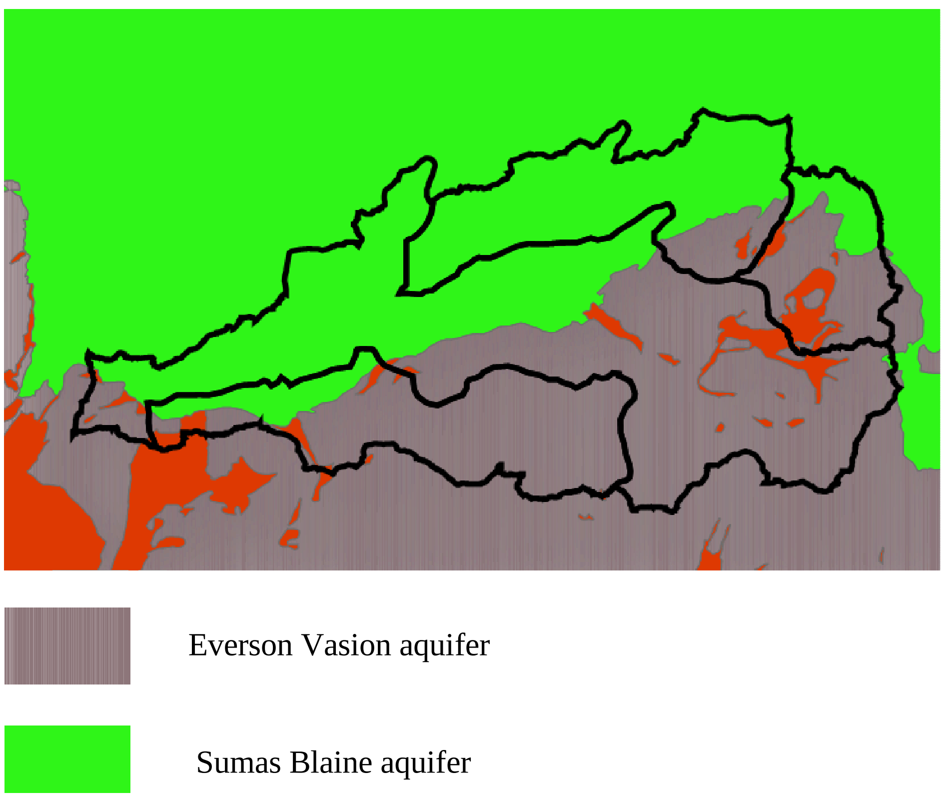


Figure 4.6. The Sumas Blaine and Everson Vashon aquifers in the Tenmile watershed (Vaccaro, Hansen, and Jones,1998)

Spatial discretization

The groundwater system was conceptualized as a water table aquifer and therefore considered to have only a single layer. A single layer was used in the groundwater flow model, as additional lithological information would be required to justify increasing the complexity. The watershed was divided into a regular grid. Each grid cell represented 300 ft in the x-direction and 300 ft in the y-direction. All cells outside the watershed boundary are inactive cells. Figure 4.7 shows the watershed with active cells (shaded) which all lie within the watershed boundaries. Elevation of the model grid was imported from the 90ft DEM for the watershed. The DEM was resampled to a 300ft grid.

Model time steps

A daily time step was selected for MODFLOW simulations. The choice of daily time steps was made considering limitations of time-step length on model convergence. Generally MOFLOW simulations are done using a monthly time steps. However, experience has shown that daily time steps are sufficiently small for MODFLOW to converge to a solution for most simulations

Model boundaries

Boundary conditions in the model define the locations and manner in which water enters and exits the active model domain. The general conceptual model for the Tenmile Creek watershed is that water enters the system as precipitation and exits the system as stream flow and groundwater discharge near the mouth of the watershed at Ferndale.

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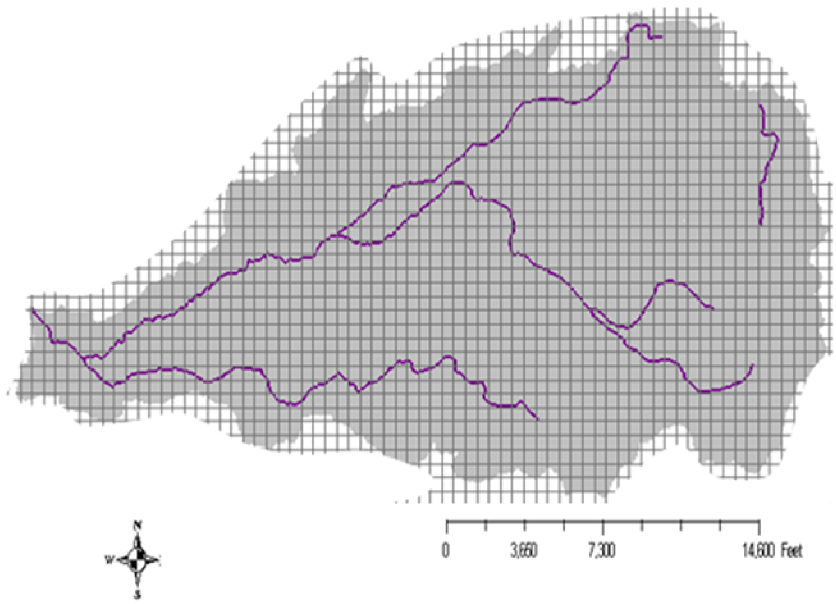


Figure 4.7. MODFLOW grid cells for Tenmile Creek watershed.

GeoEngineers (1998) and Golder Associates (1995) estimated most of the no flow boundaries using bed rock outcrops as their main criteria. Cox and Kahle (1999) also estimated the boundaries using ground water flow patterns. Areas where ground water flow is parallel to model boundaries are classified as no flow boundaries. Three types of boundaries were used in the Tenmile Creek watershed model: no-flow (outer model boundary), head-dependent flux (rivers, drains, and general head) and specified-flux (recharge).

The boundaries of the model coincide as much as possible with natural topographic, geologic, and hydrologic boundaries. Major topographic divides primarily define the lateral model boundaries. These natural features act as no-flow boundaries as they are considered coincident with groundwater divides. The topographic divides are either exposed bedrock or bedrock covered by a shallow layer of unconsolidated sediments. However there are sections of the watershed where a no flow boundary will not apply. As shown in Figure 4.8 there are areas to the northeast and southeast where the water table contours are not perpendicular to the watershed boundary and these can best be treated as specified flow boundaries. However due to the unavailability of flow data, the specified flow boundaries flux was set to zero.

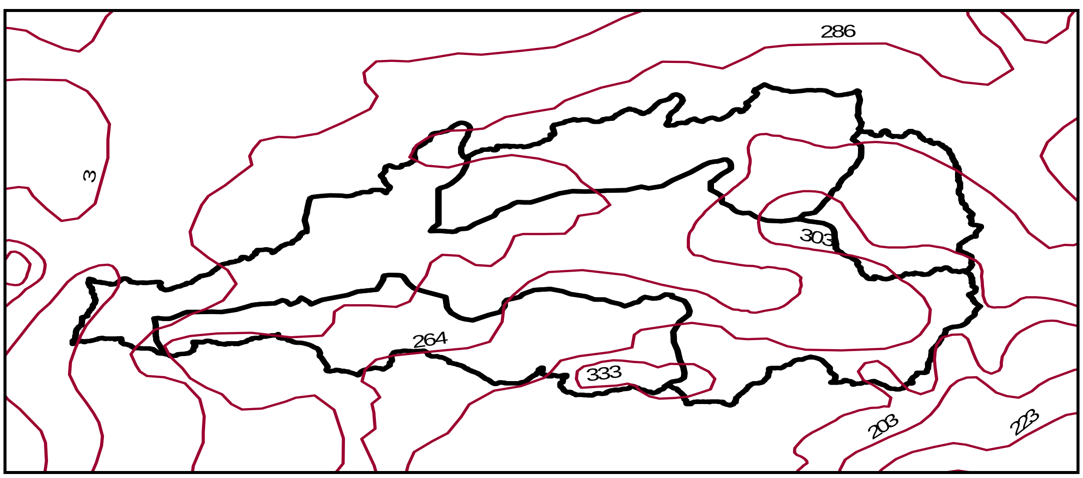


Figure 4.8. Steady-state water table contour map.

MODFLOW packages utilized

MODFLOW requires a variety of data as input, depending upon what packages are included in the simulation. For this study, the Basic Package, the Block-Centered Flow Package, the Output Control Package, the Recharge Package were used (McDonald and Harbaugh, 1988). The Evapotranspiration Package was not included in the coupled model because it was assumed that surface water model simulations would account for this part of the water budget. The Basic Package includes information about the aerial extent of the aquifers, and their location within the system (McDonald and Harbaugh, 1988). The input to this package defines the locations of the layers, and the aquifer boundaries. The Block-Centered Flow Package includes information about the hydraulic properties of each cell within the grid (model domain) (McDonald and Harbaugh, 1988). The properties include hydraulic conductivity, transmissivity, storage, and vertical conductance. The Output Control Package also does not require any physical data, but it simply controls the type and frequency of output for each simulation. The Recharge Package requires a user-defined recharge flux for each cell within the model domain. This value can be set equal to zero if no recharge is occurring in a given location.

Groundwater model parameters

Hydraulic characteristics. On average, the Sumas outwash deposits have horizontal hydraulic conductivities ranging from 7 to 7800 ft/day (Tooley and Erickson, 1996). Pump tests performed in the Strandell well field to the south of Sumas City gave values for transmissivity (T), specific yield (Sy), and horizontal hydraulic conductivity. These are shown in Table 4.1.

Table 4.1. Groundwater hydraulic parameters

Parameter	Average Value
Hydraulic conductivity	130ft/day
Transmissivity	2420 sq ft/day
Specific Yield	0.2

Recharge. Recharge to the model consisted of infiltration from precipitation. Mean annual recharge was used as a boundary condition for the model. In general, precipitation based recharge varies spatially with land-surface permeability, which is a function of soil characteristics and land use. Total flux to each cell is obtained by multiplying the recharge rate by the area of the cell.

Vaccaro, Hansen, and Jones (1998) estimated recharge values for the Puget Sound aquifer system which encompasses Tenmile watershed. They used linear regression between precipitation and ground water recharge. Data for estimating the regression equations were obtained from precipitation and recharge estimates of 26 small watersheds within the aquifer. The recharge estimates were obtained from previous studies using a deep percolation model and the Hydrogeological Simulation Program – Fortran (HSPF). Annual recharge estimates were adjusted based land use and land cover. These estimates were used in this model. Using GIS tools, Almasri and Kaluarachchi (2004) obtained polygon shapefiles of ground water recharge distribution for WRIA 1 from Vaccaro, Hansen, and Jones (1998). The annual recharge estimates were adjusted based on land use and cover. Each polygon in the GIS shape file corresponds to an area of WRIA 1 with a specific recharge value. For the Tenmile watershed, the recharge values varied from 11 to 25 inches/year. A mean annual recharge of 18 inches/year was used in the steady state model.

Water withdrawals and discharges. Pumping wells were simulated with the MODFLOW Well Package. Withdrawals from pumping wells were simulated as specified flows from the aquifer. Flow rates in the steady state model were assumed equal to average annual withdrawal and discharge rates for the study period 1968 to 1972.

Another simplifying assumption was that all simulated wells were in the single layer. Figure 4.9 shows the distribution of wells within Tenmile watershed.

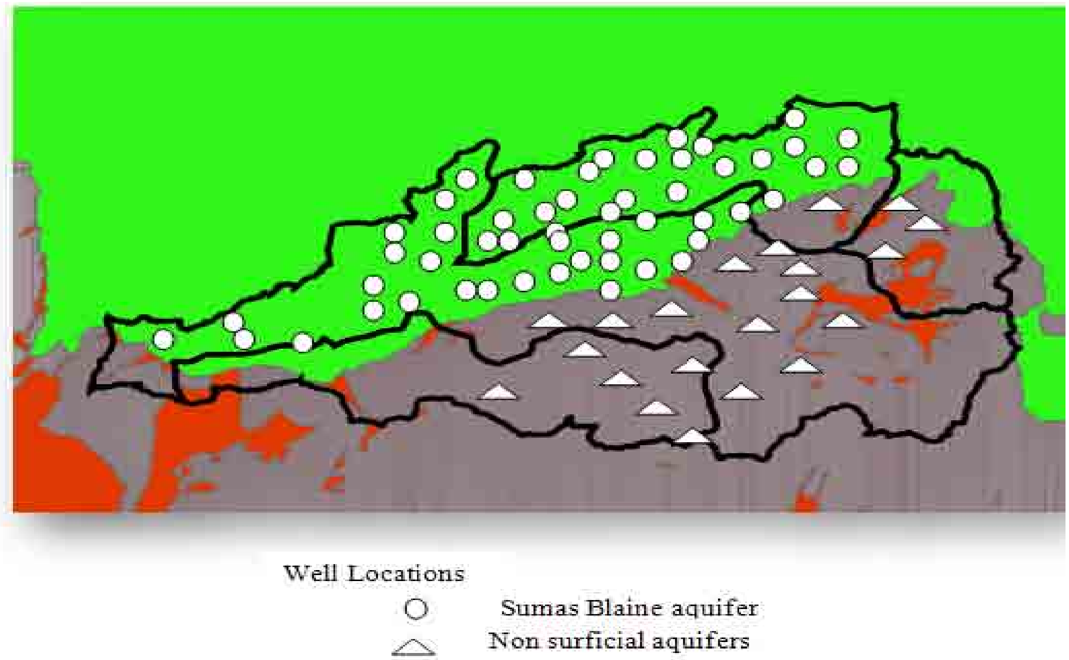


Figure 4.9 Well distribution in Tenmile watershed.

Model calibration

Two basic steps were followed in modeling the aquifer: a steady state model was developed to determine the spatial distribution of hydraulic conductivity, and a transient model was run for the period 1968 to 1972 by using monthly recharge and pumpage data. Spatial distribution of hydraulic conductivity was determined using trial and error procedures. Specific yield was calibrated using the transient model.

Steady-state model calibration

A steady-state model was developed to determine the distribution of hydraulic conductivity during average discharge using average recharge as input to the model. Measured water levels for the period May 1968 to May 1970 were used to evaluate the steady-state model calibration. There are seven calibration wells within the watershed. However, only five wells have data that could be used for calibration. The location of the calibration wells is shown in Figure 4.10.

Transient model calibration.

Transient models were developed to simulate the variations in hydrologic conditions within an average annual cycle. The transient models are based on the steady-state models but incorporate time-varying hydraulic stresses and boundary conditions. The spatial discretization of the model grid, boundary conditions other than specified flows, and spatial variations in stresses and hydraulic conductivities are the same in transient and steady-state models. The transient models were calibrated by comparing water levels to average monthly levels estimated for the 1968–70 period.

With the transient model, the low-flow periods of the annual cycle can be simulated and these are of particular concern in the evaluation of the effects of water-management alternatives. During these periods, the effects of water withdrawals and other management practices on aquatic life and stream-water quality often are greatest, because their effects are combined with naturally low flows and ground-water levels. Water demands also typically are highest during summer months.

Simulated heads and the calibrated distribution of horizontal hydraulic conductivity from the steady state model were used as input for the transient model. Monthly stress periods were used for transient simulations, which resulted in a total of 60 stress periods for the 5-year simulation. The initial estimate of specific yield of 0.01 was based on data from Slade et al., (1985). A specific storage value of $4.5 * 10^{-6}$ ft^{-1} was used. This was adopted from the study by Vaccaro, Hansen, and Jones, (1998).

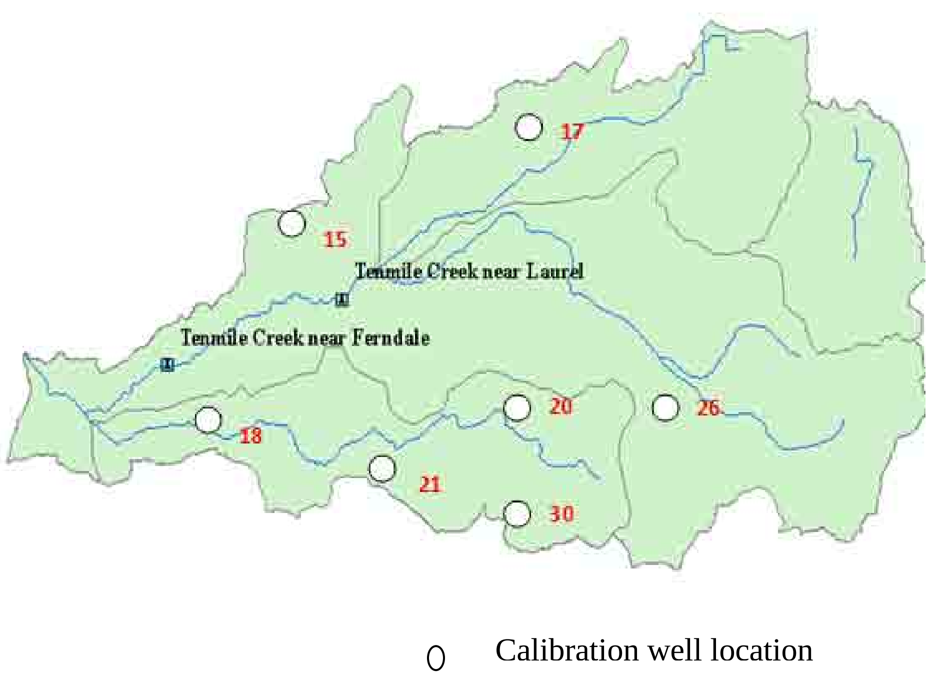


Figure 4.10. Location of calibration wells in the watershed.

Rainfall Runoff Modeling Using TOPMODEL

Surface water flow characteristics and baseflow

The surface water system of the WRIA 1 lowlands has been extensively altered by people. In its natural condition, large areas of the lowlands were swampy. To make these lands inhabitable and conducive to agriculture, people have installed drainage systems to lower the water table and dry the land since farming by settlers started in the area, in about 1850. Parts of the drainage systems consist of open ditches, while other parts consist of underground structures.

Other alterations to the surface water system include the diking and redirecting of the Nooksack River, to minimize damage from flooding that occurs periodically. There are two USGS gaging station on Tenmile Creek: Station 12212900, Tenmile Creek at Laurel; and 12213000, Tenmile Creek near Ferndale further down stream. Table 4.2 shows the location of the two gages and the available stream flow records for each gaging station. The stream flow time series plot for station 12213000 is shown in Figure 4.9. Streamflow data for station 12212900 could not be obtained.

Table 4.2. Stream flow gauging points in Tenmile Creek watershed

Station Number	Station name	Latitude	Longitude	Drainage Area (sq. miles)	Period of record
12212900	Tenmile creek at Laurel	485149	1222945	22.7	1968-1972
12213000	Tenmile creek at Ferndale	485115	1223225	23.6	1954

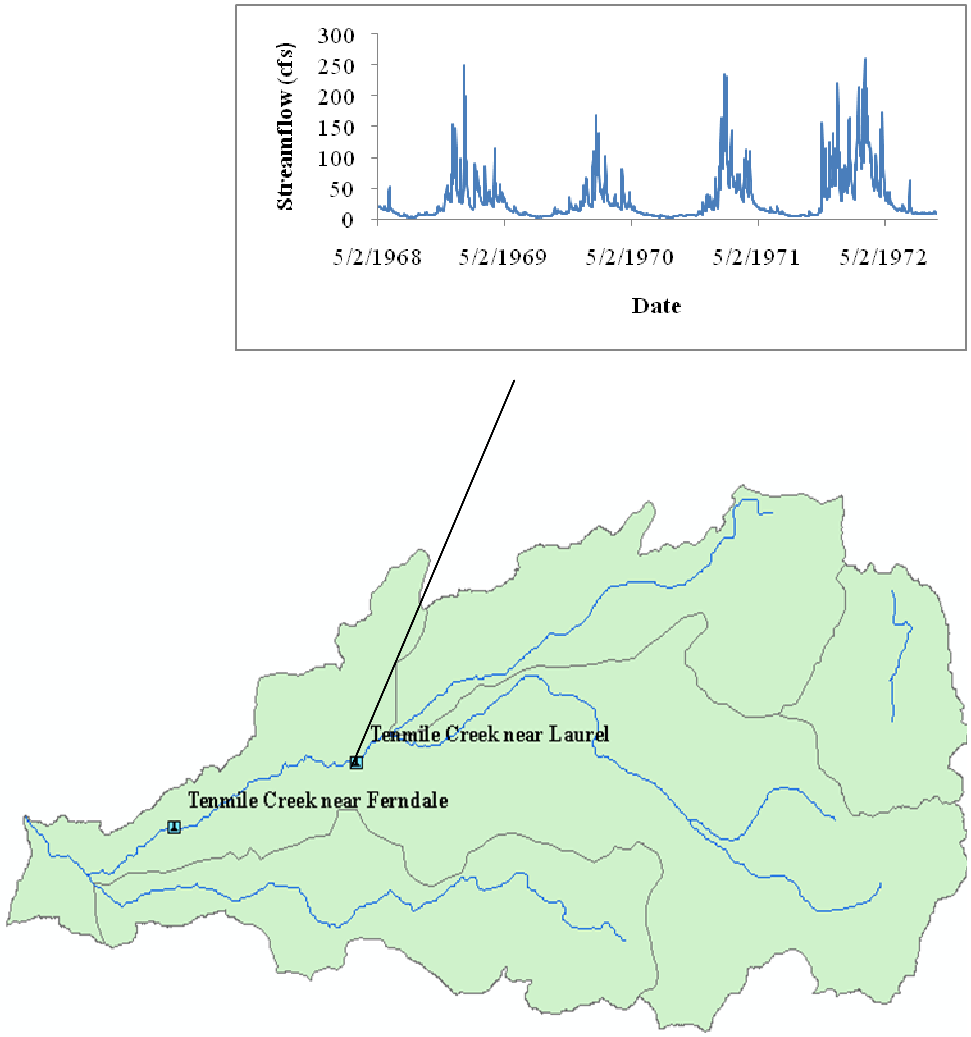


Figure 4.11. Tenmile water shed and the location of stream flow gaging stations.

Parameterizing TOPMODEL for Tenmile watershed

In this research, we used TOPMODEL, a rainfall–runoff model. The model uses gridded digital maps of land surface topography and a dynamic numerical framework that accounts for the movement of water within the soil and at the surface. The mean depth of the water table and the topographic index are used to compute the saturated areas of the watershed and the shallow groundwater flow that supports it. Thus, at any point in time a mosaic of cells, each with a local model surface wetness which, taken as a whole, represents the surface conditions of the entire watershed. This surface wetness depicts the spatial variability of conditions at the land surface that result from terrain and integrated weather based forcing inputs.

TOPMODEL is designed for the simulation of rainfall/runoff response times and prediction (both spatial and temporal) of saturated zones in the watershed. In this study the Tenmile watershed is modeled as a single watershed with no subwatersheds. Parameter values were optimized to obtain an adequate calibration, as opposed to being assigned by measurement or estimation. Initial parameters used for Tenmile watershed are shown in Table 4.3. These parameters are described in Equations 3.1 to 3.6.

Table 4.3. TOPMODEL Hydrological parameters

Parameter	Value
m	0.007
Ln(To)	0.10
SRmax	0.07
SRinit	0.01
ChVel	39

Topographic index

TOPMODEL requires a topographic index distribution function for the watershed. The topographic index values were determined using the procedure referred to as GRIDATB which is a subroutine in TOPMODEL. A detailed discussion of the procedure is given in Quinn et al., (1995). GRIDATB uses a multiple flow direction algorithm that requires a digital elevation model of the catchment. For this research, the procedure grouped the topographic index values into 14 classes as shown in Figure 4.12.

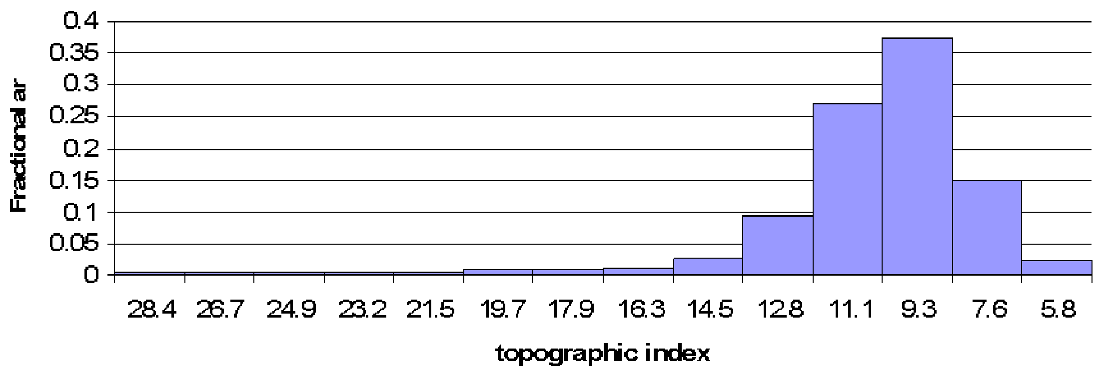


Figure 4.12. Distribution of TOPMODEL's topographic index for Tenmile watershed.

Climatic data

The data necessary to implement TOPMODEL are a DEM, the stream flow data, evapotranspiration, and the precipitation data for the study period. There is no climatic station with long-term climatic data within the watershed. The nearest climate stations with sufficiently long record are that of Bellingham 2N and Bellingham FCWOS, AP. For precipitation, the average readings of the two stations are used in this study.

Channel velocities and distance area data

Channel velocity refers to the velocity within a stream segment or within a distance increment. Routing velocity refers to the velocity that translates runoff from a given location to the catchment outlet. The average channel and routing velocities are estimated by determining equivalent stream slopes for each subcatchment using lengths of stream segments and slopes of each stream segment (Ponce, 1989).

The approximate average velocities for different stream types based on the stream slope can be found in the hydrologic literature (Chow, Maidment, and Mays, 1988). The distance area data are used to determine the routing velocity time to the outlet. The streams in the catchment are divided into equal distance segments. An approximate average velocity is determined, and this routing velocity is assumed to be constant in all the stream segments. At the catchment outlet, the distance and the fractional contributing area are zero. Thereafter, each distance and associated fractional area given are cumulative as one moves to the catchment divide upstream from the outlet, with the final accumulated fractional area being equal to one.

Calibration and verification.

The model is calibrated using a trial and error procedure for the study period. The criteria used in the calibration process determine the parameter set yielding the highest Nash and Sutcliffe efficiency value (Beven et al., 1995) defined as:

$$E=1-\left[\sum_{i=1}^{n}\left(Q_{obs}-Q_{i}\right)^{2} / \sum_{1=1}^{n}\left(Q_{obs}-Q_{av}\right)^{2}\right] \quad (4.1)$$

where Q_{obs} is the observed stream flow, Q_i is the simulated stream flow, Q_{av} is the average observed stream flow and n is the total number of time steps used in the flow simulation. According to Equation (4.1), if the simulated and observed flows are equal, the value of E will be equal to 1 or 100 percent.

The two TOPMODEL parameters, m and ln T_o, are utilized for calibration. To calibrate the single catchment model, all parameters are assigned initial values. Starting with m, the value is varied, holding all other parameters at the values previously set, to determine which value of m yields the highest Nash and Sutcliffe efficiency value, E. With this value of m and all other parameters held constant, ln T_o is varied with an effort to further maximize this efficiency. The other parameters are then varied until a final parameter set is obtained that shows little or no improvement in the efficiency, E. The model was calibrated using data for the years 1971 and 1972.

TOPMODEL- MODFLOW Coupling

Model coupling methodology

TOPMODEL and MODFLOW are both distributed, time-dependent models, written in Fortran, that possess the traditional input-solve-output model code structures shown in Figure 4.9. Both models begin by reading their input parameters and boundary conditions, then execute a time step loop, writing their results after each step. The original version of TOPMODEL executes the entire time step loop for each subcatchment.

Figure 4.13 depicts the overall control flow of the program with respect to potential coupling points. Each light colored arrow indicates the flow of control, and the dark arrows indicate potential coupling points/interfaces (PCI), or places where the values of state variables (the variables that represent the state of the physical quantities being modeled) can be exchanged with other models. In this coupling, we first determined the interactions between the physical systems which are to be studied and where these interactions occur. The runoff from a catchment is heavily influenced by the rainfall over the catchment, as well as the groundwater beneath the catchment. When simulating the runoff from a catchment, TOPMODEL makes very simple assumptions about the behavior of the groundwater beneath it.

In this coupling, the idea is to replace TOPMODEL's simple groundwater head calculations with MODFLOW's full simulation of the saturated zone. The physical quantities involved are the water table head of the aquifer, and the unsaturated zone–saturated zone flow (recharge). These quantities influence each other along the lower boundary of the catchment's unsaturated zone (TOPMODEL's coupling surface) and along the upper boundary of the aquifer (MODFLOW's coupling surface). In this case, the two dimensional coupling surfaces of each model are superimposed vertically.

The next step was to determine the state variables representing the physical quantities involved in the interaction. The water table height of the aquifer is represented by MODFLOW's h_{new} array. In TOPMODEL, the recharge between the catchment and the aquifer is represented by TOPMODEL's q_{uz} variable.

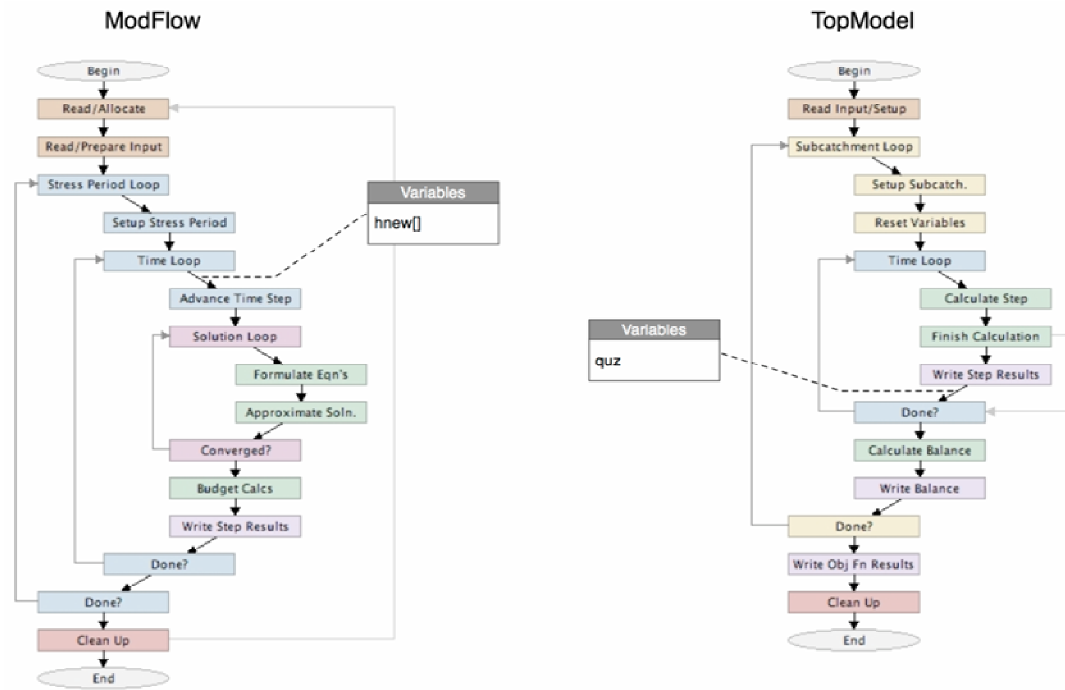


Figure 4.13. Overall control flow diagram for TOPMODEL and MODFLOW.

An analysis of the two models shows an incompatibility in that MODFLOW's water table height variable, h_{new}, is spatially distributed over a regular grid, while TOPMODEL's recharge variable, q_{uz}, is spatially distributed over a set of irregularly shaped subcatchments. In order for these quantities to interact, their variables must be mapped to the same space. Here, each MODFLOW cell must be mapped to the TOPMODEL subcatchment in which it is located.

Next, the locations within each model code where the variables should be accessed for communication with the other model must be identified. From the PCI, we know where it is permissible to access variables within the individual model codes, but not how those access points line up between the codes. Different access points within a model code correspond to different points in simulation time. Data exchanges must happen at points corresponding to the same (or at least coordinated) points in simulation time. The next step was, therefore, to determine where in the model codes, the data exchanges between state variables should take place. It is clear from the two PCIs that simply accessing state data at the start of each time step would not work because TOPMODEL's time loop is within a spatial loop. As a result, each TOPMODEL time step is simulated multiple times (once for each subcatchment in a situation where there is more than one subcatchment), while each MODFLOW time step is simulated only once.

Furthermore, because groundwater moves much more slowly than surface water, the length of a MODFLOW time step is longer than the length of a TOPMODEL time step. These structural differences represent another incompatibility between the models. One way to resolve them is to have TOPMODEL simulate all the subcatchments for a small set of short time steps for each long time step of MODFLOW. In such a situation, the h_{new} variable of MODFLOW would be accessed at the start of each time step, and the q_{uz} variable of TOPMODEL would be accessed after the subcatchment loop. As shown in this example, the models often need some control that was not in their original code; in this case, TOPMODEL simulations are conducted over several time steps of shorter duration corresponding to a single time step loop of MODFLOW.

The final step is to specify precisely how the state variables affect each other. Often the values of the state variables must be transformed before they can be exchanged. These transformations are functions, specified by the scientist, that compute new values for variables based on the values of variables from both models. If the data is spatially distributed as part of this function, it may be necessary to map the simulation spaces to each other.

Coupling specification

Figure 4.14 describes how MODFLOW and TOPMODEL are coupled. The figure shows where the interactions take place with respect to the model PCIs. There are two important things to note about this coupling. Firstly, since the original TOPMODEL uses an outer spatial loop and an inner temporal loop, the whole TOPMODEL simulation must be executed for each time step of MODFLOW. Therefore, an extra outer loop is needed in TOPMODEL to allow the full simulation to be executed repeatedly.

Secondly, the coupler must support spatial aggregation and de-aggregation of the exchanged data. This is because MODFLOW simulates the full spatial extent on each time step, while TOPMODEL simulates only a single subcatchment on each time step.

Using the coupling points from Figure 4.14 the sequence of events performed by the integrated model:

1. The coupler begins by executing each model. The models start up and read their input parameters exactly as they usually do when not coupled. The coupler sends some configuration information to each model (spatial mappings data, etc.).
2. A daily time step is used by MODFLOW which simulates the first time step and sends the water table depth (HNEW variable) to the coupler at coupling point A.
3. Using a daily time step, TOPMODEL receives the water table depth from the coupler and uses it to set the SRZ array (the initial root zone deficit) at coupling point B and also receives the baseflow from the coupler at coupling point C and uses it to set the *Qb* variable

4. After each time step at coupling point D, the drainage from the unsaturated zone is sent to the coupler from TOPMODEL. This value represents recharge to the saturated zone.

5. MODFLOW receives the updated water table heights that are used in its next time step computation. This sequence is illustrated in Figure 4.15

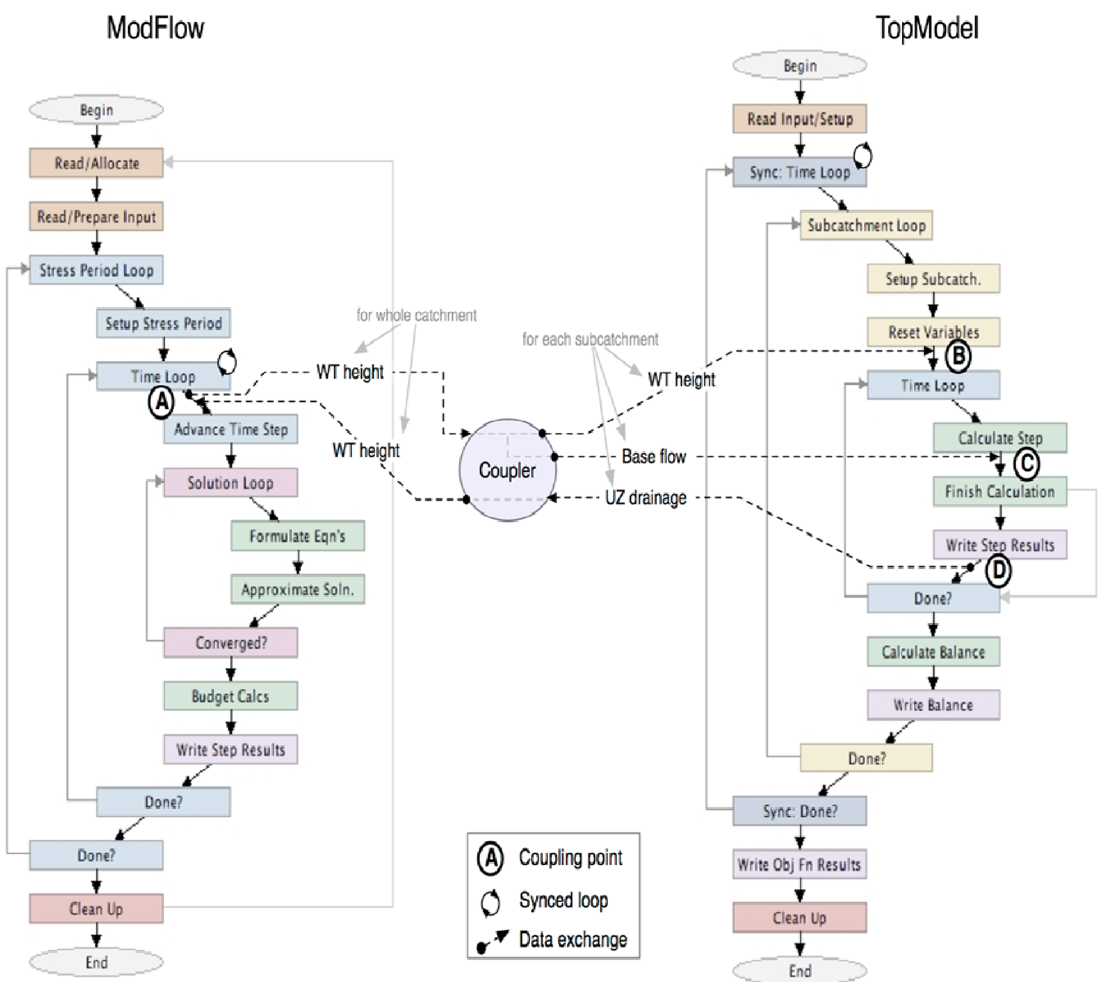


Figure 4.14. Model coupling specification.

The coupling description

Having created the specific model PCIs, the next step is to describe the coupling achieved. In this case, there is a unidirectional coupling, which means that only TOPMODEL is being influenced by MODFLOW. Figure 4.16 shows the coupling description, and it is clear form the diagram that the models will interact or communicate within their respective time step loops.

STEP 1

STEP 2

STEP 3

STEP 4

STEP 5

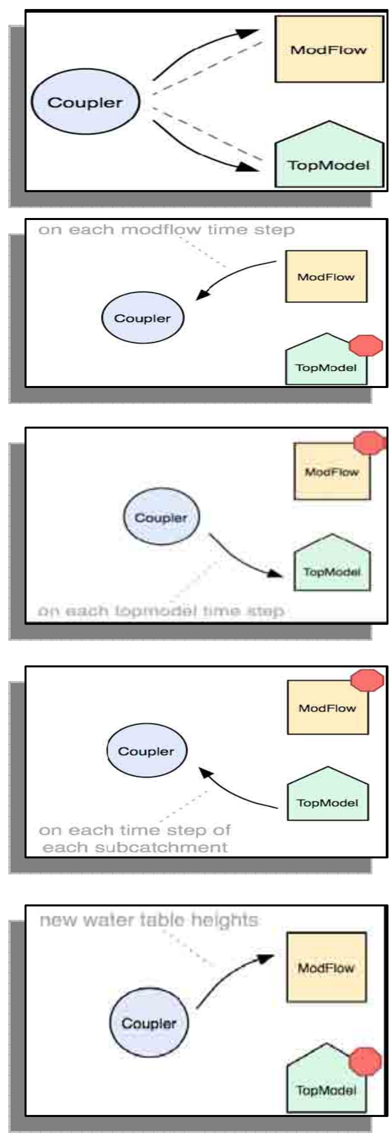


Figure 4.15. Event sequence in the coupled model (Bulatewicz, 2006).

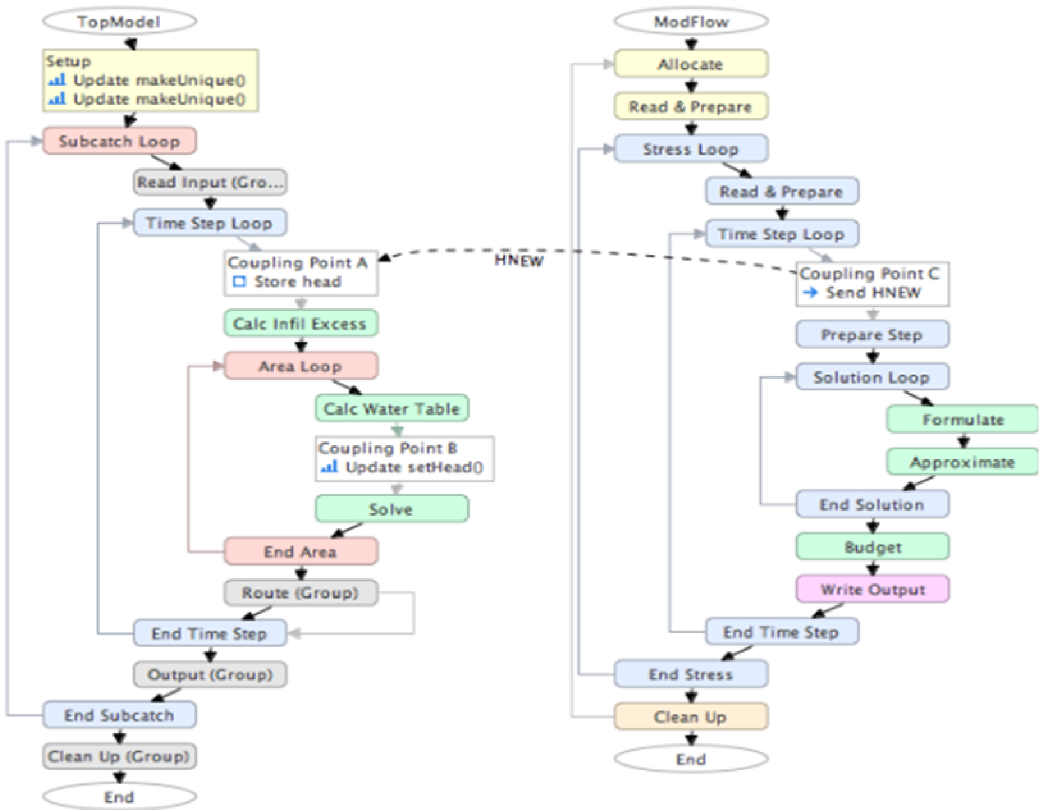


Figure 4.16. The TOPMODEL–MODFLOW unidirectional coupling description.

As highlighted in earlier sections, the two models, TOPMODEL and MODFLOW, generally use different temporal discretizations. TOPMODEL uses smaller time steps, usually hours; or to test long-term trends, days. MODFLOW uses longer time steps on the order of weeks or months, and in some cases years because of the relatively slow movement of groundwater compared to surface water movement. Thus it is important to address the synchronization of these differences in the temporal discretization in the model coupling. As long as the coupled model uses time steps that are similar for both models then it is not necessary to have any special function to synchronize the time steps. In situations where MODFLOW uses a longer time step, for example two days while TOPMODEL uses a one day time step, then there is a need to identify the greatest frequency at which the models will interact. In this case, it will be every two days. Thus TOPMODEL executes two time step loops for a single MODFLOW time step. In this study both models are set to run using daily time steps.

The coupling description includes three coupling points (A, B, and C).

Coupling Point A: While it is clear that the new head variable, *hnew*, obtained from MODFLOW is used at Coupling point C in the flow diagram, the value is sent to Coupling Point A and stored because Coupling Point C is located within a loop, and communicating with MODFLOW at that point would cause the models to become unsynchronized.

Coupling Point B: Since TOPMODEL needs to use MODFLOW's *hnew* variable, a Send Action is added to Coupling Point B in MODFLOW, which sends the variable's value to TOPMODEL, making it accessible at Coupling Point A.

Coupling Point C: To set the value of TOPMODEL's saturation deficit *(sd)* variable, an update action is added to Coupling Point A to apply the custom update function, *setHead*, which sets the value of the *sd* variable based on the value of MODFLOW's *hnew* variable. Note that MODFLOW's *hnew* value is an elevation, whereas TOPMODEL's *sd* variable is a depth. In order to set the *sd* value to the *hnew* value, the elevation must be converted into a depth. This requires knowledge of the elevation of the surface, since the depth is equal to the difference between the surface elevation and the water table elevation. A custom update function called *setHead* was written in Fortran to resolve this issue.

Model Calibration and Verification

Calibration of the model was carried out by comparing measured and calculated stream flow and groundwater heads for different observation points in the watershed. For effective assessment of model performance, reproducing past and present hydrological conditions and predicting future stresses are all goals within the modeling process. Thus, it is crucial to assess the degree of confidence that can be placed on model predictions. In this study, manual adjustment of parameters in both the ground water and surface water components of the coupled model was used as a calibration method to improve simulation results.

Results and Discussion

Stream flow

The main aim was to evaluate how effective the coupled model can be used as a tool to simulate the hydrological dynamics of the watershed. Therefore, the developed coupled model performance was compared to the performance of TOPMODEL in simulating stream flow and also compared to MODFLOW in simulating ground water flow. Model evaluation involved comparison of the simulated stream flow with the measured stream flow for Tenmile Creek near Laurel gaging station. The Ferndale gaging station was not used for calibration because of non availability of long-term stream flow records.

Early model comparison studies used graphical and statistical methods to evaluate results (World Meteorological Organization, 1975); however, evaluation of graphical criteria is subjective, as discussed by Houghton-Carr (1999), and no ‘best’ statistical quality criterion has been identified for hydrologic models (Weglarczyk, 1998). Therefore, two objective assessment criteria were used to compare TOPMODEL and TOPMODEL–MODFLOW model output in accordance with a method described by Perrin et al. (2001). The criteria were the Nash–Sutcliffe model efficiency, (E_f), (Nash and Sutcliffe, 1970) and the Mean Absolute Error (MAE).

E_f , which ranges from -∞ to 1, where 1 represents a perfect fit, is given as:

$$E_f = 1 - \frac{\sum_{i=1}^{n}\left(Q_{obs,i} - Q_{sim,i}\right)^2}{\sum_{i=1}^{n}\left(Q_{obs,i} - \overline{Q}_{obs,i}\right)^2} \qquad (4.2)$$

where $Q_{obs,i}$ = observed stream flow at time step i;

$\overline{Q_{obs,i}}$ = mean observed stream flow during the evaluation period, and

$Q_{sim,i}$ = model simulated stream flow at time step i

The second criterion, also transformed to a scale of -∞ to 1, was derived from the MAE which is given in the following relationship (Anderson and Woessner, 1992)

$$MAE = \frac{\sum_{i}^{n}\left|Q_{obs,i} - Q_{sim,i}\right|}{n} \qquad (4.3)$$

MAE is a weighted average of the absolute errors with the relative frequencies as the weight factors. Results for this analysis are shown in Table 4.4.

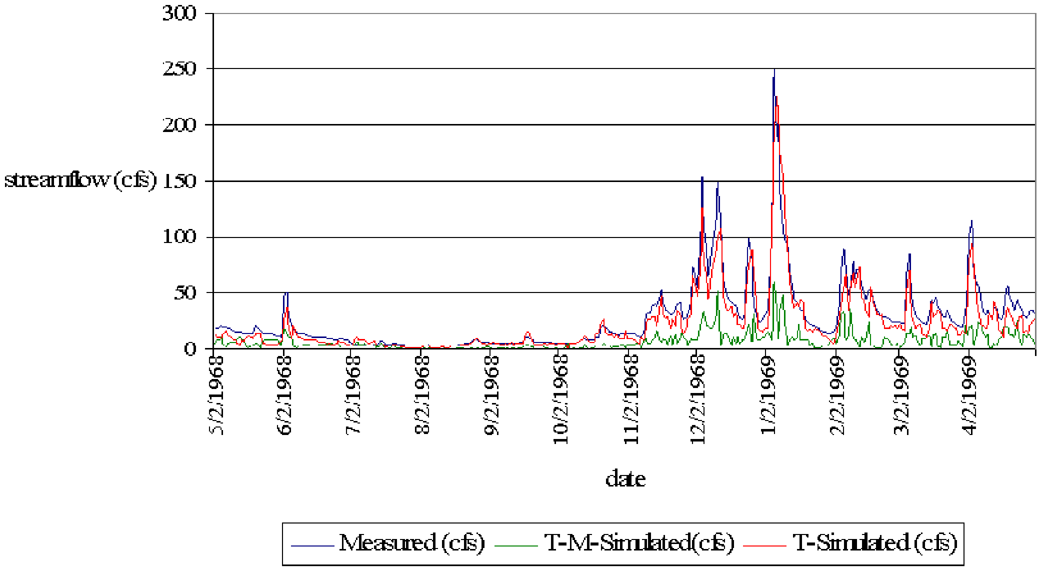
Table 4.4. Summary of TOPMODEL and TOPMODEL–MODFLOW model results for simulations of stream flow in the Tenmile watershed for the period 1968 to 1972

Period	Mean Stream flow [a] (cfs)	E_f TOPMODEL	E_f TOPMODEL–MODFLOW	MAE TOPMODEL	MAE TOPMODEL–MODFLOW
1968-69	28.1	0.88	0.64	0.050	0.32
1969-70	22.6	0.92	0.43	0.047	0.36
1970-71	29.8	0.86	0.51	0.104	0.52
1971-72	51.7	0.88	0.62	0.049	0.37
MEAN	33.05	0.885	0.55	0.063	0.39

[a] measured stream flow

The annual hydrograph generation was based on daily stream flow, which is the sum of surface runoff and subsurface flow to streams (baseflow). The stream flow integrates the hydrological responses from across the watershed and can, therefore, be used to assess the overall predictions of the model.

From the modeling results there is an apparent mismatch between observed stream flow and stream flow modeled using the coupled model (Figure 4.17a-d). Stream flow prediction during the calibration period, 1968-1970, generally agreed well with measured flow for TOPMODEL with an average Nash-Sutcliffe R^2 of 0.87. The Nash-Sutcliffe R^2 represents the fraction of the total squared error that is explained by the model or a statistic that will give some information about the goodness of fit of the model. Thus values approaching one are desirable.



“T-M-Simulated” means TOPMODEL–MODFLOW simulated stream flow
“T-Simulated” means TOPMODEL simulated Stream flow

Figure 4.17a. Measured and simulated stream flows for the years 1968-1969.

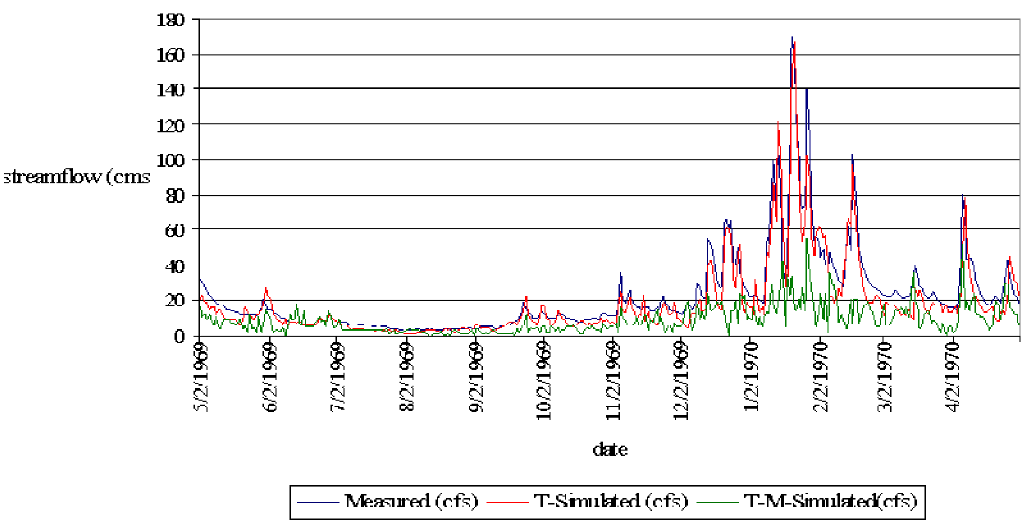


Figure 4.17b. Measured and simulated stream flows for the years1969-1970.

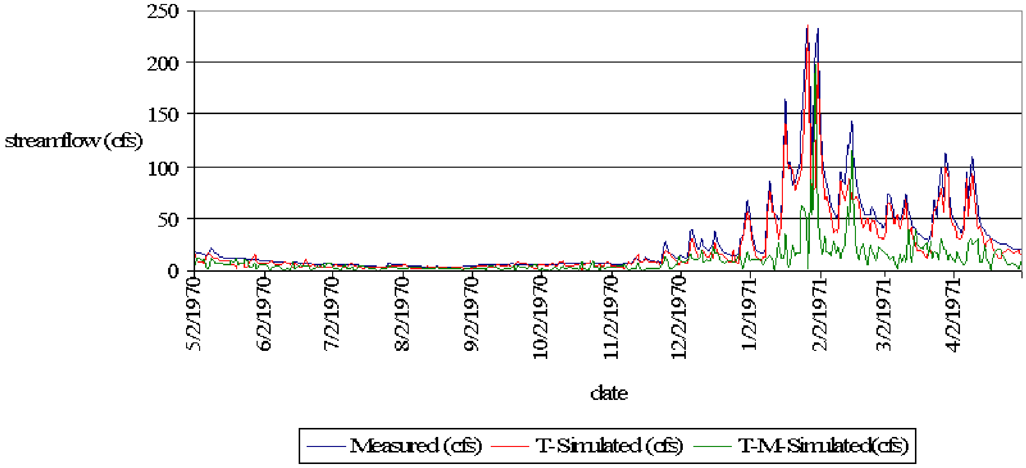


Figure 4.17c. Measured and simulated stream flows for the years1970-1971.

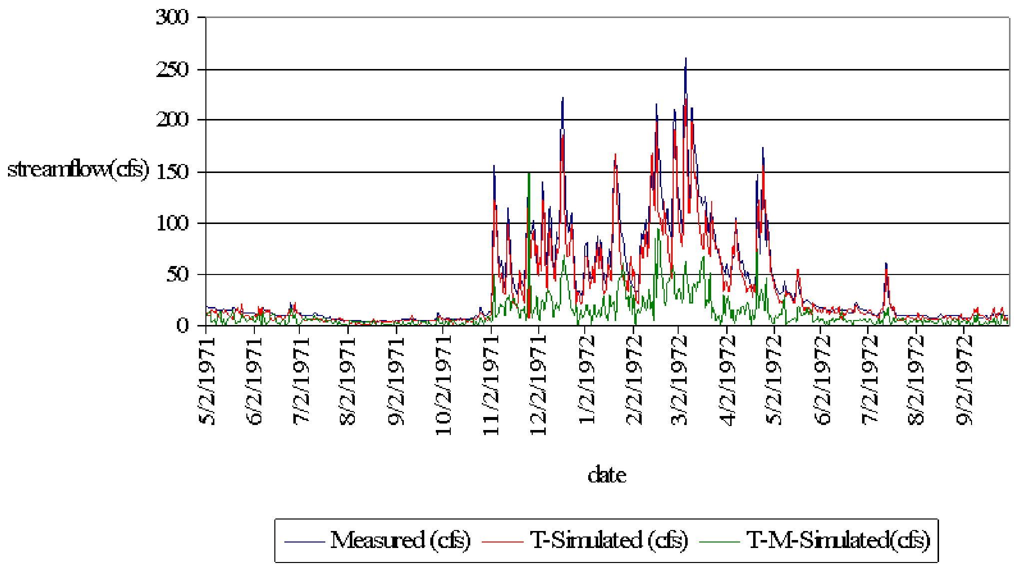


Figure 4.17d. Measured and simulated stream flows for the years1971-1972.

However, for the coupled TOPMODEL–MODFLOW model there was under simulation of stream flow for most of the period, and an average Nash-Sutcliffe R^2 of 0.54 was obtained for the same calibration period. For both TOPMODEL and the coupled

TOPMODEL–MODFLOW during most of the calibration period, the simulated low flow is quite close to measured flows, except for the 1969-70 period.

This mismatch is also evident in the scatter plot shown on Figure 4.18 and the corresponding poor total squared error that is explained by the model, R^2, obtained.

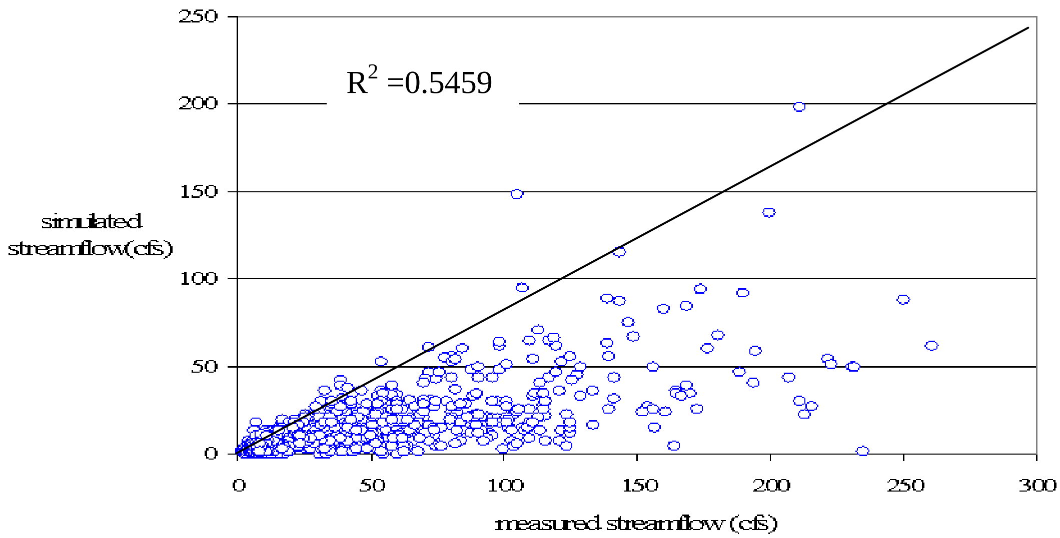


Figure 4.18. Correlation for measured and coupled TOPMODEL–MODFLOW simulated stream flow.

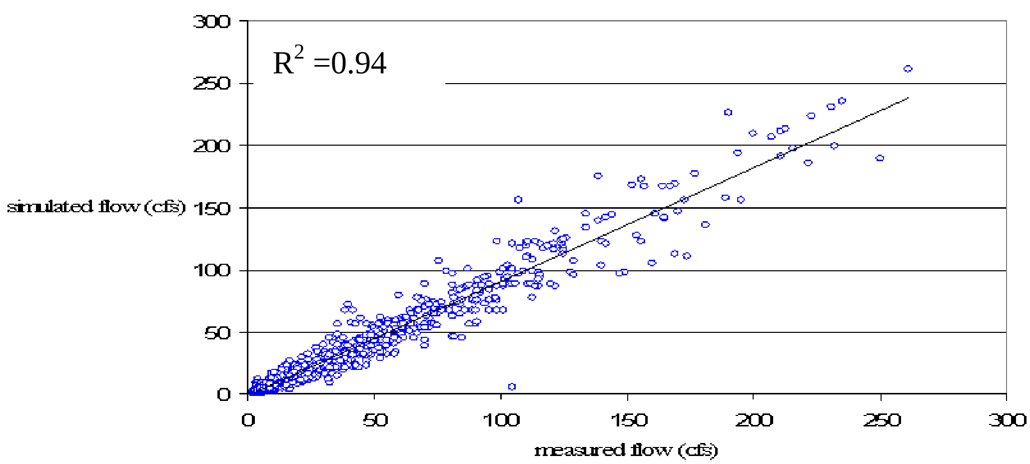


Figure 4.19. Correlation for measured and TOPMODEL simulated stream flow

Thus the model performed poorly in terms of point-wise comparisons. As an example, in the 1969-70 year, the low flow simulated by the model is generally underestimated and there are clear deviations. In general, the model properly describes the increase and recessions in the river flow, but the peak levels are more often underestimated for some rainfall events.

As shown in Table 4.4, TOPMODEL on its own produced good simulation results with average efficiency of 0.88 and a MAE of 0.0625 compared to the coupled model mean efficiency of 0.55 and MAE of 0.39. From these results, it appears that the poor overall simulation ability of the coupled model is mainly a result of problems in the groundwater model. Inadequacies in the groundwater model could be due to the various simplifying assumptions used.

The relatively large deviation between simulated and measured stream flows for the coupled model is most likely due to both model and measurement errors. Of particular interest, the uncertainty associated with extrapolating rainfall point measurements over the study area may represent a significant component of the error in the simulated flows. This affects both the coupled model and the stand alone TOPMODEL since the same rainfall data is used in both cases. However the error might be greater in the coupled model because any error in the data is propagated since the rainfall data is used to drive the recharge rates to ground water which impacts of wetness Indices and thus saturation status of an area and subsequently affects stream flow. As mentioned earlier, the rainfall used for this study was obtained from two climatic stations that are outside the watershed. Precipitation used in this study was obtained by averaging precipitation recorded at the two stations: Bellingham 2N and Bellingham FCWOS, AP.

TOPMODEL keeps track of the depth to water table state variable for each wetness index class, and this is utilized to determine the spatial distribution of the saturation status of an area and subsequent stream flow. However, in the coupled model, there is no storing of the same state variable at the wetness index class level as this complicates the modeling. Thus, there is a coarser distribution of water table depths using the coupled model compared to TOPMODEL alone. This can contribute to the lack of accuracy from the coupled model in describing stream flow. This can be improved by subdividing the catchment into small subcatchments. If subcatchments can be defined sufficiently small that they have relatively little variation in elevation, the relationship between the ground surface and the water table can be reasonably represented by a single value.

The depth to water table is critical in determining stream flow and spatial extent of saturated zones in a watershed. Generally shallow water table depths results in more runoff and stream flow compared to deeper water tables. The spatial variation of depths to the water table was plotted using TOPMODEL and also using the MODFLOW.

TOPMODEL results in more saturated areas compared to MODFLOW. These results show that the coupled model is not able to effectively describe stream flows in the watershed because of the influence of the ground water component of the coupled model. The error margin due to the groundwater component is higher. Obvious causes of this are the major simplifying assumptions made in the ground water model.

Differences between simulated and observed stream flow may have resulted from various causes, including model calibration error, discretization effects, or inadequate simulation of aquifer geometry, storage properties, recharge or other hydrologic processes.

<u>Watershed water balance</u>

The coupled model was used to understand the interactions between surface and groundwater resources by analyzing coupled model simulation results for the period 1968-1972. It was possible to prove that the coupled model was able to simulate baseflow fairly well compared to the uncoupled models as well as in comparison to baseflow separation results. This is shown in Figure 4.20 while Figure 4.21 shows the net recharge to groundwater, baseflow, and stream flow obtained using the coupled model in relation to rainfall.

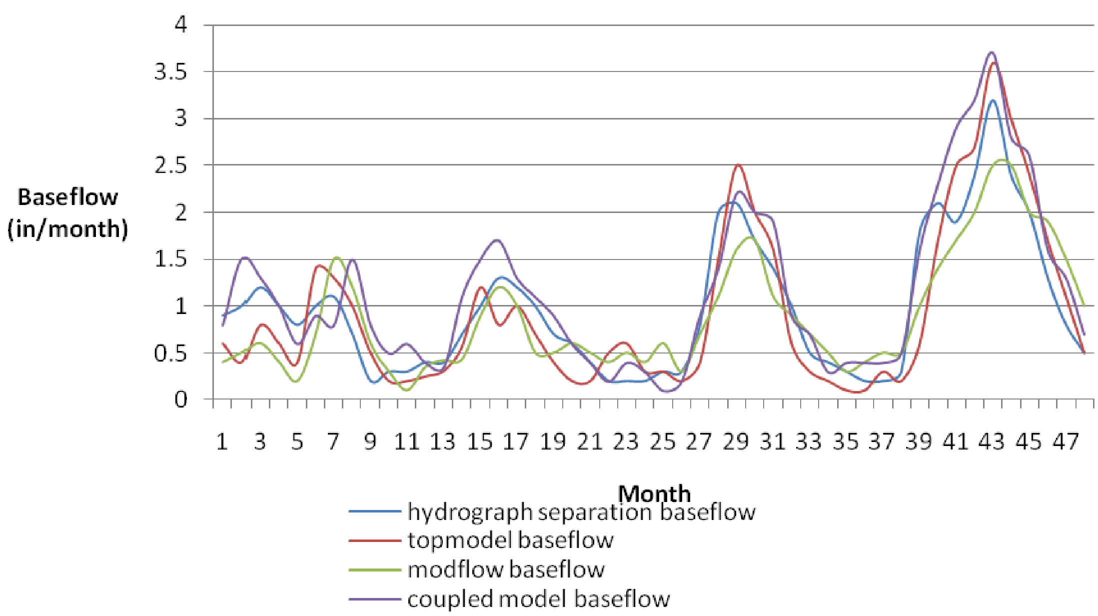


Figure 4.20. Mean monthly baseflow values for the period 1968-1972.

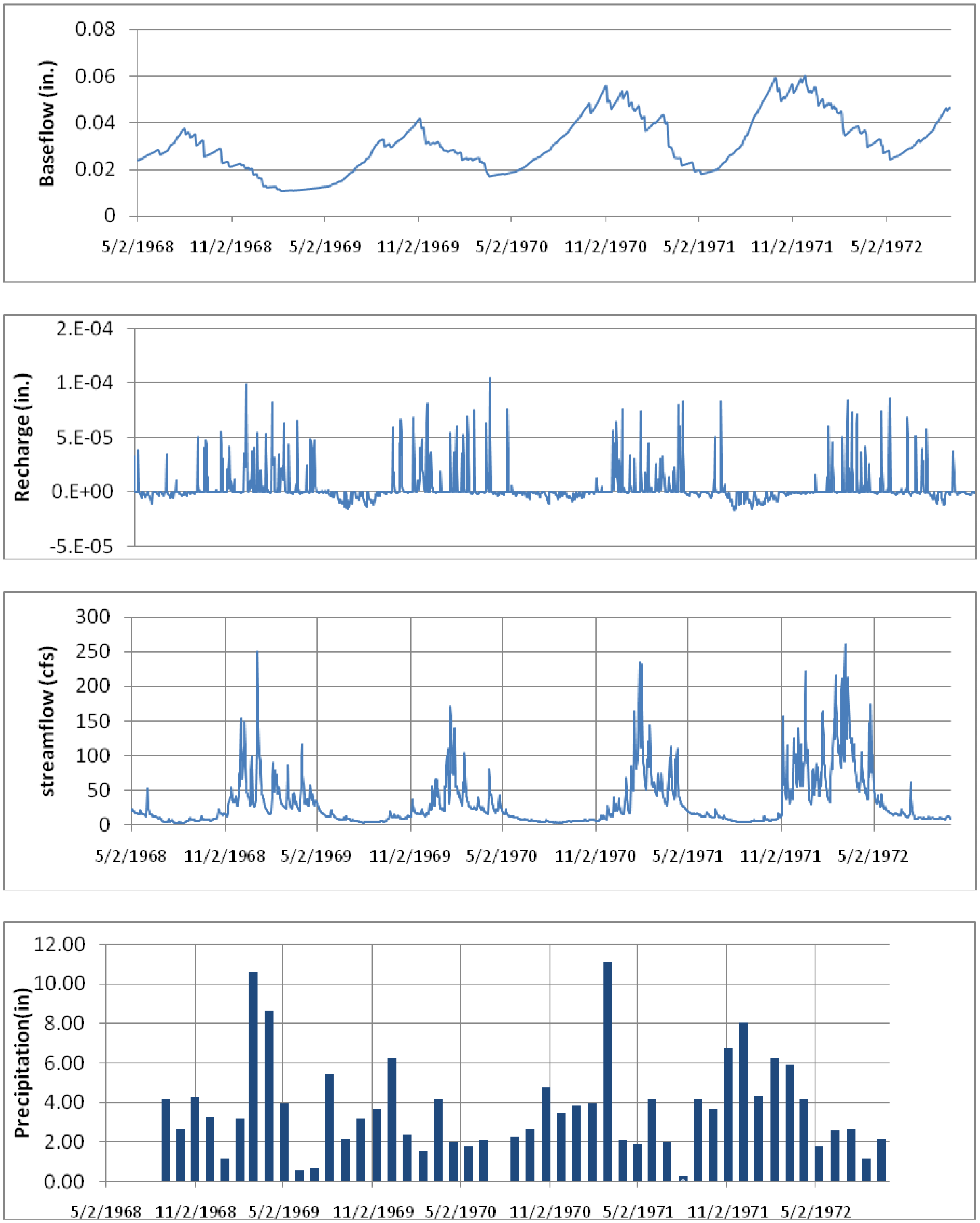


Figure 4.21. Dynamics of modeled groundwater recharge, baseflow, and stream flow in the Tenmile watershed.

Apparently, the net flow direction changes in time, occasionally very rapidly, caused by quick changes in surface water stages or intensive precipitation events. Although the interaction between surface and ground water bodies is important for the water balance of the watershed, the discharge of groundwater to the surface water, represented as negative recharge, is almost negligible compared with recharge to groundwater.

The discharge to the soil zone showed seasonal variations, as illustrated by the difference in the negative fluxes over time. Groundwater discharge to the soil zone can contribute interflow and surface runoff and thus, groundwater can contribute flow to streams through surface runoff and interflow. Seasonal variation in groundwater discharge to the soil zone is likely in this area due to the large seasonal variation in precipitation and stream flow. This phenomenon of upwelling of groundwater into the soil zone is common in lowland areas and wetlands. In these areas the soil zone moisture content is increased due to the presence of the shallow water table, influencing the water available for soil evapotranspiration and reducing infiltration capacity. The upwelling of water from the saturated zone to the soil zone represents a removal or loss of water from the groundwater saturated zone.

The increase in soil zone moisture content also causes a reduction in infiltration capacity and this will also lead to increased infiltration excess runoff. Another effect of upwelling of water into the soil zone is increased relative plant available water and also increased evapotranspiration withdrawals from the soil zone. Evapotranspiration withdrawals from the soil zone will result in negative net recharge to MODFLOW.

Although Figure 4.21 shows the temporal variation of exchange flow direction and intensity, it does not describe the spatial variability of the flows because in this study we only considered the sum of inflows and outflows.

The characteristics of the vertical dynamics of water movement in the watershed may be masked by the dynamical lateral interactions. This fact may also mean that the total groundwater balance in the watershed is strongly influenced by the lateral water flows rather than vertical flows, and the influences of lateral processes are much stronger than the vertical groundwater recharge. As a result of this, lateral groundwater flow and groundwater–surface water interactions have a major impact on the water balance of this watershed.

As Figure 4.21 illustrates, groundwater recharge generally occurs from winter to spring. From the relation between vertical groundwater recharge and lateral fluxes, it can be assumed that the groundwater recharge during this period is mainly caused by effective infiltration of surface water into the groundwater. This tendency is decreasing until the early summer, when the conditions begin to reverse and a period characterized by groundwater discharge starts. This is most likely due to higher transpiration losses during this time. During this period also stream levels decrease and there is a higher retention of the groundwater leading to water movement form the sub surface zone into the unsaturated zone which subsequently cause a negative water balance as shown by the negative recharge values in Figure 4.21. This process continues until the winter period when the water balance becomes positive again because of increasing surface water levels and subsequently more infiltration out of the river into the underlying aquifer and the influence of snow and snow melt.

A mean annual recharge of 18 inches per year was used in the development of the steady state ground water model. The coupled model utilizes recharge obtained from TOPMODEL. A mean recharge of 11 inches per year was obtained using TOPMODEL. The difference in the recharge values is most likely due to the influence of snowmelt which is not considered in TOPMODEL. Vaccaro, Hansen, and Jones (1998) considered the effects of snowmelt on recharge to ground water in their development of regression equations for precipitation and recharge. The mean recharge value from TOPMODEL is significantly lower than the mean value used in the steady state ground water model and this therefore is a possible reason why the coupled model under simulates ground water heads compared to MODFLOW on its own. However the TOPMODEL recharge brings in an advantage of spatial variation compared to a single value used in MODFLOW.

The mean ground water budgets for the watershed using MODFLOW and also the coupled model are shown in Tables 4.5 and 468. The tables show that the coupled model results in a suppressed water budget. There is a general under prediction. This is also manifested in a comparison of the recharge values obtained from the coupled model and the mean recharge values from Vaccaro, Hansen, and Jones (1998). Vaccaro, Hansen, and Jones (1998) also obtained recharge values ranging from 12-18in /year for the watershed while the coupled model estimated recharge values average is 8 in/yr. According to Vaccaro, Hansen, and Jones (1998) the 12-18 in/year recharge from precipitation represents about 43-58% of annual precipitation and this is typical. The coupled model recharge rate is thus a suppressed or under prediction. This difference in recharge values is the likely cause of the larger error in the water balance obtained using the couple model.

Table 4.5. MODFLOW model calculated steady state hydrologic budget for the Tenmile watershed

Hydrologic Budget component	Rate of Flow Cubic feet/ sec
Inflow	
Recharge from Precipitation	52
Total Inflow	52
Outflow	
Stream flow ground water discharge to streams	43
Evapotranspiration	2
Water withdrawal	3
Total Outflow	49
Budget error	3

Table 4.6. Coupled TOPMODEL-MODFLOW model calculated steady state hydrologic budget for the Tenmile watershed

Hydrologic Budget component	Rate of Flow Cubic feet/ sec
Inflow	
Recharge from Precipitation	49
Total Inflow	
Outflow	
Stream flow and ground water discharge to streams	37
Evapotranspiration	2
Water withdrawal	3
Total Outflow	42
Budget error	9

<u>Groundwater head</u>

A comparison was also made of the effectiveness of the coupled model in simulating ground water flow and heads against that obtained using MODFLOW alone. This is important in order to understand whether the model coupling is a worthwhile exercise and understand whether model coupling results in any improvement in groundwater flow simulation.

An analysis of modeling results using MODFLOW alone is given in the next section as is an analysis using the coupled model. Figure 4.22 (a) – 4.22 (e) show time series of measured and MODFLOW simulated groundwater heads.

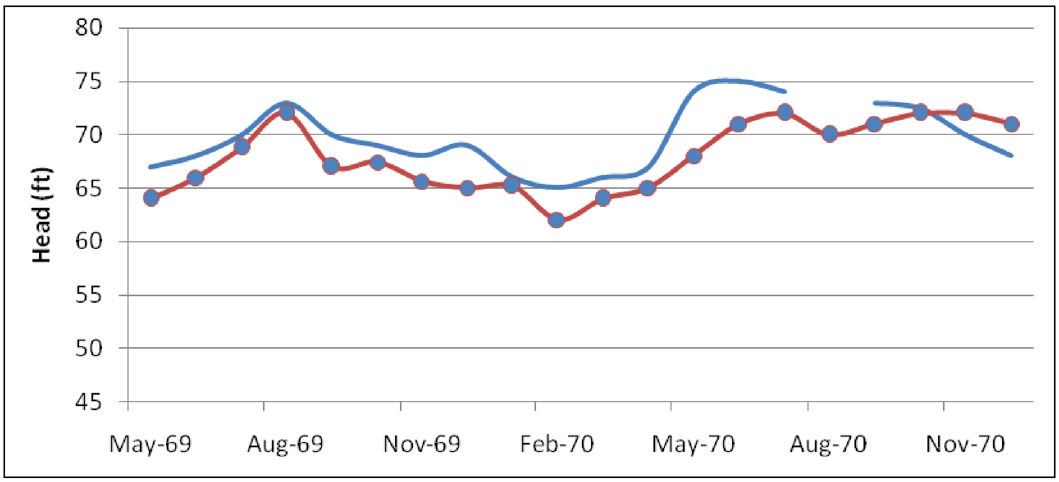


Figure 4.22. (a) Measured and simulated groundwater heads for calibration well 15.

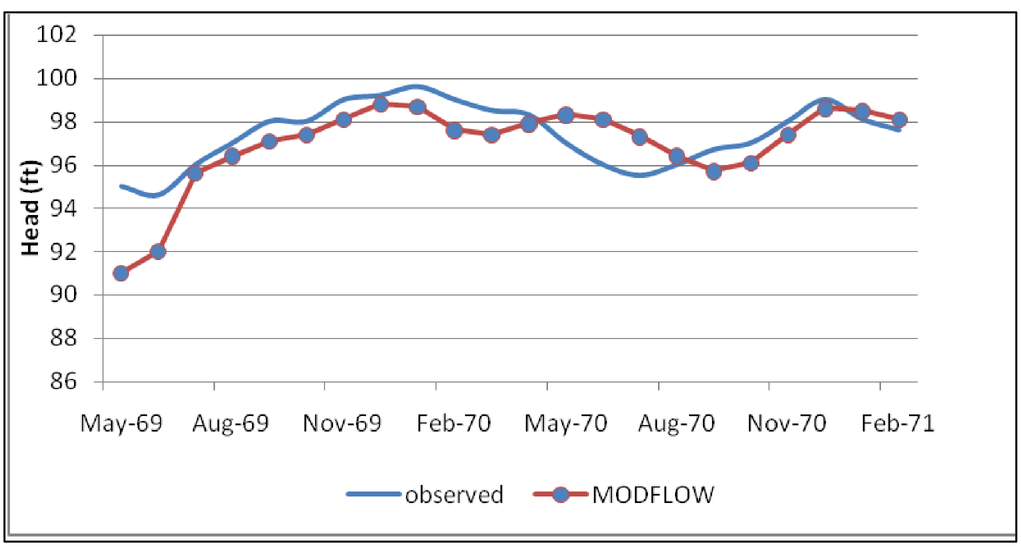


Figure 4.22. (b) Measured and simulated groundwater heads for well 17.

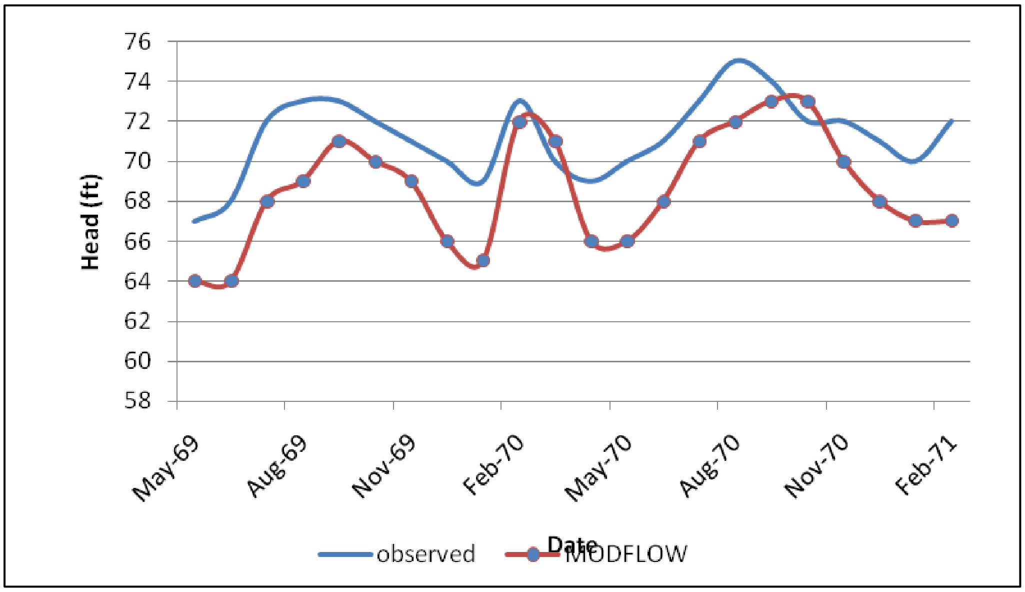


Figure 4.22. (c) Measured and simulated groundwater heads for well 18.

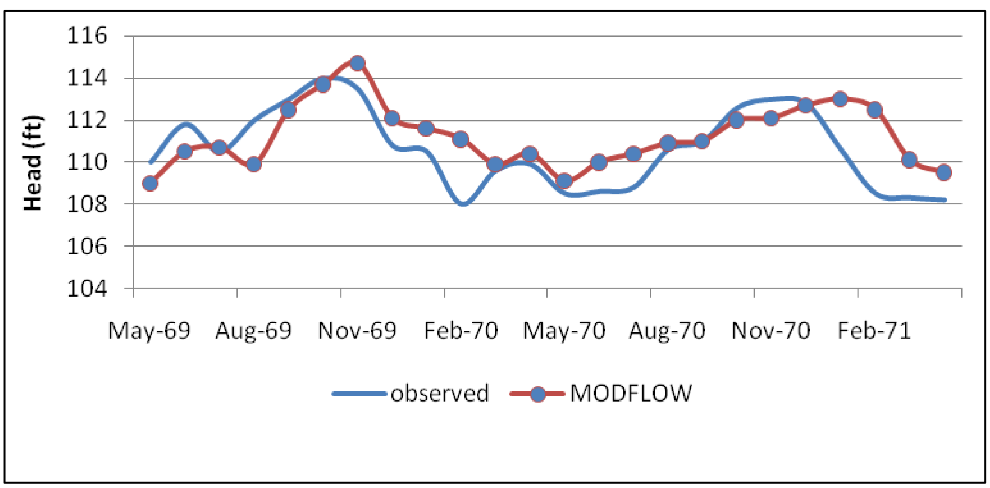


Figure 4.22. (d) Measured and simulated groundwater heads for well 20.

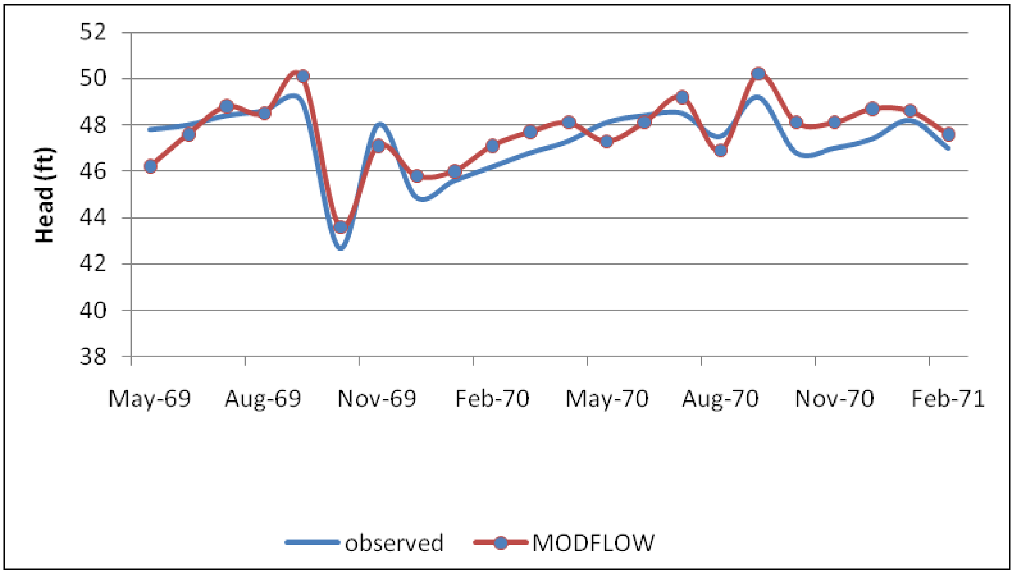


Figure 4.22 (e) Measured and simulated groundwater heads for well 21.

The hydraulic head correlation for the Tenmile calibration wells are shown in Appendix D. Generally the ground water model was able to simulate the temporal trends in ground water heads. The massive blip shown on Figure 4.22 (a) is mostly due to a measurement error in the observed groundwater head in August 1970. There is also a general trend of under simulation of groundwater heads which is clearly seen on well 18 (Figure 4.22 (c)). Correlations between simulated heads and measured heads would give a clearer picture on the performance of the model. These correlations are shown in Appendix E for each of the calibration wells.

The steady-state and transient flow models of the Tenmile Watershed provide a watershed-scale simulation of ground-water flow in the aquifers in the study area. As with all mathematical models of natural systems, the simplifications and assumptions incorporated into the models result in limitations to their appropriate uses and to the interpretations that may be made of simulation results. Hydrologic processes and spatial variability in hydraulic properties and stresses are simplified and approximated to a degree consistent with this scale. The model calibration also represents the best fit to estimates and observations made throughout the watershed. Thus, the agreement between simulated water levels in specific areas of the flow system may not be adequate to support local-scale model applications.

Studies by Cox and Kahle (1999) observed that there exists more than one layer in parts of WRIA1. However, the scarcity of hydraulic parameters for the deeper aquifer, piezometer readings in those aquifers that are used to constrain the uncertainty associated with the parameters as well as aquifer layering information make it difficult to build, calibrate and verify a multi layer ground water quantity model. There are no piezometer readings at places where the existence of two layers is recognized. Also there are no vertical hydraulic conductivity estimates that would enable a good multi-layer model to be developed. In some areas well logs penetrated only the surficial aquifer while in other areas (like public water supply drillings) there exist deeper aquifer depth information. This variable aquifer depth data could not be interpolated/extrapolated without several uncertainties. Thus a single layer model was adopted and the results are not very representative as shown by the poor correlation between simulated and measured heads.

The simulated groundwater flow and water levels may not match the measured values because the hydraulics of the surface water system, such as storage provided by impoundments and wetlands, were not simulated. The effects of these controls on surface water flow were not included in the simulated stream base flows. The approach of representing stream stage by a fixed value representing average conditions may lead to some inaccuracies in flow rates between aquifers and streams, particularly during periods of high flow.

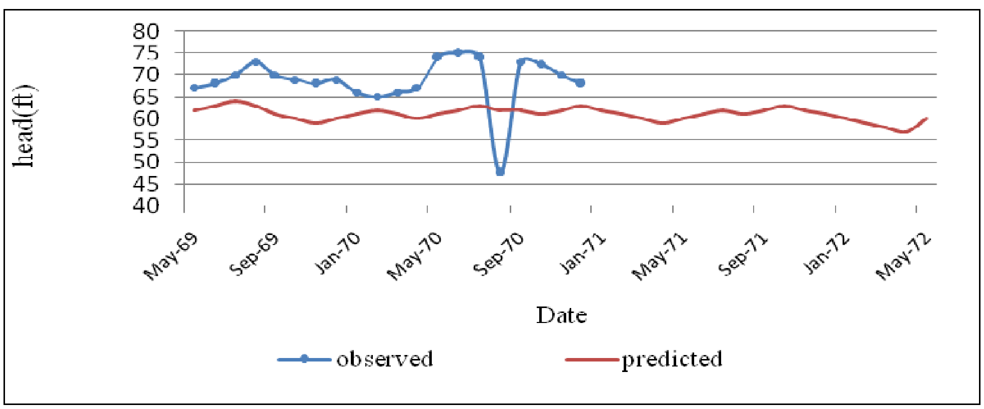
Figures 4.23(a-e) show time series of predicted and observed groundwater heads for the Tenmile watershed using the coupled model. A range of well-matched and poorly matched hydrographs is shown to provide an overall indication of the performance of the model. The predicted heads to the end of 1972 are shown to give an indication of predicted long-term trends and seasonal fluctuations in predicted head. Figure 4.23(a) shows subdued, regular seasonal groundwater fluctuations at well 15. This piezometer is close to the catchment boundary, but observed groundwater heads vary between three to six feet below ground level. The model consistently under predicts the heads, and the predicted trend is initially downward; however, predicted heads are within 7.5 feet of the ground level and the trend is slightly upward by the end of the simulation. Figure 4.23(b) shows that the heads at well 17 are over predicted. However, the amplitude of the seasonal variation in groundwater level is well matched, apart from the large fluctuations between May and December 1969. The main source of error at this well could be that there is no proper or correct representation of intermittent pumping to simulate the actual abstraction that occurs. Therefore, groundwater levels that will vary significantly, depending on pumping, are represented by average conditions simulated by average, seasonal abstraction rates.

Figure 4.23(c) shows that heads at well 18 are over predicted and that the seasonal variations in groundwater are under estimated. The long-term trend in observed groundwater heads, however, is well matched by the model. Figure 4.23(d) shows that the model consistently under predicts groundwater heads at this location. Geological assessment of this area shows the presence of several dykes. This can be a reason why it is particularly difficult to match the measured groundwater heads and simulated heads because the heads are controlled by factors operating at a smaller spatial scale than the model. Figure 4.23(e) shows the predicted and observed heads at well 21. Observed data shows a slight downward trend which is moderately approximated by the model in the initial times steps. From September 1970 to January 1971 the coupled model does not capture the increasing trend in ground water heads shown by the observed data. However between January 1971 and May 1971 the coupled model is able to describe the decreasing trend in ground water heads.

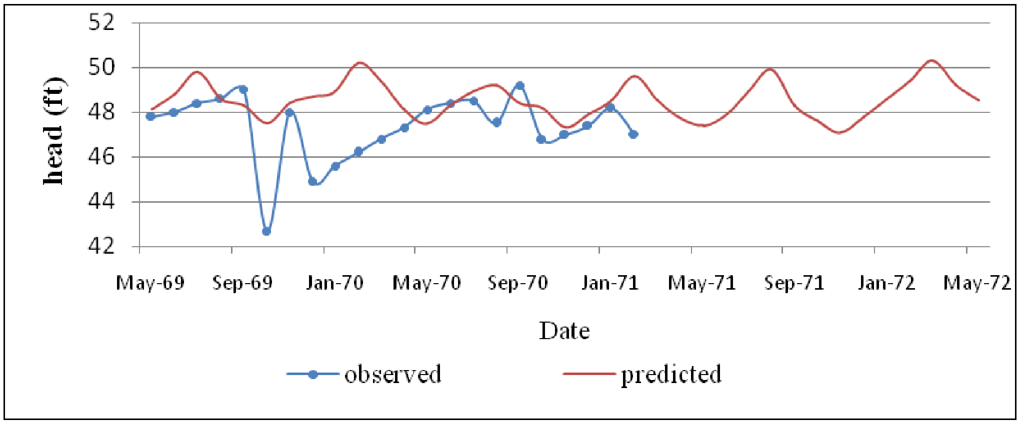
In comparison, an average correlation coefficient of 0.604 was obtained between MODFLOW simulated and measured heads while an average of 0.52 was obtained between the coupled model and measured heads. This shows that MODFLOW alone was able to describe the ground water dynamics in the watershed better than the coupled TOPMODEL-MODFLOW model.

Groundwater levels fluctuate in the watershed in response to a variety of driving forces working at different temporal and spatial scales. Long-term climate cycles are the dominant driving force and annual recharge cycles also work in the watershed. Canal leakage and irrigation, also important driving forces, affect only parts of the model area and have an annual cycle that is different from the timing of natural recharge. Other forces include stream-stage variations and ground-water pumping. These work at a variety of scales, generally small relative to other stresses. The model simulates the fluctuations caused by climate cycles, natural recharge from precipitation. Canal leakage and recharge from irrigation is not captured in the model. Stream-stage variation is also not explicitly represented in the model, so water-level fluctuations resulting from variations in stream stage are not simulated. Although groundwater pumping is included in the model, drawdown effects are averaged over entire model cells, so large fluctuations close to pumping wells are not simulated.

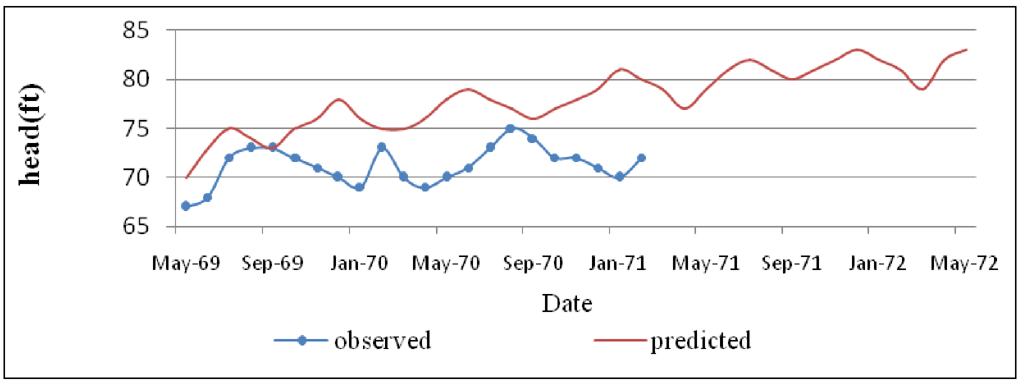
The model is probably not reflecting the moderating effect of leakage from the rivers and creeks on long-term fluctuations. In addition, simulated water levels do not show the seasonal fluctuations in the measured water levels. This is expected because the stream stage variations that drive water table fluctuations in this area are not present in the model. Thus simulated water levels do not show the water level rise caused by the onset of wet climatic conditions seen in the measured data.



a) Well 15

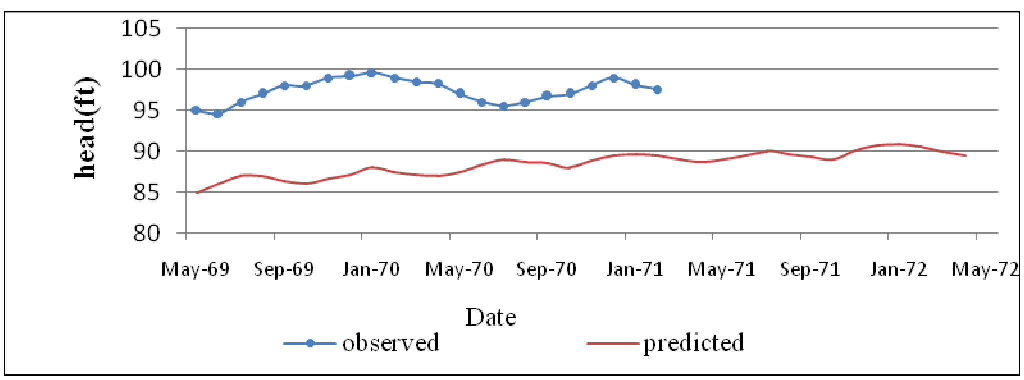


(b) Well 17

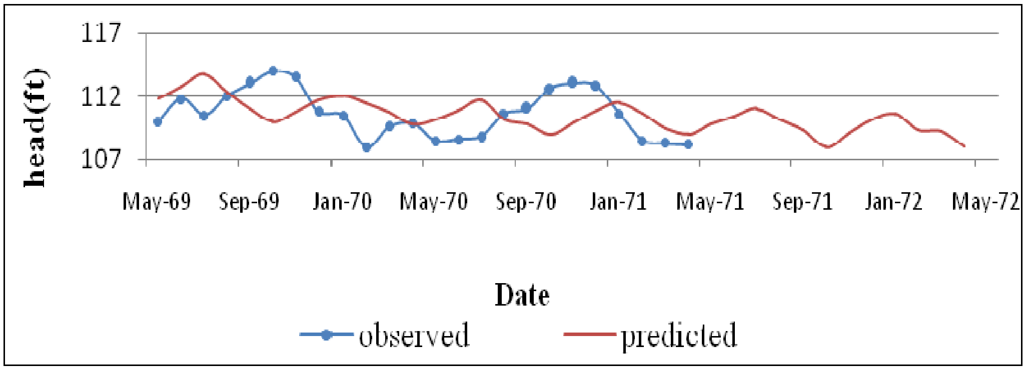


(c) Well 18

Figure 4.23 (a-e). Time series comparison of modeled and measured groundwater head at the observation wells using the coupled TOPMODEL-MODFLOW model.



(d) Well 20



(e) Well 21

Conclusions

Fundamentally, the study was designed to be a "first cut" for demonstrating the potential usefulness of the model coupling tool for simulating surface water–groundwater interactions in a watershed. Based on the published literature, a conceptual model of the hydrological conditions of the Tenmile watershed study area was chosen to represent the complex process structure of the watershed. For a successful simulation of the water balance and groundwater dynamics of the study area, it was possible to show that an adequate model has to deal with the temporal and spatial dynamics of runoff generation processes, the interactions between groundwater and surface waters, and their variable impacts on the catchment water balance. Based on the conceptual model, a coupled modeling system has been developed that simulates rainfall runoff processes, groundwater recharge, and exchange fluxes between the saturated and unsaturated zone. It was demonstrated that the coupling of TOPMODEL and MODFLOW is a potentially useful approach to characterizing the water balance of a typical watershed and to analyzing surface water–groundwater interactions, including the quantification of transfer fluxes even though the coupled model results were not of the expected levels. More refinement is required in the ground water component of the coupled model to improve on the simulations.

Model coupling was achieved using the Potential Coupling Interface technique and was evaluated for its ability to simulate stream flow and groundwater stages over a four-year period in the Tenmile watershed. TOPMODEL gave adequate simulations of stream flow while the coupled model underestimated stream flow. Statistical evaluation criteria that were applied showed both TOPMODEL and the coupled TOPMODEL–MODFLOW models to produce acceptable results for the entire period. However, a detailed water year evaluation of the four year study period showed differences in accuracy. The comparative analysis indicated that TOPMODEL on its own was more accurate than the coupled Model in the simulation of stream flow.

The importance of lateral exchange processes for the watershed groundwater dynamics and water balance could be quantified using the coupled model. The results of annual water balance simulations also shows the importance of groundwater–surface water interactions on the change of groundwater storage in the watershed. Use of the couple model also revealed groundwater upwelling into the unsaturated zone and surface water infiltration and how these two processes dynamically affect the water balance simultaneously. It is most likely that the overall groundwater dynamics in the watershed are also mainly controlled by lateral interactions between groundwater and surface waters.

Even though the temporal variations of hydrological processes are modeled fairly well compared to measured results, the coupled model does not suffice to simulate the complex surface-groundwater system of the Tenmile watershed at the moment. However, the approach of using spatially and temporally variable recharge derived from a surface water model, TOPMODEL, is promising with better knowledge of aquifer properties. In order to adequately account for the full effects of surface water–groundwater interactions, a better understanding of the groundwater aquifer is required so that some of the applied simplifying assumptions used in MODFLOW can be modified. The model also needs to be tested over a longer time period. More measured data is required to ensure that a more accurate groundwater model can be developed, calibrated and used. Since much of this information has yet to be quantitatively determined for this watershed, the described model may be regarded as a best first estimate given the data available. Further research is therefore necessary with more emphasis on the accuracy of the sub surface simulation model accuracy and elimination of too many assumptions.

CHAPTER 5

INTEGRATION OF TOPNET AND MODFLOW MODELS WITH APPLICATION TO THE BIG DARBY WATERSHED, OHIO

Introduction

The primary objective of this case study was to use the Potential Coupling Interface tool for the development of a coupled model with application to the Big Darby watershed, Ohio. The water movement through the hydrological cycle, which starts with precipitation and, after going through the processes of evapotranspiration, direct flow, infiltration, and groundwater recharge, ends its "journey" as baseflow in the main stream, will be simulated using two models, TOPNET and MODFLOW in an integrated way.

In this section, there will be an analysis of the role of the groundwater aquifer in the transport of precipitation through the groundwater aquifer towards the connected stream. Besides the development and testing of the coupled model, an initial question of interest was to establish any physical conditions in the watershed warranting the development of a coupled model. A review of available literature was also necessary to establish whether the physical and hydrologic conditions in the watershed reflect evidence of ground water surface water interactions. The Big Darby watershed was selected because of the availability of hydrological data, both surface and sub-surface data, needed for model parameterization and calibration.

Study Area Description

Location

The Big Darby Creek watershed is located 40 kilometers west of downtown Columbus, Ohio, and covers 1440 square kilometers (Figure 5.1). The watershed encompasses portions of seven counties: Logan, Clark, Union, Champaign, Madison, Franklin, and Pickaway. From its headwaters in Logan County to its confluence with the Scioto River near Circleville, the Little and Big Darby creeks traverse rolling hills in the headwaters and large flat plains in the midsection, and drop into large floodplains near the mouth. There are large expanses of relatively flat, poorly drained soils, well-suited for agriculture with proper drainage throughout most of the watershed.

The main stem, Big Darby Creek, originates in Logan County and flows southeat for 132 kilometers to its confluence with the Scioto River, north of Circleville. The major tributaries are Flat Branch, Spain Creek, Buck Run, Treacle Creek, Sugar Run, Little Darby Creek, Hellbranch Run, Spring Fork, and Robinson Run.

Climate

The Darby Creek watershed lies in the temperate climate of Central Ohio. The Midwestern Regional Climate Center (MRCC) collects historical climate data from various observer stations throughout the Midwest. In the Darby Creek watershed, these stations are located in Irwin, Marysville, and Circleville. Data from the Irwin, Ohio station is represented in Table 5.1 as a representation of the climate in the Darby watershed. This data is also shown in Appendix C. for the Columbus weather station

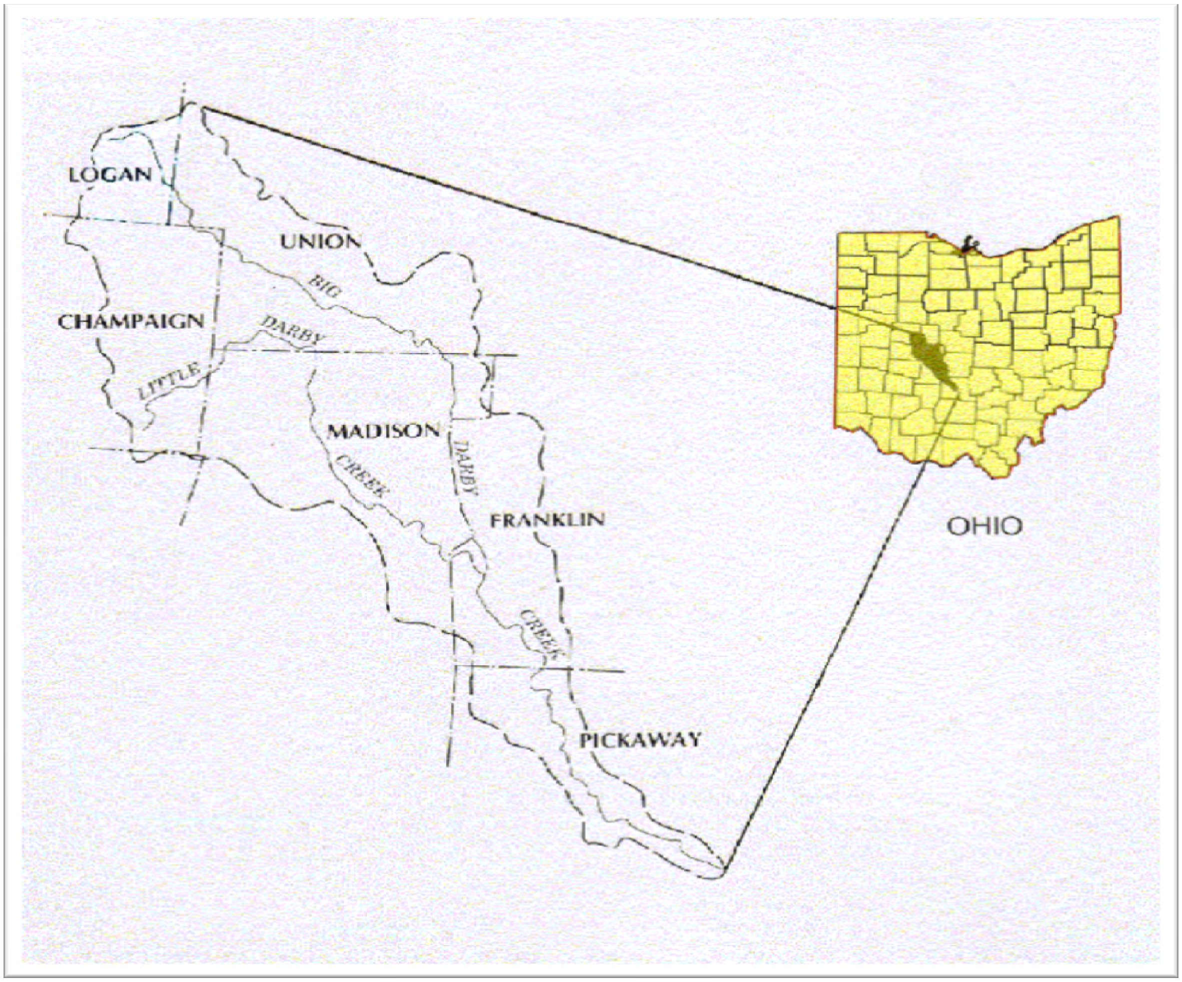


Figure 5.1. Location of the Big Darby watershed, Ohio.

Table 5.1. Irwin, Ohio station. Weather data collected 1991-1997

Temperature		
Temperature range	27.0 F (January)	74.1 F(July)
Mean Temperature	52.2 F	
Precipitation		
Precipitation range	49.5 mm (February)	118 mm (July)
Average annual rainfall	969 mm	

Soils

Soils in Ohio have been analyzed on varying levels of detail since the late 1800s. The soils data is from both the Ohio Department of Natural Resources (ODNR) and United States Department of Agriculture (USDA).

The Darby Creek watershed falls within the Eastern Corn Belt Ecoregion of Ohio. The glacially created soils in the Darby watershed can be generalized into ten soil associations (Table 5.2). The name of the soil association defines the predominant soil types within the watershed. Table 5.2 defines the soil associations and their frequency within the watershed.

Table 5.2. Big Darby watershed soil associations

Soil Association Name	Percentage Area
Kokomo Crosby Miamian	28.00
Miamian Celina Crosby	20.30
Brookston Crosby Celina	16.30
Crosby Miamian Brookston	11.90
Blount Glynwood Morley	10.20
Nappanee St. Clair Paulding	6.00
Eldean Ockley Sleeth	5.00
Blount Pewamo Glynwood	1.30
Eldean Westland Patton	0.77
Miamian Eldean Crosby	0.18

In the STATSGO soil associations there are five dominant soils making up 70 percent of the watershed: Crosby, Brookston, Miamian, Blount, and Kokomo. The most extensive soil unit is the Kokomo silty clay loam and this is a poorly drained soil with nearly level topography. It has moderately slow permeability and has very slow to ponded runoff. Kokomo soils were the primary “prairie soil” that supported most wet prairies in the Darby Plains during early settlement (Gordon and Simpson, 1994). The

Crosby is a poorly drained silt loam with nearly level to gently sloping topography. It has slow permeability and slow to medium runoff. Crosby soils, the second most extensive soil type in the Darby Plains, supported the mixed oak forests at the time of early settlement (Gordon and Simpson, 1994).

Brookston is a very poorly drained silty clay loam with level to nearly level topography and slow permeability. Miamian soils are a well drained silt loam on gently sloping to very steep topography. They have moderately slow permeability and medium to very rapid runoff. Blount is a somewhat poorly drained silt loam on nearly level to gently sloping topography with slow to moderate permeability and runoff. Fieldwork is delayed in spring due to wetness for many of the soil types. To counteract this, much of the landscape has been drained by ditches and tiles. Other soils found in the watershed include Celina, Eldean, Glynwood, Morley, Nappanee, Ockley, Patton, Paulding, Pewamo, Sleeth, St. Clair, and Westland.

Geological framework

All watershed drainage patterns, topography, soils and water chemistry are influenced by their underlying geology. The geology of the Big Darby Creek watershed is, in large part, the result of interactions between four successive glacial periods. Figure 5.2 shows an idealized geological cross section of the watershed. Throughout the watershed, substrates are derived from the calcareous sedimentary parent materials seen in the region’s bedrock. Igneous substrate constituents and glacial remnants also appear in the Darby system. They were carried to Ohio by the continental ice sheets. A series of end moraines in the Big Darby Creek watershed resulted from the advances and retreats of the glaciers which influence the stream system itself and the watershed landscape in multiple ways. Among these are watershed topography and spring water contribution to tributaries and the mainstem.

The predominant bedrock unit is the Silurian formation which provides water for farm, domestic, and industrial purposes in the watershed. The transmissivity of the carbonate aquifers ranges from 190 to 400m^2/day while the storativity varies from $1.0*10^{-3}$ to $1.0 * 10^{-5}$. Ground morain or silty clay till covers 85% of the watershed and ranges in depths from one to 30m in some places.

Sand and gravel bodies are common in the till and are sources of water for approximately half of the farms and domestic wells. Recent alluvial deposits are thin and not an important source of groundwater. The hydraulic conductivity of the various sand and gravel aquifers ranges from 40 to 120m/day. The variation of hydraulic conductivity has a big influence on the ground water surface interactions as this affects the rate of water movement within the subsoil strata.

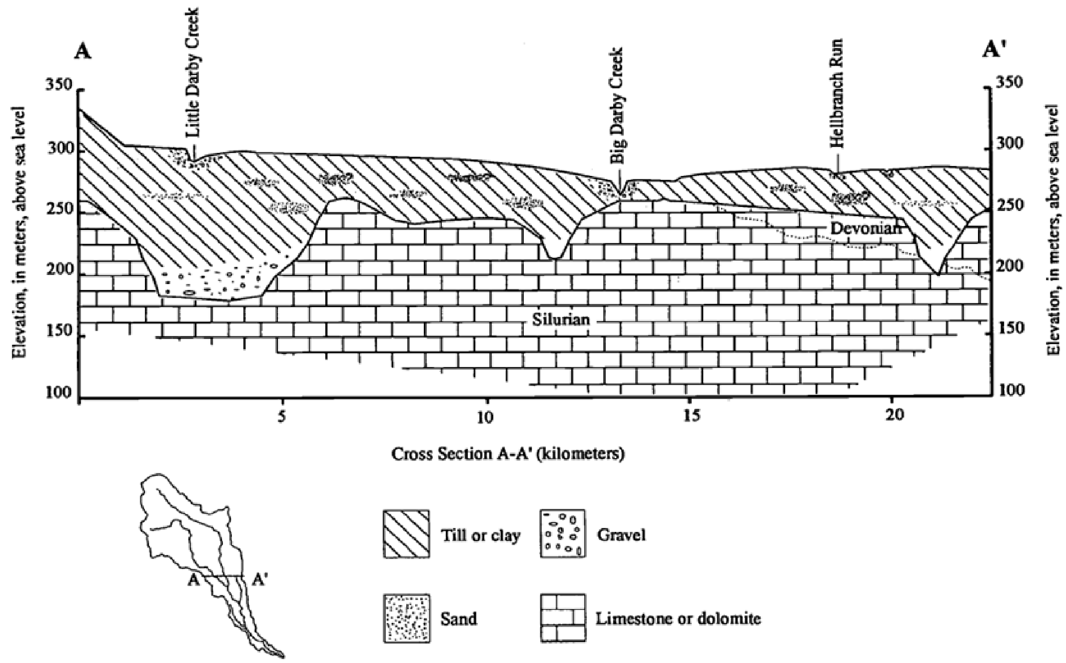


Figure 5.2. Generalized cross section of Geology in the Big Darby watershed, Ohio (Yu and Schwartz, 1998).

Groundwater

Groundwater is an important source of drinking water for many of the rural residents in Ohio including the Big Darby watershed inhabitants. According to Darby Joint Board of Soil and Water Conservation report (2002), more than 90% of the population uses groundwater for domestic, agricultural and industrial user. Appendix B shows the locations of ground water wells within the counties in which the Big Darby watershed lies in. Larger amounts of groundwater are typically found in porous geologic formations such as sand, gravel, or certain types of bedrock. Geologic formations that produce usable amounts of groundwater are referred to as aquifers. The most productive aquifers in Ohio are typically buried valley aquifers that consist of thin layers of glacially deposited sand and gravel surrounded by bedrock. Buried valleys in the Darby watershed are likely to be remnants of Wisconsinan glacial activities and possibly tributaries to Teays River which cuts through parts of Madison, Franklin and Pickaway Counties within the Darby watershed. The buried valley follows the present day course of the southern part of the Little Darby Creek and then follows the Big Darby Creek south to Circleville. Another area that contributes a large amount of groundwater is an area in Champaign County, in the headwaters of the Little Darby Creek. This area is another buried valley of a tributary to the Mad River. The yields in the remainder of the watershed are largely dependent on the thickness of the glacial deposits and the presence of the highly porous areas of sand and gravel.

Though sand and gravel deposits are excellent sources of groundwater, many wells in the Darby watershed extend to layers of porous Devonian and Silurian age limestone. The depth to the limestone or bedrock varies across the watershed. Domestic wells in the Darby watershed typically yield between 5 to 7 cubic meters per hour depending on the type of limestone, number of fractures, and proximity to streams.

The 2002 Darby watershed inventory report by the Darby Joint Board of Soil and Water Conservation also notes that ground water is the essential source of water for streams during prolonged dry spells. The large amount of groundwater flow in streams in the watershed is believed to be one reason why the Darby creeks achieve high levels of water quality. Ohio EPA documented noticeable groundwater contributions from the Cable Moraine Complex in Logan, Champaign, and Madison County to portions of Upper Big Darby Creek, Little Darby Creek, Clover Run, Hay Run, Pleasant Run, and

Spain Creek. The London Moraine influences groundwater flow to Gay Run, Smith Ditch, and Springwater Run in eastern Madison and southern Franklin counties.

Groundwater resources in the Big Darby watershed vary considerably. The highest yields are seen in the Big Darby flood plain, extending from I-70 south to the confluence with the Scioto River and the flood plain of Little Darby Creek, east of Mechanicsburg in Champaign County. These areas, most particularly the Big Darby flood plains, are underlain by the most extensive buried glacial valleys in the watershed.

Groundwater–Surface Water Interactions in the Big Darby Watershed

It is imperative to understand interactions between groundwater and surface water because the linkages and feedbacks between these two systems affect both the quantity and quality of available water to meet human and ecosystem needs.

From their study on development in the Big Darby watershed, Dufour et al. (2001) observed that human needs are expanding in the Big Darby watershed and this is evident in the projected increase in population (Appendix E) which is expected to double in the next two decades. This increase means an increase in water demands. Ecosystem needs are being recognized, as in stream flow programs attempt to establish minimum flow requirements to maintain healthy ecosystems. Because groundwater and surface water form a single resource, factors such as development or contamination of groundwater may impact surface water and vice versa. Increased development of groundwater can change streams from gaining to losing status, affecting the quantity of surface water available for water rights and in stream flows. Contamination of groundwater can impact nearby surface water bodies where groundwater discharges to surface water. Therefore, quantitative assessment of the existence of groundwater–surface water interactions is important. This section will present an analysis of the interactions between surface water and groundwater in the Big Darby watershed. This will consist of quantitative analyses of hydrological parameters as well as review of available literature. The main aim is to identify the existence of surface–groundwater interactions within the Big Darby watershed and, thus, the two major questions to be answered are:

1. What work has been done in the watershed on groundwater–surface water interactions?
2. Do the physical processes in the watershed exhibit the existence of interaction between surface water systems and groundwater systems to warrant the development of a coupled groundwater–surface water model for the watershed?

A joint study by the Ohio Department of Natural Resources (ODNR) and USGS analyzed flow data in Ohio to determine groundwater recharge rates and is summarized in the report, *Use of Stream Flow Records and Basin Characteristics to Estimate Groundwater Recharge Rates in Ohio* (Dumouchelle and Schiefer 2002). The report estimates groundwater recharge rates and the mean baseflow to mean stream flow ratio for USGS gages in Ohio. The three active gages in the Big Darby watershed were included in the report. The hydrologic cycle for the Big Darby Creek and its subwatersheds was simulated using the Generalized Watershed Loading Function or GWLF model (Haith, Mandel, and Wu, 1992). The model predicts stream flow based on precipitation, evapotranspiration, land uses, and soil characteristics. Groundwater recharge is determined by tracking daily water balances in the unsaturated and shallow saturated zones; these zones act as reservoirs and have inputs and outputs. The input to the unsaturated zone is the infiltrated water calculated as the amount of the precipitation received, less the surface runoff. Outputs of this zone include the moisture lost via plant root uptake (which is lost to the atmosphere in a process called evapotranspiration) and percolation down to the saturated zone.

Evapotranspiration is estimated based on the available moisture in the unsaturated zone, the potential evapotranspiration based on day length and temperature, and a cover coefficient based on the type of plant or crop in the area of interest. Percolation occurs when the unsaturated zone volume exceeds the soil–water capacity; the shallow saturated zone receives the percolated water. This zone is treated as a linear reservoir. It can discharge water to the stream as baseflow or lose moisture to deep seepage.

Stream flow is computed as the sum of the groundwater discharge from the shallow saturated zone and the surface runoff. The model computes the daily water balance and resulting stream flow allowing comparison of the GWLF-predicted values to a daily record of stream flow such as is collected at USGS flow gages. The findings for these USGS gages based on this study and the GWLF results are also shown in Table 5.3.

Table 5.3. Baseflow and stream flow from the GWLF model and USGS data

Gage #	Stream	R^2 Value	Predicted to Observed Ratio	ODNR/USGS study % of Mean Stream flow		GWLF Model % of Mean Stream flow	
				Baseflow	Runoff	Baseflow	Runoff
03230310	Little Darby Ck	0.883	1.02	50.8	49.2	51.2	48.8
03230450	Hellbranch Run	0.884	0.99	41.2	58.8	41.4	58.6
03230500	Big Darby Ck	0.895	1.03	46.2	53.8	46.5	53.5

Table 5.3 shows that baseflow contributes half of the total stream flow recorded at all three gaging stations. This clearly shows that groundwater influences surface flow to a large extent. The GWLF model also shows the same trend, and this clearly shows the existence of groundwater–surface water interactions within the watershed.

Influence of hydrogeology on groundwater-surface water interactions

The purpose of this analysis is to identify watershed areas with the potential to receive a proportionately greater groundwater contribution to stream flow than the watershed as a whole. The analysis is based upon a characterization of the physical properties of the underlying subsurface material. The material that composes the watershed subsurface varies downward along its vertical profile and laterally across the watershed surface. It is this spatial variation that partially explains the relative presence or absence of groundwater in stream flow. This section summarizes results from the Ohio EPA (2005) study on the glacial and bedrock water resources of the Big Darby subwatershed, and the effect they have on groundwater surface water interactions. The geological make up of the underlying strata in the watershed is classifies in order of increasing hydraulic conductivity as till, till with sand and gravel, fines with sand and gravel, sand and gravel with till, and sand and gravel with fines. Estimated yields given provide a measure of the aquifer's capability to provide water for wells. While hydraulic conductivity is a better measure of the connectivity between ground and surface water, yield can be useful if its limitations within this context are considered. Spatial variability of yield is only significant when it results from a change in the local geology of the underlying deposits.

Upper Big Darby Creek watershed

According to the Ohio EPA report, the upper Big Darby Creek watershed is dominated by till with sand and gravel as shown in Figure 5.3. There is a general likelihood of high groundwater contributions to the streams because of the presence of streams that cut into the high yielding bedrock aquifers in the north western part of this sub watershed.

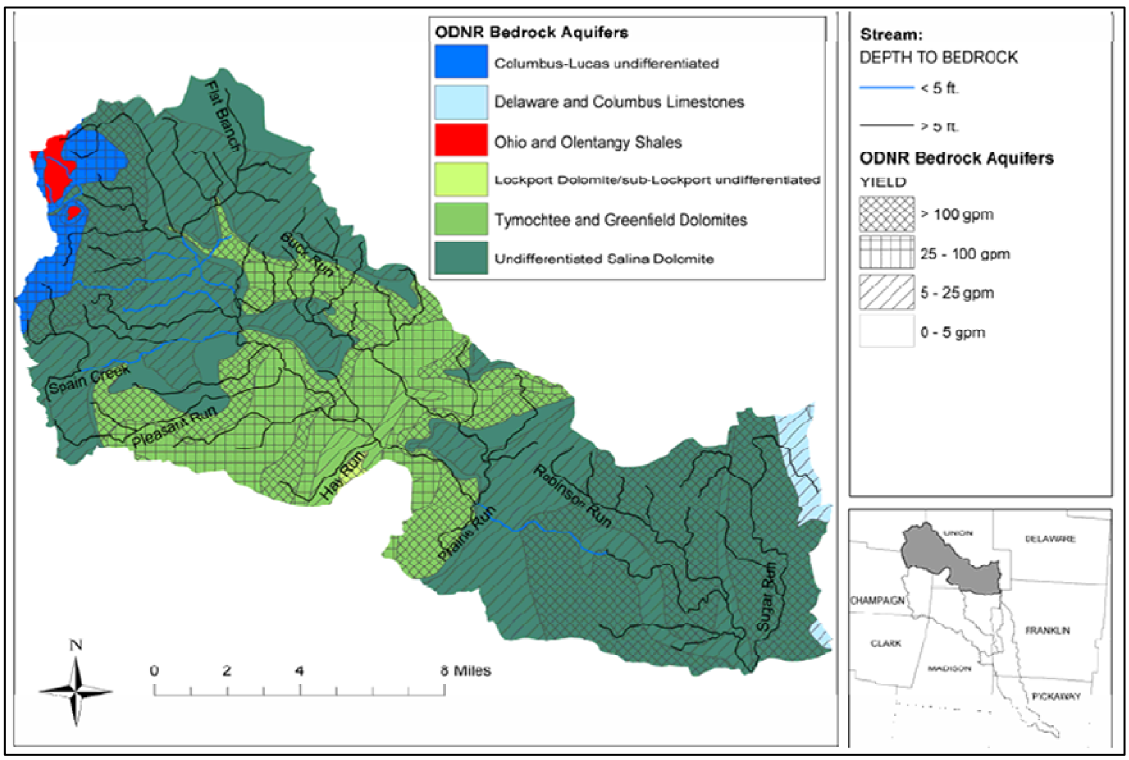


Figure 5.3. Bedrock aquifer type and yield in the upper Big Darby Creek watershed (Ohio EPA, 2005).

In areas where the underlying bedrock transitions from one type to another, there is usually greater fracture and this gives a higher potential for water storage and movement. Even though the aquifer types in this sub watershed are low yielding there is potential for a greater groundwater contribution to the stream because of the convergence of multiple rock types in the area.

Middle Big Darby Creek watershed

The aquifers in this sub watershed are shown in Figure 5.4. The sub watershed is dominated by till with sand and gravel and alluvial fines with sand and gravel are found along the mainstem. Groundwater yields of the glacial till are low to moderate, and the alluvial fines have moderate yields. Bedrock beneath Fitzgerald Ditch is Tymochtee, Greenfield, and Salina dolomite of moderate to high yield and this may lead to a greater groundwater contribution to this stream. Bedrock beneath the mainstem segment is low-yielding limestone, and offers little potential for a significant contribution.

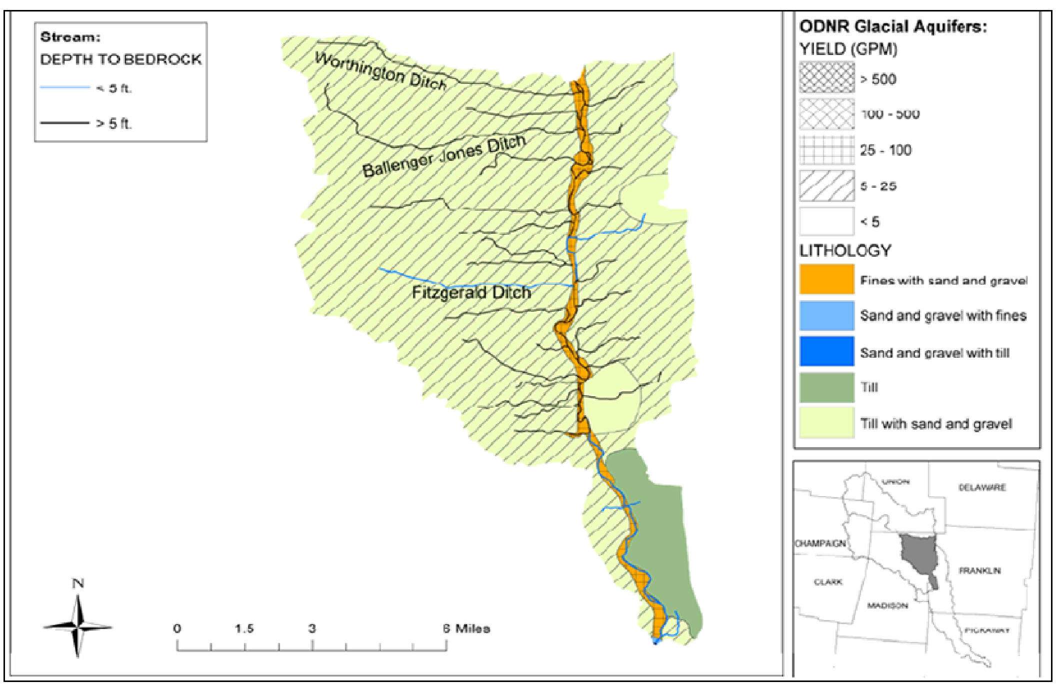


Figure 5.4. Bedrock aquifer type and yield in the middle Big Darby Creek watershed (Ohio EPA, 2005).

Lower Big Darby Creek watershed

Figure 5.5 shows the underlying aquifers in the lower Big Darby Creek. The valleys are the area of greatest potential for a large groundwater contribution in the entire watershed. There is a great potential for percolation to the aquifer and lateral transport to the stream due to the high hydraulic conductivity of the sand and gravel as well as the greater permeability of the watershed soils. Due to these factors, groundwater is likely a large component of stream flow during dry periods. Figure 5.5 also shows that there is little connection between the streams and bedrock in the lower areas of the subwatershed.

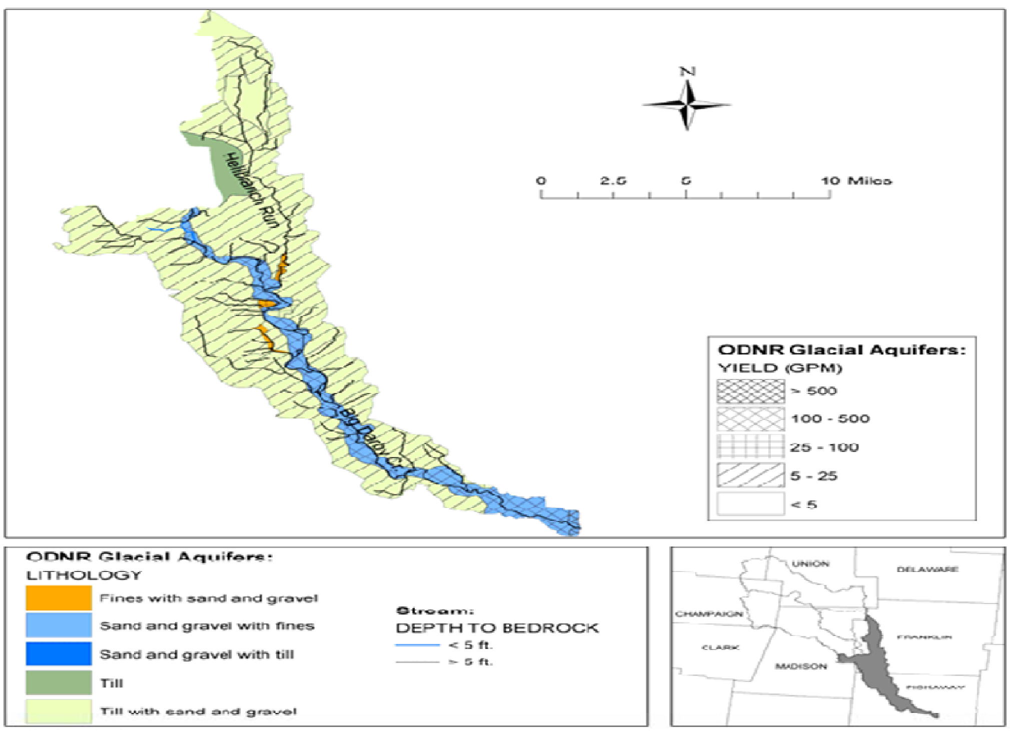


Figure 5.5. Aquifer types and yield in the lower Big Darby Creek watershed (Ohio EPA, 2005)

<u>Little Darby Creek subwatershed</u>

The lithology and yield for the Little Darby Creek subwatershed are illustrated in Figure 5.6. This sub watershed is dominated by till with sand and gravel, which have low to moderate yields. The headwaters of the Little Darby Creek runs through sand and gravel, sand and gravel with till, and sand and gravel with fines in a buried valley setting. The sand and gravel have higher hydraulic conductivities than the surrounding deposits, and thus, a greater groundwater contribution to stream flow is possible. Alluvial fines with sand and gravel exist in the middle and are characterized by moderate yields. As the Little Darby Creek reaches its confluence with the Big Darby Creek, the subsurface deposits change from alluvial to buried valley leading to underlying strata that is coarser and having higher conductivities and higher ground water contribution to stream flow.

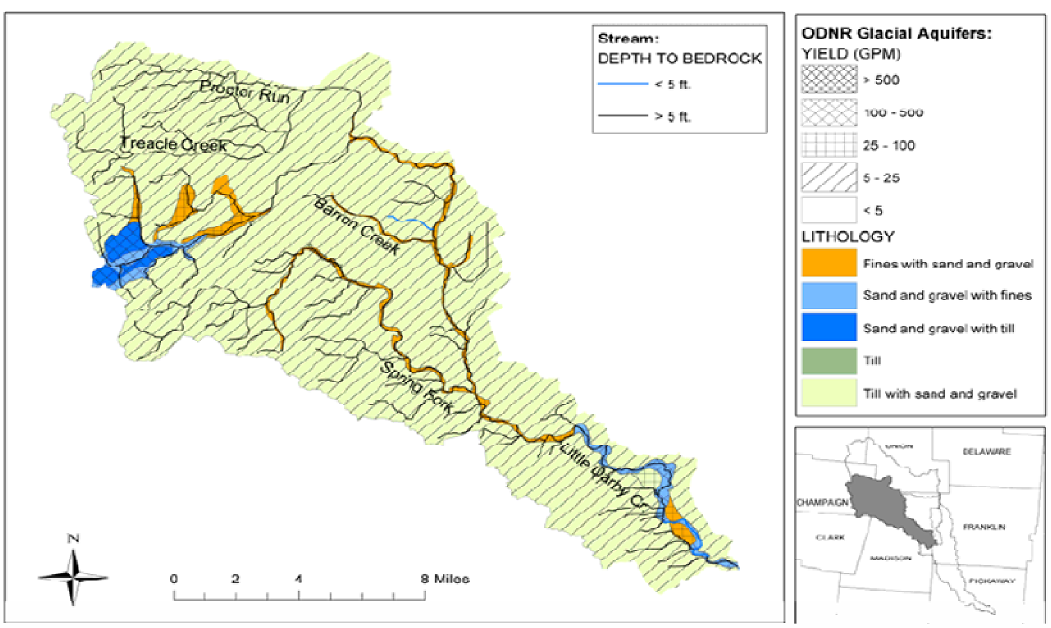


Figure 5.6. Bedrock aquifer type and yield in the Little Darby Creek watershed (Ohio EPA, 2005).

Groundwater Model for the Big Darby Watershed Using MODFLOW

This section outlines the development and calibration of a groundwater flow model which will subsequently be used in conjunction with a surface water simulation model to understand basin wide water dynamics for the Big Darby watershed, Ohio. A conceptual model of the watershed was constructed on the basis of hydrogeologic data collected during this investigation. A steady state numerical model, developed on the basis of the conceptual model, was calibrated and optimized using MODFLOW-96. The steady state simulation was used to obtain initial parameters that were used in the transient simulation. The main question to be addressed by this study was to evaluate how a MODFLOW based ground water movement model simulates spatial and temporal patterns of groundwater movement in the Big Darby watershed.

Methodology

Construction of the groundwater flow model was accomplished by discretization of the hydrologic properties of the groundwater system; establishment of model boundaries that represent conceptual hydrologic boundaries; determination of recharge rates and groundwater withdrawal rates for the steady state simulation and each stress period of the transient simulation; and assignment of model parameters to recharge, discharge, and hydrologic properties. The model chosen was the United States Geological Survey (USGS) finite difference model, MODFLOW (McDonald & Harbaugh, 1988). This package was selected because it is widely used and accepted within the hydrogeological community (Osiensky & Williams, 1997; Anderson and Woessner, 1992).

Conceptual model of groundwater flow in the Big Darby watershed

The Big Darby watershed represents a single hydrogeologic basin because no subsurface flow occurs across the periphery of the basin. Recharge to the Big Darby watershed occurs mainly in the form of infiltration from precipitation. The water table mimics topography, such that surface and groundwater divides typically coincide, especially in uplands. Precipitation recharges groundwater in till and bedrock upland areas and in the stratified glacial deposits; surface runoff from uplands also recharges the stratified glacial deposits at the edges of valleys. Groundwater levels and flow directions, particularly in the stratified glacial deposits, are strongly influenced by the locations and elevations of streams which, along with wetlands and pumping wells, are the discharge points for the groundwater flow system.

For the groundwater flow simulations, the Big Darby watershed was conceptualized using a single layer model. The layer represents the unconfined portion of the White Limestone aquifer and the alluvial aquifer. Hydrologic stresses represented in the model are net recharge and discharge. Recharge to the model layer was estimated by subtracting the total estimated evapotranspiration and surface runoff from total precipitation. Local variations in recharge occur due to lithology, slope, and geology. Additional recharge to the aquifer can occur along streams and rivers, subsurface inflow from neighboring basins, and leakage from ponds, lakes, and reservoirs. Additional discharge can occur as baseflow to streams and rivers, although this amount is small in comparison to subsurface discharge.

Simplifying assumptions

By definition, a model is a mathematical representation of a process or a system. In that regard, a single layer was used to represent the aquifer in the model described in this dissertation. Other simplifying assumptions in the model include: 1) the system is isotropic, causing hydrologic properties to be spatially invariant; 2) all pumpage in a model cell can be simulated as coming from the cell center; 3) the pumpage throughout a stress period is applied equally throughout the stress period; 4) recharge is invariant over large periods of time; and 5) small scale variations of hydraulic conductivity within cells are negligible. It was also assumed that the sediments that comprise the unconsolidated material in the model layer and the system of fractures in the bedrock that supply a majority of wells in the area transmit water as an equivalent porous media.

Darcy's Law can then be assumed to apply to groundwater flow, and the use of MODFLOW to simulate this flow system is thus appropriate. This assumption has been made by other investigations, particularly in fractured or conduit-flow aquifers (Glenn et al., 1989; Nelson, 1989), and is valid due to the scale of the model. At a regional scale, the fractures in the system are assumed to represent the primary porosity of the system and approximate the porosity of a continuous porous medium at a regional perspective.

Model design

Model design represents the process of translating the conceptual model for groundwater flow in the aquifer into a numerical representation which is generally described as the model. The conceptual flow model defines the required processes and attributes for the code to be used. In addition to selection of the appropriate code, model design includes definition of the model grid and layer structure, the model boundary conditions, and the model hydraulic parameters. Each of these elements of model design and their implementation are described in the remainder of this section.

Spatial discretization of model grid

The geographic boundaries of the model grid were determined by using a map created in Arc Map covering the extent of the watershed. This map represents the areal extent of the physiographic region. A finite difference grid superimposed over the 554 square miles study area was designed and constructed based on the simplification of a conceptual model representing the physical properties of the groundwater system. The physiographic boundaries of the ground water model were set to coincide with the boundary of the watershed. The grid network was a constant spacing of 500 m by 500 m. A total of 166 rows and 121columns and 5417 cells are used to cover the study area. The single model layer slopes with the land surface, and thickness is highly varied

Model data input

The finite difference model was developed by incorporating geologic data and measured and inferred hydrologic data for the period 1998. A contour map of the potentiometric surfaces of the aquifer were developed and are based on the interpolation and extrapolation of heads from measured points. The map was used to provide an initial specified head as reference elevations with which steady state heads could be calculated. Input parameters to the model included horizontal hydraulic conductivity of the aquifer; initial hydraulic head, estimated recharge values, streambed conductance, and layer aquifer thickness. Each active grid cell was assigned values according to its location within the study area.

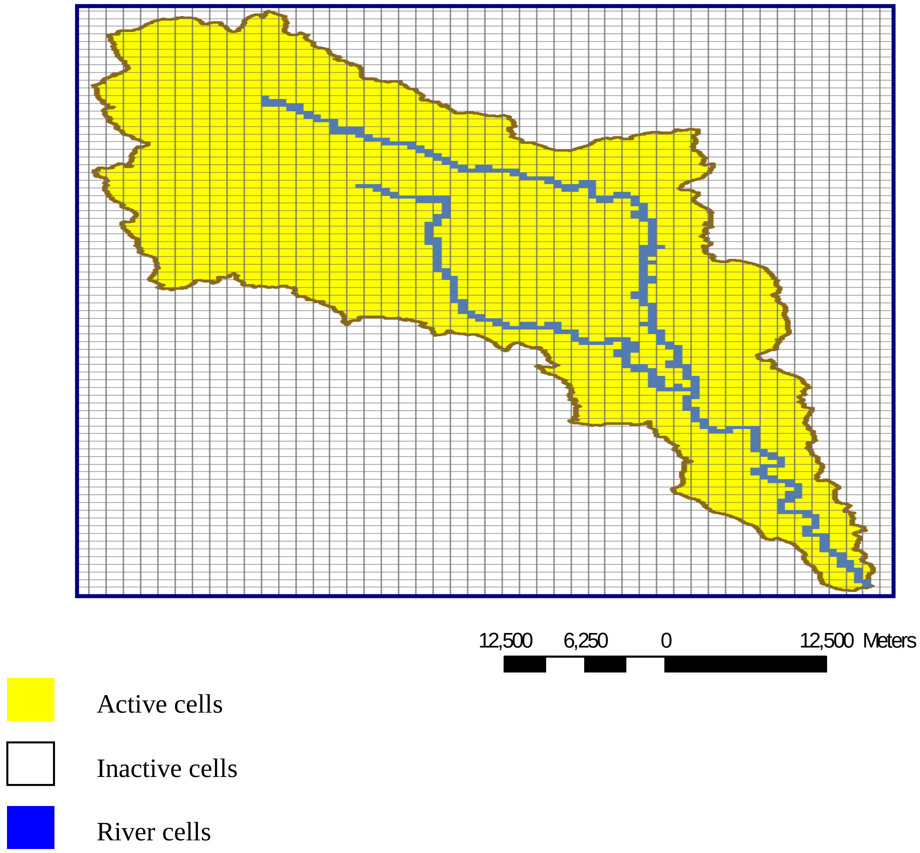


Figure 5.7. Model grid showing active and inactive cells and the river network.

<u>Initial net recharge</u>

Because an evaluation of groundwater availability is largely dependent upon recharge (Freeze, 1971), it is an important model input parameter warranting careful examination and meaningful implementation. In typical model applications, recharge is either homogeneously defined as a percentage of the yearly average precipitation or calibrated as an unknown parameter. However recharge varies spatially and temporarily because recharge is a complex function of precipitation rate and volume, soil type, water level and soil moisture, topography, and evapotranspiration (ET) (Freeze, 1969). It is therefore necessary in modeling to include spatially and temporarily varying recharge. Recharge data for MODFLOW were obtained from the Ohio Division of Water report (1965), Stowe (1979), and Garner (1983). Recharge rates provided by Sheets and Yost (1994) for the Mad River in Clark County were also found to be applicable for similar terrains in Franklin County. Values of 18cm to 25cm per year of recharge were assigned to areas with highly permeable soils (e.g. sandy loams) and vadose materials (e.g. outwash), shallow depths to water, and relatively flat topography. These areas typically occur along terraces or floodplains flanking the streams.

Model hydraulic parameters

For the steady state model, the primary parameter to be estimated and distributed across the model grid is hydraulic conductivity. For the transient model, we must add the storage coefficient. Data for hydraulic conductivity were derived from transmissivity data from the Ohio Division of Water (1965), Bennett and Williams (1988), and Eagon (1988). Values for hydraulic conductivity were calculated by taking the transmissivity and dividing by an estimated (or given) value for the saturated thickness. In some reports, actual data for hydraulic conductivity or permeability were given. Textbook tables (Freeze and Cherry, 1979; Fetter, 1980; Driscoll, 1986) were useful in obtaining estimated hydraulic conductivity values for a variety of sediments.

Boundary and initial conditions

A boundary condition can be defined as a constraint put on the active model grid to characterize the interaction between the active simulation grid domain and the surrounding environment. There are generally three types of boundary conditions; specified head (First Type or Dirichlet), specified flow (Second Type or Neumann), and head-dependent flow (Third Type or Cauchy). The no-flow boundary condition is a special case of the specified flow boundary condition. Boundaries can be defined as being time independent or time dependent.

For this research the altitude of the top of each model cell was set equal to the altitude of land surface obtained from the DEM. The extent of the model area was defined with no-flow cells along the topographic divide of the Big Darby watershed. In a large river catchment such as the Big Darby watershed, it can be assumed that the surface watershed has the same extent as the subsurface groundwater catchment. Therefore no-flow boundaries were assigned all around the model domain. Many studies of local and regional groundwater systems (Faye and Mayer, 1990; Robinson et al., 1997), especially those conducted for small basins, are based on this fundamental assumption. This assumption sometimes is violated as reported by Tiedeman, Goode, and Hsieh (1998), in their study of fractured rock near Mirror Lake, New Hampshire. The boundaries in this investigation, however, will be assumed to be no-flow boundaries that correspond to the surface water drainage basin to minimize the variability in the input parameter set.

Groundwater monitoring piezometers

Piezometer readings of groundwater head are used for verifying whether groundwater head predictions are relatively similar to measured heads. Within the Big Darby watershed there are fifteen piezometers with time series of depths to the water table. However, three of those have short data periods; therefore, only twelve piezometers were used in this research. These twelve have data for the period 1988-2000. Usually piezometer readings provide depth to groundwater table information that needs to be changed to groundwater head using surface level information. Figure 5.8 shows the location of these piezometers in the watershed.

Groundwater pumping

Public water supplies in the watershed are all supplied by groundwater sources expect for the city of Marysville which utilizes surface water from Mill Creek for two thirds of their drinking water supply. The Darby Creek watershed has no surface water removal for water supplies. The Ohio Environmental Protection Agency (OEPA) created a database providing the name and location of all public water supplies in the state. A total of 171 public water supplies exist in the watershed and their location is shown in a series of maps by county in Appendix A. The majority of the population within the Darby Creek watershed uses groundwater as its source of water. Groundwater is extracted at an annual rate of $3.8 *10^{7}$ m^{3}.

The monthly pumping distribution for transient simulations is determined from seasonal pumping schedules given by the OEPA. Monthly pumping rates range from a minimum of $4.5 *10^{-2}$ m^{3}/s per well during the months of November through February to a maximum of $8.4*10^{-1}$ m^{3}/s per well during the months May through August.

Model calibration

Model calibration was accomplished by varying the model-input parameters within plausible ranges to produce the best fit between simulated and observed hydraulic heads in the watershed. Steady-state and transient simulations were analyzed to determine the best combination of model-input parameters. Water-level measurements for 12 wells that were used for estimation of potentiometric surfaces of the watershed aquifer were considered for calibration of the steady-state model. The numerical model was used to simulate average (steady-state) flow conditions for the 10-year period (1992-2002). Steady-state conditions were numerically approximated for the steady-state simulation by using a transient simulation with one 20-year stress period using constant input flow rates.

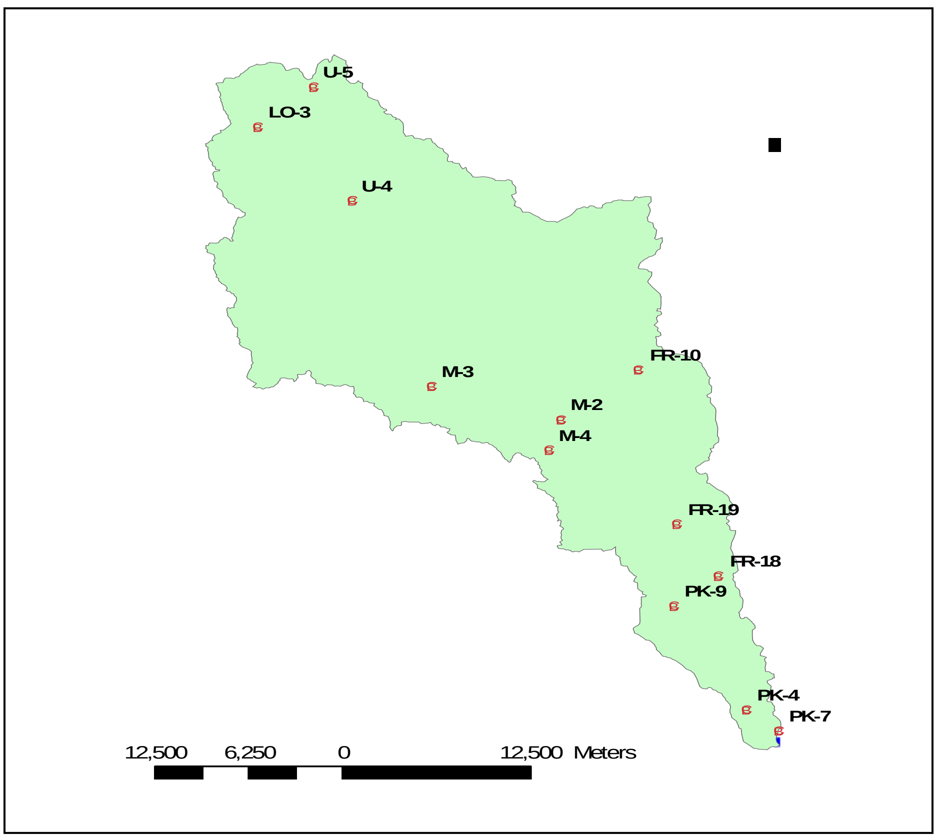


Figure 5.8. Location of groundwater monitoring wells in the study area.

The hydraulic head surface from the steady-state simulation established initial conditions for the transient simulation. Calibration requires development of calibration targets and specification of calibration measures. The primary calibration target is hydraulic head. Simulated heads were compared to measured heads at specific observation points through time.

The root mean square (RMS) error is the average of the squared differences between measured heads (*hm*) and simulated heads (*hs*):

$$RMS=\left[\frac{1}{n}\sum_{i=1}^{n}\left(h_m-h_s\right)_i^2\right]^{0.5} \qquad (5.1)$$

where *n* is the number of calibration measurements. The difference between the measured hydraulic head and the simulated hydraulic head is termed a residual. The RMS was used as the basic measure of calibration for hydraulic heads.

Rainfall–Runoff Model Using TOPNET

This section describes the development and calibration of TOPNET, a precipitation–runoff model for the Big Darby watershed. In general terms, the model was developed by: 1) compiling, collecting and processing needed data, 2) creating a model structure that represents the basin, 3) calibrating the model, and 4) evaluating its performance.

A 30m DEM obtained from the National Elevation Dataset (NED) was used to delineate streams and subbasins for the Big Darby Watershed. This was done using the

Terrain Analysis Using Digital Elevation Models (TAUDEM) software (Tarboton, 2002). TAUDEM was also used to calculate flow directions and contributing areas for each grid cell. Streams were then delineated using the DEM curvature based method. The obtained river network was pruned down by removing some of the higher order streams. This resulted in fewer model elements and a drainage density of $0.7 km^{-1}$. Figure 5.9 shows the subbasins that were delineated for the watershed. These were used as the model elements in TOPNET.

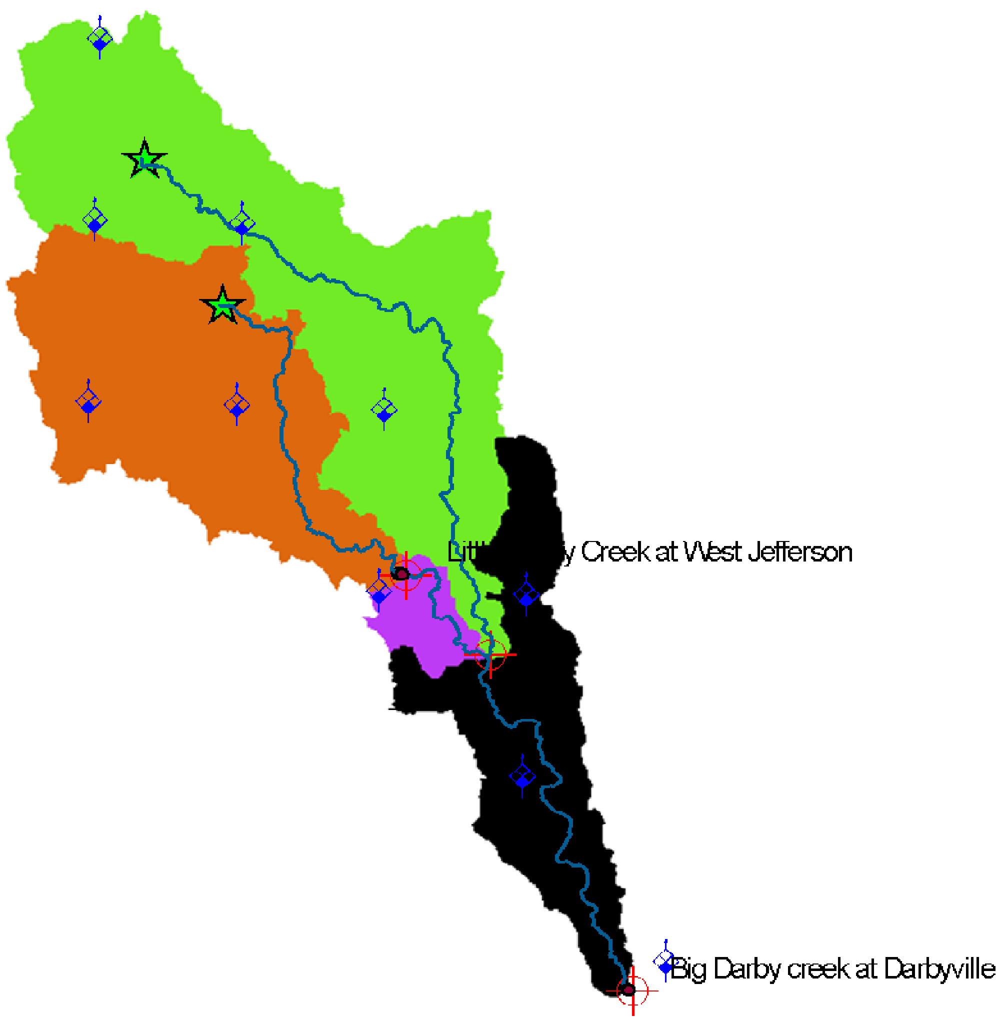


Figure 5.9. Delineated subbasins for the Big Darby watershed.

TAUDEM procedures were used to obtain slope and specific catchment area, *a*; for each grid cell in the DEM. The distribution of wetness index, $\ln(a/\tan\beta)$; for each sub-basin is represented using a histogram of the relative areal fraction of the sub-basin within each wetness index class.

Temporal inputs

The following meteorological forcings are used to run TOPNET. Daily gridded meteorological data was obtained from the Surface Water Modeling group at the University of Washington. The data can be downloaded from their web site at http://www.hydro.washington.edu/Lettenmaier/gridded_data/, the development of which is described by Maurer et al. (2002).

Precipitation. Daily precipitation in mm was based on station observations, where available. In data sparse areas, observations were filled in by interpolation using the nearest neighbor analysis method.

Temperature. Daily maximum and minimum temperature were used for each rain gage station. Dew point temperature was also utilized.

Vapor pressure and wind speed. The Mean daily vapor pressure was used as a driving input to the TOPNET model. Kimball et al. (1997) presented a procedure to calculate mean daily vapor pressure based on the daily minimum temperature, precipitation and shortwave radiation. Daily wind speed in m/s for each rain gauge station is also required by the model.

To prepare TOPNET inputs, soils and land use grids were utilized. These grids were resampled to the same 30m grid as the digital elevation model. A look-up table was used to associate a model parameter value for each grid cell. The 11-layer standard soil depth grid used by the Pennsylvania State University to grid NRCS STATSGO database was used. Figure 5.10 shows the soils map created in ArcMap for Big Darby watershed. Table 5.4 gives the names of each of the soil groups identified in Figure 5.10.

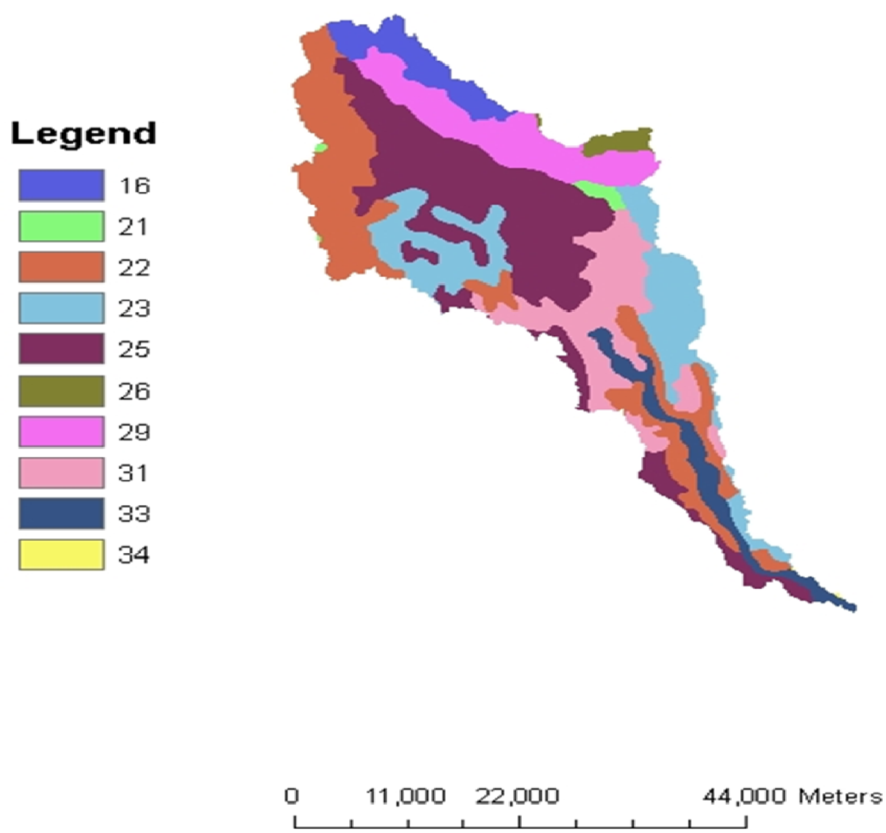


Figure 5.10. Delineated soil map for the Big Darby watershed.

Table 5.4. Big Darby watershed soil association names.

Map Unit ID	Soil Association Name
31	Kokomo Crosby-Miamian
33	Miamian Celina-Crosby
23	Brookston Crosby Celina
25	Crosby Miamian Brookston
21	Blount Glynwood Morley
16	Nappanee St. Clair Paulding
28	Eldean Ockley Sleeth
22	Blount Pewamo Glynwood
29	Eldean Westland Patton
34	Miamian Eldean Crosby

A detailed description of the procedures to estimate these parameters is outlined in Bandaragoda, Tarboton, and Ross (2004). The Matlab program, FITF (Woods, 2003) was used to obtain the soil parameter values. These parameters are shown in Table 5.5.

Table 5.5. TOPNET soils parameters estimated using FITF.

Parameter	**Definition**
$\Delta\theta_1$	Drainable porosity
$\Delta\theta_2$	Plant available porosity
F	Saturated store sensitivity
Ko	Surface Saturated Hydraulic conductivity
Ψ_f	Wetting front suction

Parameter estimation using FITF, a MATLAB routine, is based on 11-layer soils polygons, which have soil texture information in each soil layer, and a lookup table developed by Clapp and Hornberger (1978). The saturated store sensitivity parameter, f , is estimated for each polygon by first assigning a saturated hydraulic conductivity, Ksat, at each soil layer. An equation of the form:

$$Ksat(z)=K0*exp(-f*z). \qquad (5.2)$$

is fitted where K0 is initial estimate of hydraulic conductivity. The saturated store sensitivity parameter is obtained from the linear regression:

$$\log (Ksat(z))=\log(K0) + (-f)*z . \qquad (5.3)$$

The K0 from the regression is used as the TOPMODEL parameter value.

The values for $\Delta\theta_1$, $\Delta\theta_2$ and ψ_f are averages down to but not including the first occurrence of bedrock. Table 5.6 shows the obtained values for these parameters for each soil class in the Big Darby watershed.

Parameter values for lapse rate, soil zone drainage sensitivity, and hydraulic geometry were left at the default values set in TOPNET. Parameter values for land use are given in Table 5.7.

Table 5.6. Soil parameter values utilized by the TOPSETUP program

ZONE	f(1/m)	K0(m/h)	$\Delta\theta_1$	$\Delta\theta_2$	ψ_f (m)
OH016	1.64	0.0152117	0.095	0.146	0.6
OH029	2.57	0.0126793	0.111	0.130	0.4
OHO33	0.57	0.019244	0.111	0.165	0.7
OHO23	0.57	0.00537091	0.106	0.141	0.5
OHO31	0.57	0.0190243	0.115	0.165	0.6
OHO22	1.21	0.0174466	0.119	0.155	0.5
OHO21	1.15	0.017648	0.115	0.155	0.6
OHO25	0.57	0.0198842	0.117	0.167	0.6
OHO26	3.62	0.0298052	0.104	0.144	0.5
OHO34	2.54	0.0344478	0.109	0.158	0.6

Table 5.7. TOPSETUP Land use parameter values

Vegetation Class	CC (m)	CR	Albedo	Description
1	0	1	0.08	water
2	0.001	1	0.3	Croplands
3	0.001	1	0.3	Croplands
4	0.001	1	0.1	Grasslands
5	0.0015	1.5	0.2	Urban
7	0.0015	1.5	0.2	Woodland
8	0.003	3	0.14	Forest
9	0	1	0.23	Unclassified

Model calibration and verification

The hydrologic components of TOPNET were calibrated to fit the observed daily stream flow data from two USGS stream flow gauging stations: Little Darby at Jefferson (03230310) and Big Darby at Darbyville (03230500) for the years 1989-2000. This period was chosen because it represents a combination of dry, average, and wet years within the watershed. The model was run for the 11-year period of 1989-2000, but the first two years (1989 and 1990) were used for stabilization of model runs, and simulated stream flow for 1991-2000 was only used for comparison purposes. Values of selected model parameters were varied iteratively within a reasonable range during various calibration runs until the difference between observed and simulated stream flow data was minimized. The calibrated parameters were then verified by using an independent set of stream flow data that was not used for model calibration. In this study, stream flow data for the five year period (1991-1996) was used for model calibration while data for the period (1996-2000) for the same USGS gauging stations was used for model verification.

During calibration, as well as verification, agreement between observed and simulated stream flow data on a daily basis was determined using subjective as well as quantitative measures. The fit between daily observed and simulated stream flows was checked graphically by plotting the stream flow time series. General agreement between observed and simulated time series curves indicates adequate calibration over the range of the flow conditions simulated. Figures 5.11 and 5.12 show the time series of simulated and measured stream flow for the two gaging stations during the calibration period.

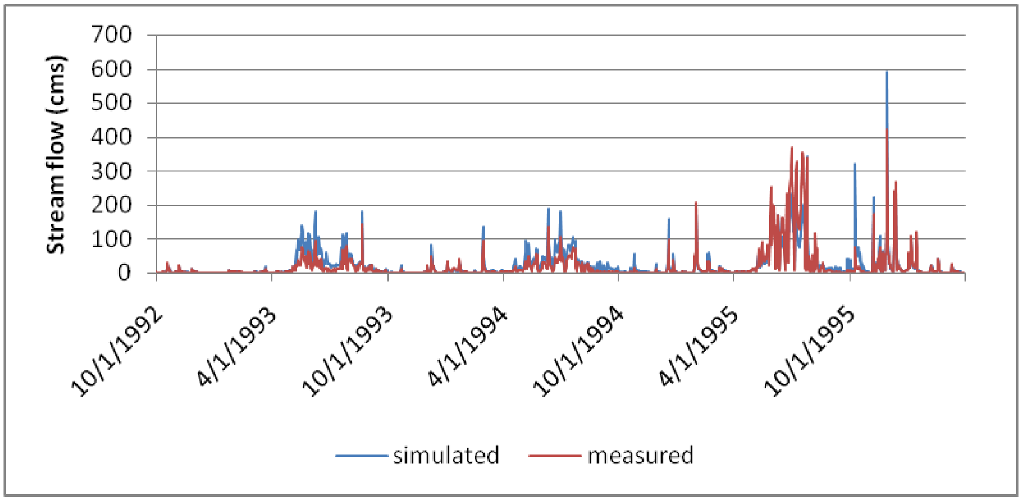


Figure 5.11. Measured and simulated stream flows at Darbyville station.

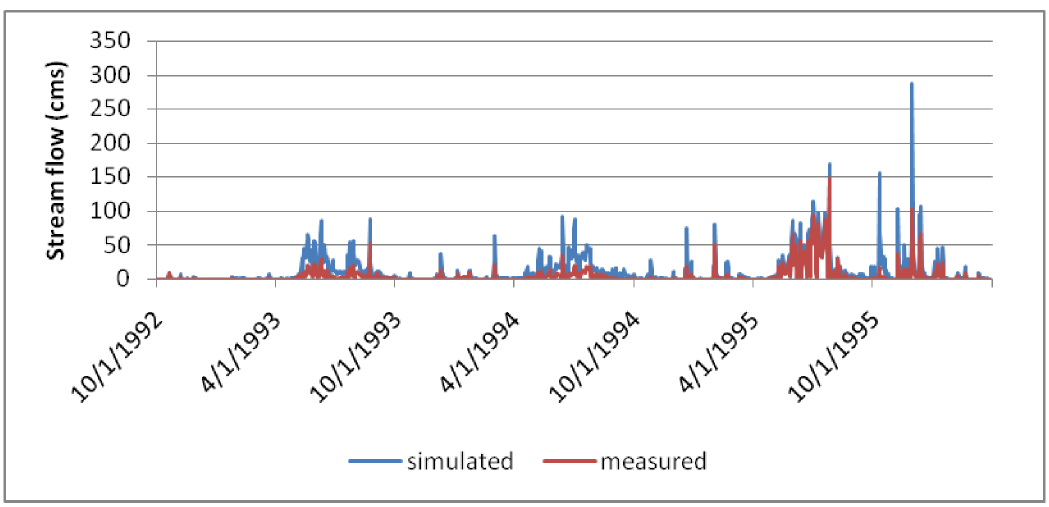


Figure 5.12. Measured and simulated stream flows at West Jefferson station.

Quantitative measures of agreement were based on observed and simulated mean daily stream flows and their standard deviations (SD), correlation coefficient (r), Nash-Sutcliffe model efficiency (NSE) (Nash and Sutcliffe, 1970), root mean squared error (RMSE), and mean absolute error (MAE). Both RMSE and MAE describe the difference between model simulations and observations in the units of the variable. Values close to zero indicate perfect fit; however, values less than half of the SD of the observations may be considered low.

The three sub basin model elements for the watershed have their own distinct model parameters and state variables which are derived from the soil and vegetation data. In the calibration process, as mentioned by Bandaragoda et al.(2004), the pattern of the spatial variability between subbasins is maintained by using multipliers for each parameter that are the same across all subbasins to scale the GIS derived sub basin parameters for each sub basin by the same factor.

The scatter plots for measured and simulated stream flows are are shown in Figures 5.13 and 5.14. During the 5-year calibration period the absolute value of the volume error between observed and TOPNET simulated annual stream flows was less than 10% in three years, and 10-15% in two years. Out of the five years, TOPNET under simulated (2.6% to 49.5%) the stream flow in four years and over simulated (-0.2 and -13.8%) it in one year for the Big Darby at the Darbyville gaging station. The model predicted mean monthly stream flows satisfactorily as indicated by an average correlation coefficient of 0.88 for the Darbyville gaging station and 0.90 for the Jefferson gaging station.

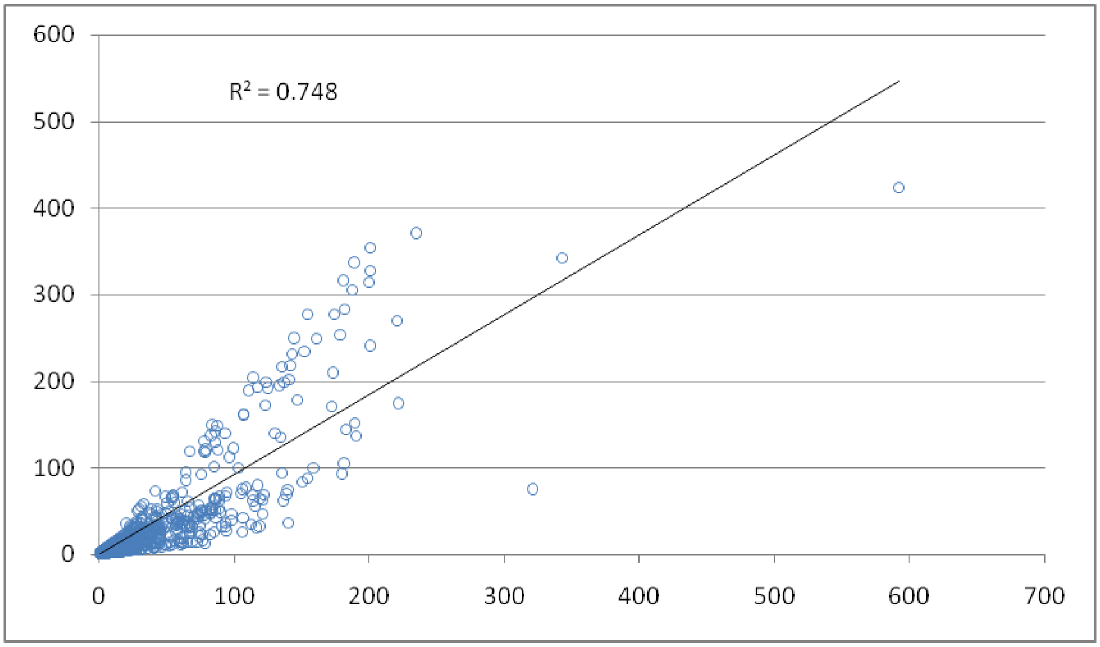


Figure 5.13. Scatter plot of measured and simulated stream flow at Darbyville.

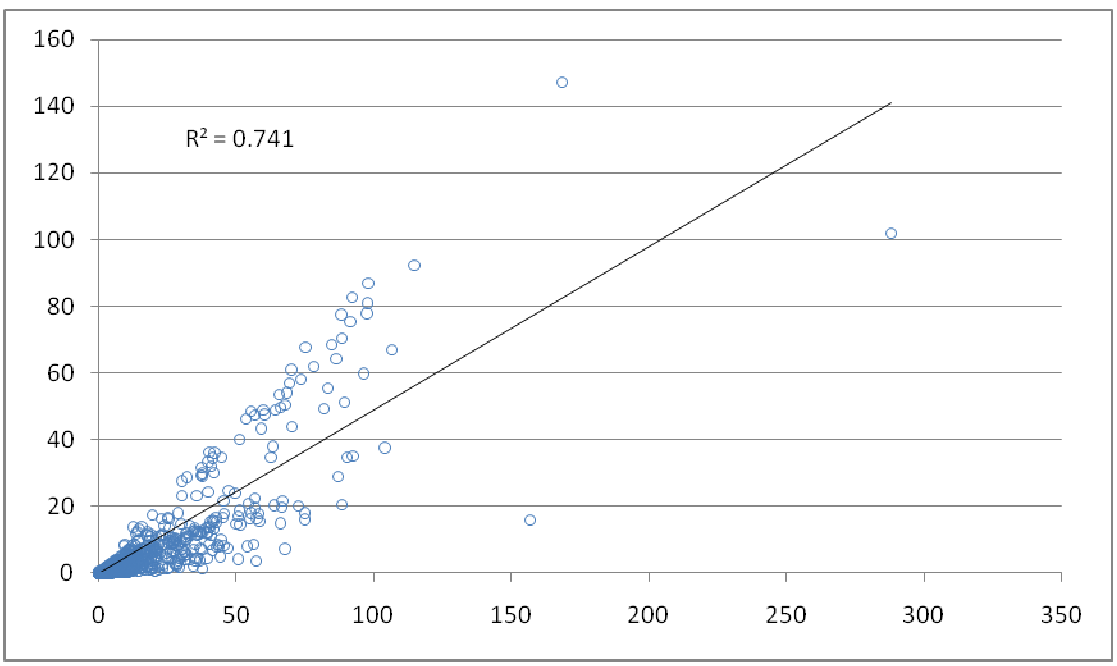


Figure 5.14. Scatter plot of measured and simulated stream flow at West Jefferson .

Visual comparison of observed and TOPNET simulated hydrographs (Figure 5.11 and 5.12) also shows this satisfactory agreement. Out of 60 months in the calibration period, the absolute volume error between observed and simulated mean monthly values was less than 10% in 44 months, 10-15% in nine months, and 15-25% in 7 months.

Despite the apparent under simulation, the rainfall runoff model was able to adequately describe the variations in stream flows with accurate simulation of peaks and lows as recorded at the two gaging stations as shown in Figures 5.11 and 5.12. The model was verified using data for the period 1996-2000. Figures 5.14 and 5.15 shows the times series obtained for the verification period and the corresponding correlation plots are shown in Figure 5.16 and 5.17 for the two gaging stations. The influence of snow melt on stream flow, which is not simulated here is the most likely reason for the under simulation of stream flow.

The model performed adequately as shown by an average correlation coefficient of 0.81 for the Darbyville gage and 0.79 for the Jefferson gaging station. Although these correlations are lower than those obtained during the calibration phase, the model showed good representation of the stream flow peaks and lows as shown in the plots in Figure 5.11 and Figure 5.12.

The model might also not have been able to very adequately model low flow stream flows. As outlined by Bandaragoda, Tarboton and, Ross (2004), TOPMODEL has a single function that models baseflow recession. Therefore any calibration of the TOPNET model has resulted in the adjustment of the sensitivity parameter f to match high flow recessions rather than low flow recessions.

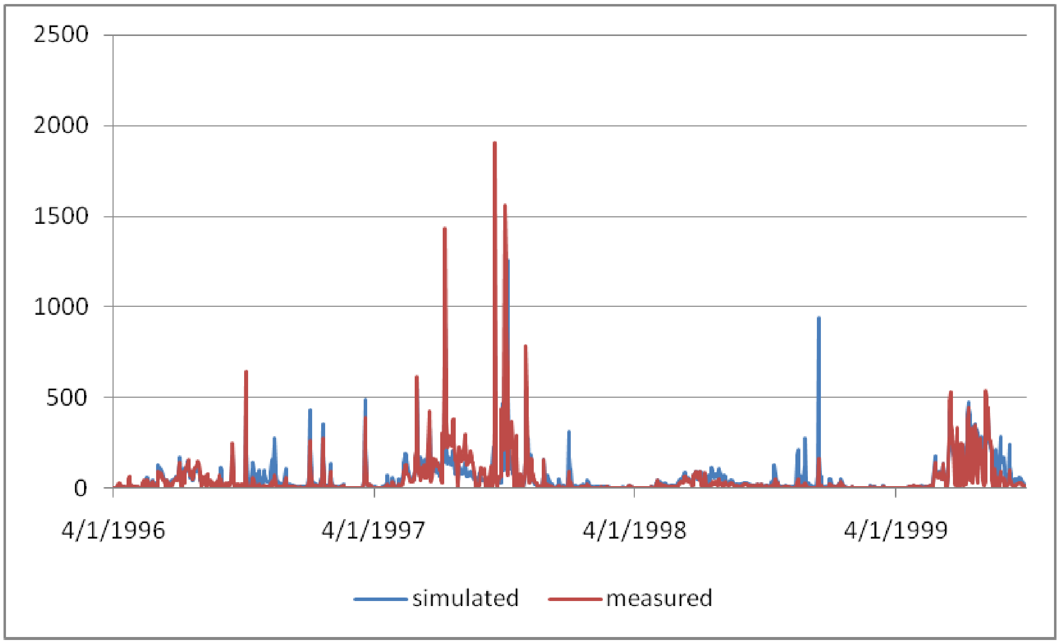


Figure 5.15. Measured and simulated stream flow at Darbyville during validation.

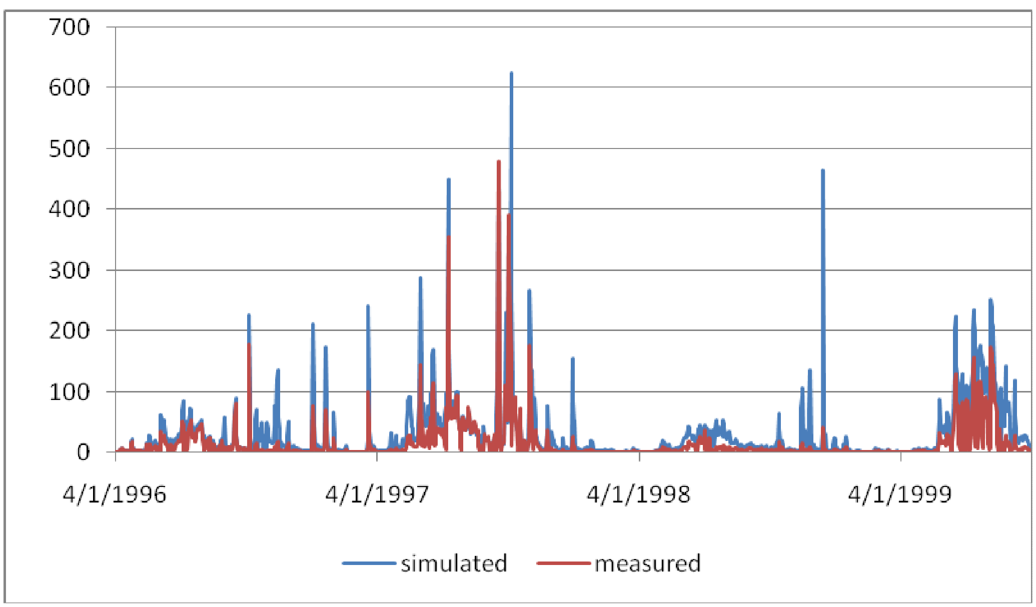


Figure 5.16. Measured and simulated stream flow West Jefferson during validation.

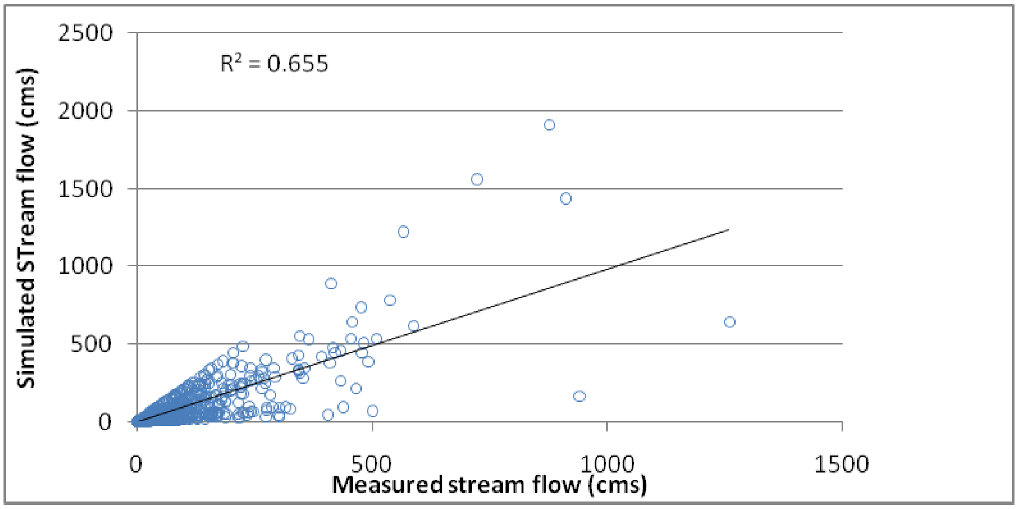


Figure 5.17a. Scatter plot for measured and simulated stream flow at Darbyville during validation period.

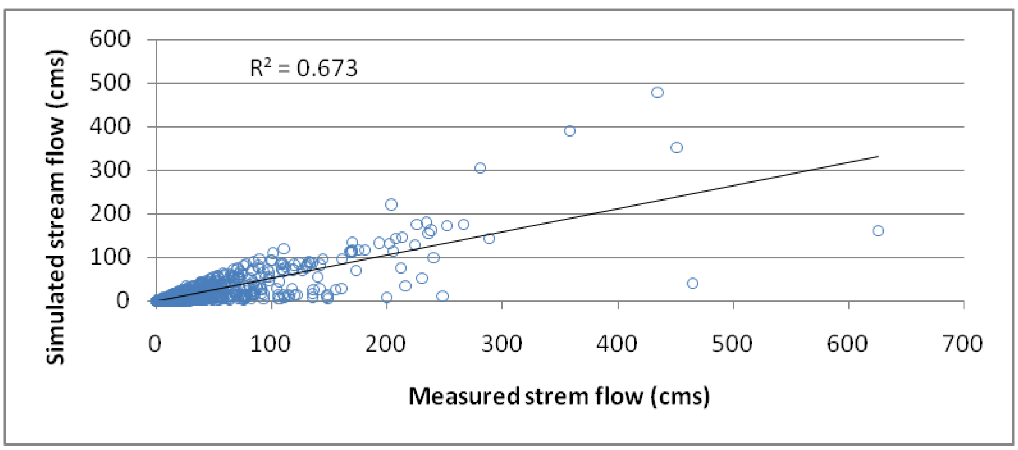


Figure 5.17 b. Scatter plot for measured and simulated stream flow at West Jefferson during validation period.

Developing the coupled groundwater–surface water simulation model

The unique characteristics of the hydrogeologic system of watersheds cause significant interactions between groundwater and surface water systems. Interaction processes involve infiltration, evapotranspiration (ET), runoff, and exchange of flow (seepage) between streams and aquifers. These interaction processes cannot be accurately simulated by either a surface water model or a groundwater model alone because surface water models generally oversimplify groundwater movement, and groundwater models generally oversimplify surface water movement. Estimates of the many components of flow between surface water and groundwater (such as recharge and ET) made by the two types of models are often inconsistent. This chapter will describe the model coupling concepts utilized in this research and also describe the basic principles of the Potential Coupling Interface tools (Bulatewicz and Cuny, 2005) for model coupling.

In the coupling of TOPNET with MODFLOW, the MODFLOW groundwater model grid points provide the depth to the water table. Therefore instead of depth to the water table measured based on wetness index, the depth to the water table for each groundwater model cell is passed from MODFLOW to TOPNET. As a result of this modification, water table depth calculations will be done for each groundwater model node, instead of the wetness index class. With this method, recharge to groundwater is modeled for each groundwater model cell. However recharge is an output of TOPNET and this variable is "lumped" over a watershed.

Therefore a sub routine that disintegrates the watershed wide recharge to specific ground water model cells is required.

In summary the interactions between MODFLOW and TOPNET proceed as follows:

(1) MODFLOW provides baseflow and depth to the water table (z) at each node to TOPNET;

(2) TOPNET uses the depth to the water table, root zone store depth, precipitation, evapotranspiration and snowmelt to determine stream flow ;

(3) TOPNET determines the net recharge to the saturated zone and passes it to MODFLOW and;

(4) MODFLOW uses the recharge to calculate ground water head variation in the watershed. This variation is both spatially and temporarily.

Therefore the coupling occurs in a cascading fashion.

Physical principles of model integration

In the coupled model, the surface water model, TOPNET, deals with the unsaturated zone and the surface process, whereas MODFLOW operates the saturated or groundwater zone. The vertical water flux from the saturated zone is calculated by TOPNET at every time step and forms the groundwater recharge to all active cells of MODFLOW. The runoff values of TOPNET are the sum of the surface water outflow (surface water runoff) and the groundwater outflow (baseflow).

Initially MODFLOW checks the type of simulation, the size of model grid, and the solution scheme to be used; reads parameter values, number of stress periods, initial heads, and boundary conditions; and allocates memory space while TOPNET reads and prepares information about hydrological and meteorological data, and reads parameter values. The two models are executed simultaneously. At the beginning of each stress period, TOPNET calculates and writes information about groundwater recharge from the unsaturated zone into the saturated zone top MODFLOW's input files.

In TOPNET, the depth of water held in the soil zone for each model element is calculated according to the equation:

$$\frac{dS_r}{dt}=I-E_s-R \qquad (5.4)$$

where I is the infiltration rate; E_s; soil evaporation rate; and R is the drainage rate or recharge to the saturated zone store from the soil store.

The recharge rate, R, is the same recharge rate, $I_{i,j,}$ used as input in MODFLOW, and from this rate:

$$Q_{R,i,j}=I_{i,j}\times DELR_j\times DELC_i \qquad (5.5)$$

where $Q_{R,i,j}$ is the flow rate applied to the model at a horizontal cell location i,j, $I_{i,j}$ is recharge flux in units of length per unit time and $DELR_j$ *$DELC_i$ is the area of a cell. Solving the partial differential equation of groundwater flow used in MODFLOW gives the finite difference equation for each groundwater grid cell.

Flow in the stream which is a combination of direct runoff and baseflow is routed in TOPNET using a kinematic wave routing algorithm (Goring, 1994). As outlined by Bandaragoda et al. (2004), the parameters used in the kinematic wave channel network routing are Manning's roughness parameter n, as well as width, slope and length for each channel segment. Slope and length are determined from the GIS based upon the DEM.

Channel width is determined as a power function of contributing area (Leopold and Maddock, 1953).

Model coupling description

Figure 5.18 provides a coupling description for the two models. This considers a unidirectional coupling as was done in the first case study; however, in this case study we also consider the situation where the TOPNET provides updated values of recharge to MODFLOW at each time step. Thus, we need to establish that link coupling.

The details of the coupling description for TOPNET–MODFLOW are similar to the details for the TOPMODEL–MODFLOW coupling, except for one significant difference: in TOPNET, the subcatchment loop is inside the time step loop (rather than the time step loop inside the subcatchment loop, as in TOPMODEL). This means that in the TOPNET–MODFLOW coupling, only one instance of each model needs to be used, which will simplify the coupling (recall that in the TOPMODEL–MODFLOW coupling we must use a separate instance of TOPMODEL for each subcatchment).

In the TOPNET–MODFLOW coupling, the water table heads (*HNEW*) are sent from MODFLOW at the start of each time step, and received by TOPNET at the start of each time step. At coupling point B, TOPNET overwrites the calculated water table depth (*ZBAR*) with the water table head received from MODFLOW at coupling point A. There is an update function at coupling point B that handles the mapping from the MODFLOW cells to the current subcatchment, and handles the unit conversions and depth calculation (based on the surface elevation).

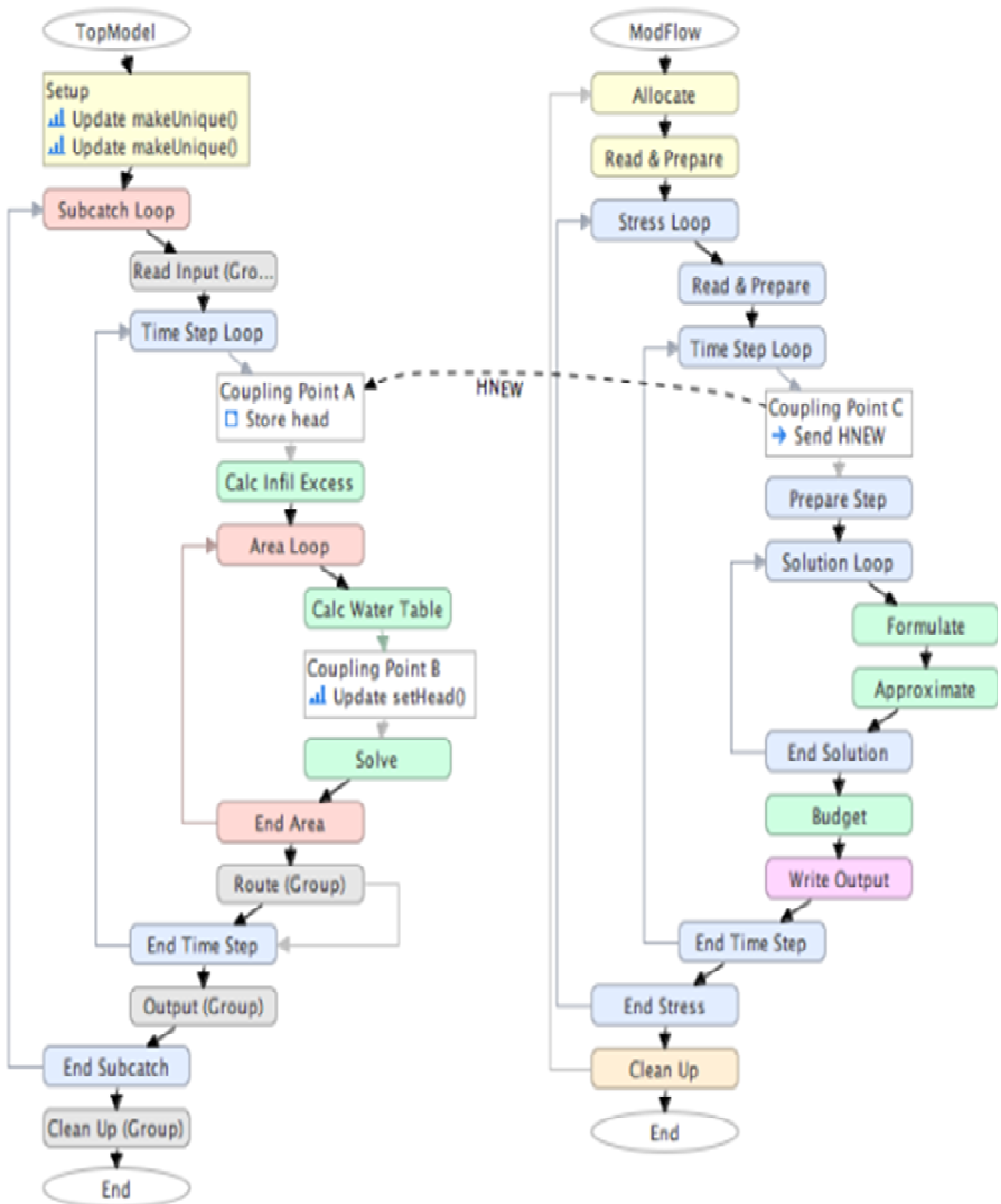


Figure 5.18. TOPNET–MODFLOW unidirectional coupling description.

Time step synchronization

The two models use different temporal discretization schemes. Generally surface water models use smaller time steps in the order of hours or days while ground water models use larger time steps. The smallest time step for groundwater modeling would be days, while use of monthly time steps is common. This is because of the low velocities of water movement in the sub surface compared to water movement in streams. Therefore in coupled groundwater surface water models it is important to synchronize the time steps in order to obtain reasonable results. In this study, a monthly time step was used for the groundwater and a daily time step for the surface water model. The synchronization of these different time steps is as shown in Figure 5.19a.

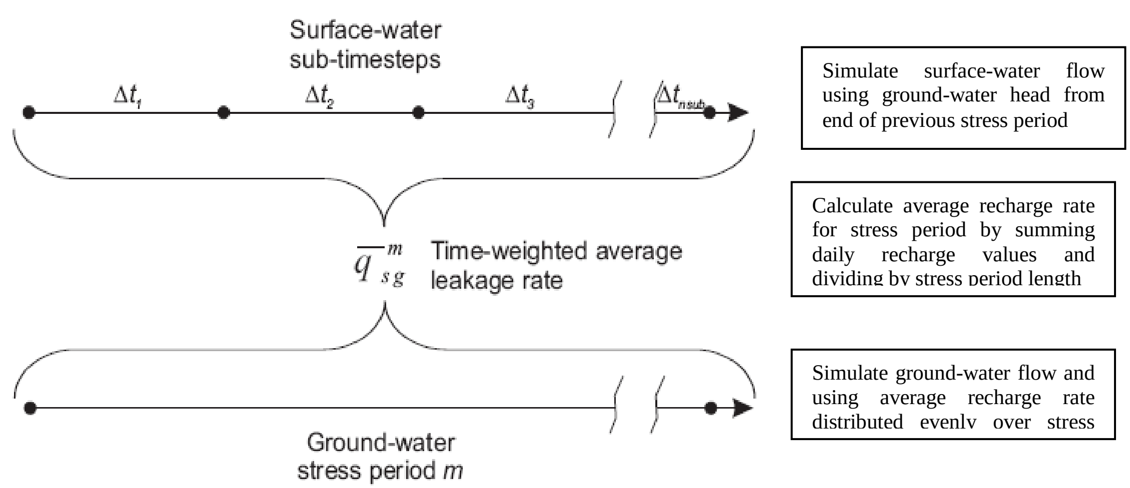


Figure 5.19a. Synchronization of time steps.

Connecting subwatersheds to Finite-Difference Cells

An important component to the coupling of the TOPNET and MODFLOW models is the process used to spatially link the subwatersheds used by TOPNET with the finite-difference cells used by MODFLOW. Thus two spatial conversions must be performed. Recharge calculated by TOPNET in a subcatchment must be distributed over the corresponding MODFLOW cells, and head values for the MODFLOW cells must be combined to produce a water table elevation for each subcatchment. GIS was used to join TOPNET subbasins to MODFLOW grid-cells by areally averaging the grid cells that fall within a particular subwatershed. Figure 5.19b shows the MODFLOW grid cells and the irregular polygons (subwatersheds) used by TOPNET. Appendix A shows the ArcGIS script developed by Bulatewicz (2006) that is used in areally averaging grid cells within a subwatershed.

Stepping through the rows and columns of the groundwater grid, the idea is to determine how many subcatchments each cell falls within and recording the subcatchment ID and the area of the portion of the cell within that subcatchment as illustrated in Figure 5.19b. A separate array of index values is also maintained that indicates how many ID/area pairs are recorded for each cell, and the total area of each subcatchment is stored as well.

Figure 5.19b. MODFLOW grid cells superimposed on TOPNET sub watersheds.

Results and Discussion of Stream Flow

Calibration Period

The Big Darby watershed coupled model was calibrated using a trial and error multi-target approach to reduce model error with respect to the mean and standard deviation of residuals. Calibration targets included stream flow measured at two gaging stations: Big Darby at Darbyville (3230500), and Little Darby at Jefferson (3230310). Depth-to-water table levels were measured at 12 monitoring wells. A split sample approach was used, designating measurements for the period 1992-1996 for calibration and 1997-2002 for verification.

The TOPNET component of the coupled model simulates runoff and outputs stream flow. Stream flow, based on USGS stream flow measurements at the two gaging stations in the watershed was uses as a calibration target. As mentioned in earlier sections, stream flow at the two gaging stations, Big Darby at Darbyville and Little Darby at Jefferson have been recorded since the 1950s. Daily measured and simulated stream flow for the calibration period is shown in Figures 5.20 and 5.21. A satisfactory match can be observed from these figures, except for consistent under simulation during high rainfall periods. This could be due to an underestimation of direct runoff from the watershed. The main emphasis during the calibration was to capture the overall trend for both stream flows and groundwater levels, and less attention was paid to a particular location in the watershed or time period.

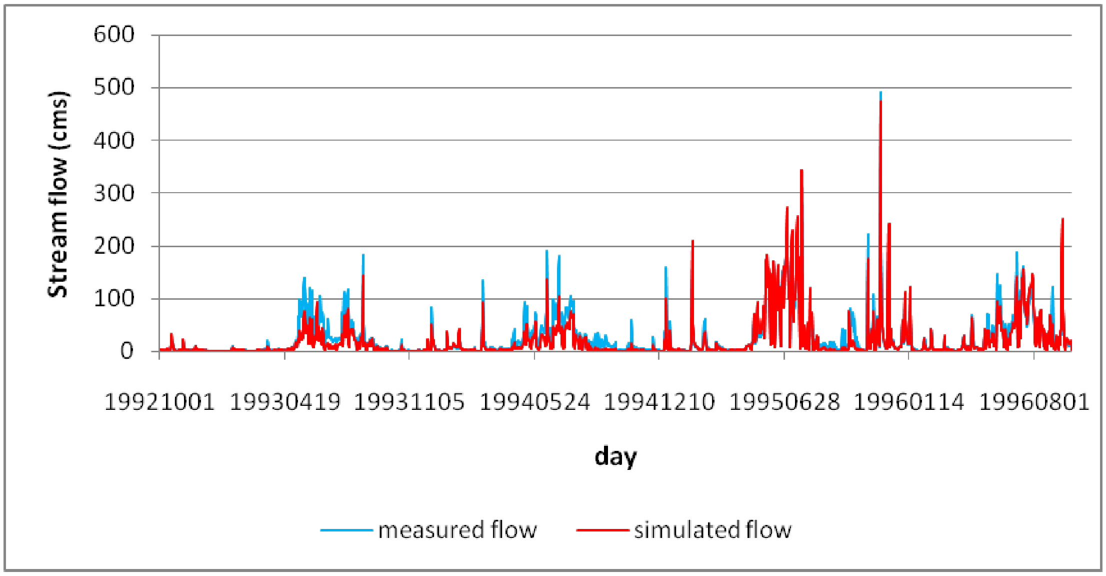


Figure 5.20. Stream flow at Darbyville during calibration period.

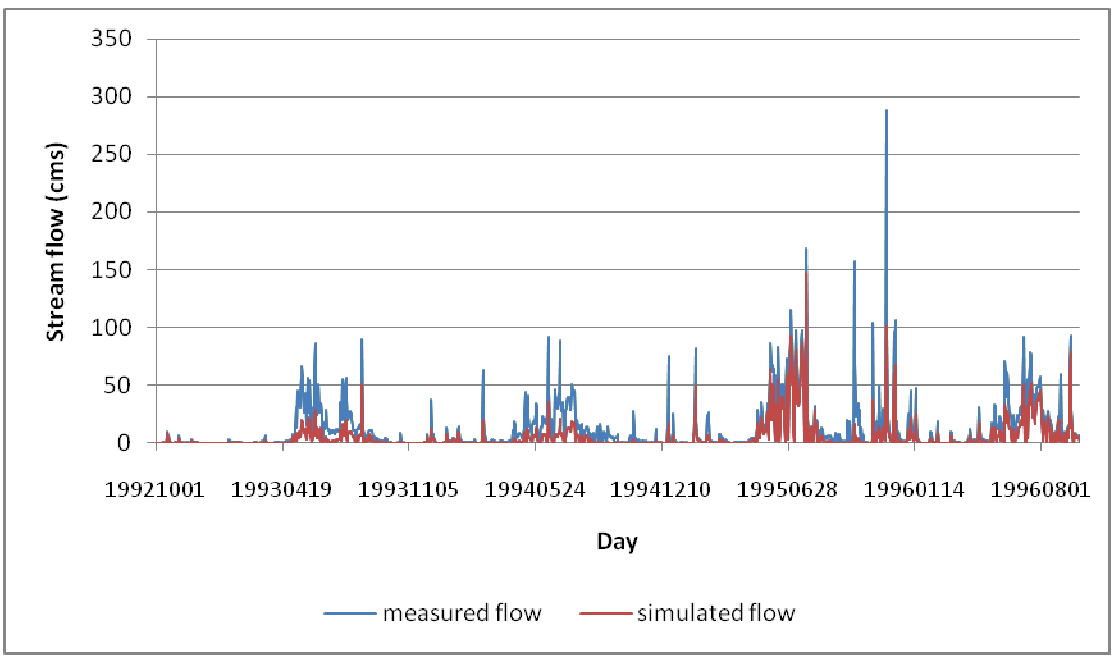


Figure 5.21. Stream flow at Little Darby at Jefferson during the calibration period.

Verification period

Model verification or corroboration is the process by which one demonstrates that the calibrated model is an adequate representation of the physical system. Because of the non-uniqueness in parameters obtained in model calibration, the set of parameter values used in calibration may not accurately represent field values. Model verification will help establish greater confidence in the calibration. The process of model verification involves comparing the model simulation results with observed data for a period other than the calibration period. This permits an independent assessment of model performance. The calibration period spanned 1992-1996 and the verification period spanned 1997-2002.

The daily stream flows for the verification period at the two gaging stations, Little Darby at Jefferson and Big Darby at Darbyville, are shown in Figures 5.22 and 5.23. A close agreement between the simulated and observed flows can be observed from these figures, except during the high rainfall periods where the coupled model underestimates stream flow. The major reason for model under simulation is the fact that the effect of snow is not simulated in this model. Some other reasons for underestimation of stream flow during these high rainfall periods are: effect of localized storm events which cannot be captured by the weather stations in the basins, probable contribution of flow from non-contributing areas, and uncertainty involved in flow measurement. However, the coupled model is capable of simulating the consistent overall trend for both the calibration and verification periods. Mean absolute errors of 4.3 and 8.5 (Table 5.7) for Darbyville and Jefferson, respectively, also show the fairly good simulation obtained using the coupled model. The results are not as good as those obtained using TOPNET alone, again for both the calibration and validation periods as shown by the statistics in Tables 5.7 and 5.8.

One of the reasons for the poor results using the coupled model compared to using TOPNET alone could be due to the simplifying assumptions used in the groundwater model. The conceptualization of the aquifer as a as single layer aquifer could have an impact on simulated heads, resulting in unrepresentative stream flows. The other assumption of the groundwater boundary coinciding with the watershed boundary may be another source of error since some of the water in the aquifer system is not accounted for, resulting in under simulation of stream flow.

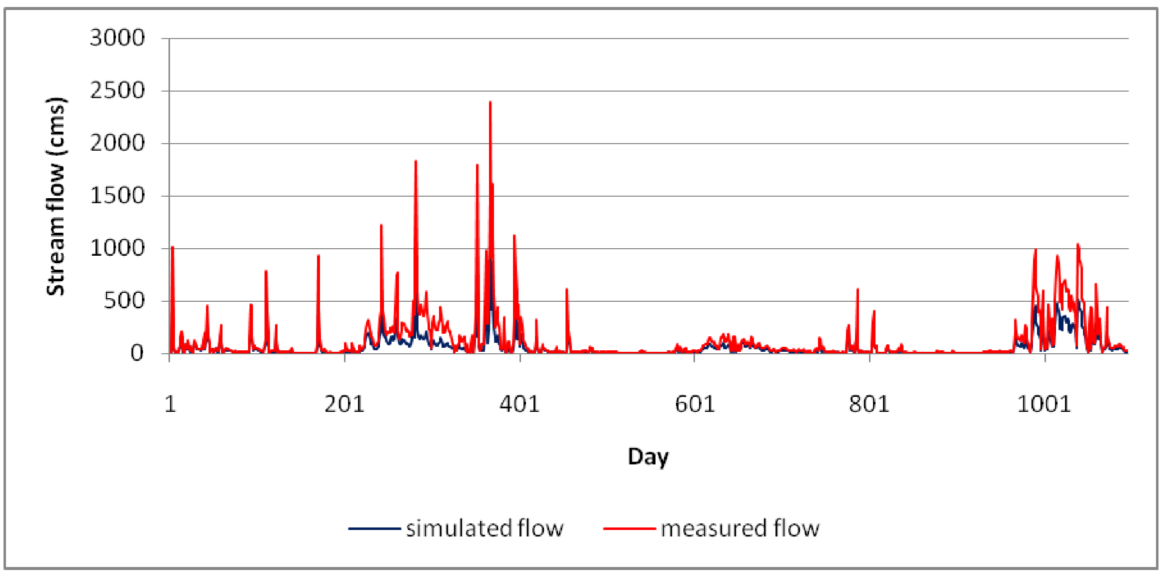


Figure 5.22. Stream flow at Darbyville during verification period (1997-2002).

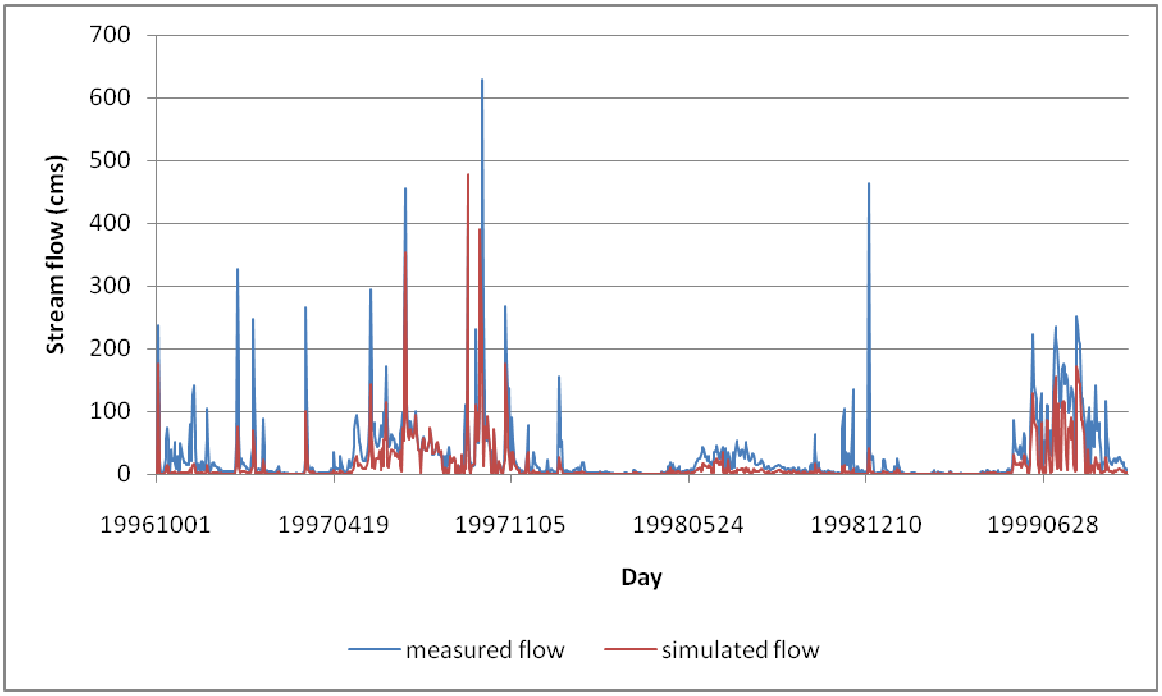


Figure 5.23. Stream flow at Little Darby at Jefferson during the verification period.

Table 5.8. Statistical evaluation of stream flow for model calibration period(1992-1996).

Statistic	Big Darby at Darbyville			Little Darby at Jefferson		
	Observed	TopN	TopN–ModF	Observed	TopN	TopN–ModF
Mean daily flow	12.6	12.4	7.6	6.3	5.9	3.8
Standard Deviation	7.9	7.6	10	12.9	7.6	6.1
Mean absolute error		2.5	6.6		1.8	4.5
Root mean square error		4.4	11.3		7.6	10.2

Note: All flows are in units of cubic meters per second (cms)

Table 5.9. Statistical evaluation for stream flow for model validation(1997-2002)

Statistic	Big Darby at Darbyville			Little Darby at Jefferson		
	Observed	TopN	TopN–ModF	Observed	TopN	TopN–ModF
Mean daily flow	10.3	9.1	6.2	4.5	5.1	2.9
Standard Deviation	8.9	7.6	5.2	8.3	6.9	6.4
Mean absolute error		1.1	4.3		2.4	8.5
Root mean square error		5.9	12.1		5.5	10.8

Table 5.10. TOPNET simulation model efficiency for calibration (1992-1996).

	Stream flow, cms			
Site	Mean Observed	Mean Simulated	Correlation Coefficient	Model fit efficiency
Big Darby at Darbyville	52.4	48.7	0.84	0.89
Little Darby at Jefferson	29.6	23.8	0.87	0.91

Table 5.11. Coupled TOPNET–MODFLOW simulation model efficiency for calibration period (1992-1996)

	Streamflow, cms			
Site	Mean Observed	Mean Simulated	Correlation Coefficient	Model fit efficiency
Big Darby at Darbyville	52.4	35.2	0.53	0.61
Little Darby at Jefferson	29.6	18.8	0.67	0.69

Results and discussion of Groundwater levels

There are twelve monitoring wells in the watershed. However only nine of these wells have measured ground water level data for long periods and these nine are used in model analysis. Figures 5.24 through 5.32 show a comparison of modeled groundwater heads and measured groundwater levels for nine monitoring wells in the watershed. The modeled heads were obtained using the coupled TOPNET–MODFLOW model. Generally, the coupled model was able to simulate the fluctuation range fairly well, but it does not represent the different peaks and their timing. The explanation for the shift in timing is not related to the groundwater model but in the recharge time series, which equals the bottom flux of the unsaturated zone model. The general trend of the groundwater dynamics is represented by the model, but the short-term, fast dynamics, is not. This could be a result of the 'smooth' and 'late' recharge time series due to the large time step used in MODFLOW. The unsaturated zone model retains and, thus, attenuates the groundwater flux too much, but represents the amplitude of the fluctuations reasonably well. The model is capable of describing the average groundwater level and the amplitude of its fluctuations in the Big Darby watershed.

The mean statistics of the residuals are shown in Table 5.12. The residuals are expected to be random with a mean close to zero and a constant variance. The residuals were examined to determine random behavior and it is clear that there is significant deviation from the expected. One of the major reasons for the deviations which are also clearly visible from visual examination of Figures 5.24 through 5.32, is probably the simplifying assumptions used in the groundwater model. The one layer aquifer model might not be a good representation of the groundwater physical system. The consolidated water budget for the watershed is shown in Table 5.13 using TOPNET, the coupled TOPNET–MODFLOW model, and that derived from baseflow separation techniques. As shown in Table 5.13, the basin wide average value of evapotranspiration is the major loss from the system. The basin wide average annual average values for precipitation, surface runoff and total groundwater recharge vary over a wide range of values during the entire simulation. For example, minimum precipitation is about 33mm and the maximum precipitation is about 1300mm. Similarly, minimum and maximum for groundwater recharge are about 0.5cm to 23cm respectively.

The surface runoff and recharge values follow the precipitation pattern: high during wet periods and low in dry years. It is clear from Figures 5.28 and 5.29 that surface runoff and groundwater recharge are critical hydrological components in the basin. This is especially true during dry years where the runoff and recharge are very low. Most of the stream flow in these low flow periods is a result of baseflow showing the interconnection between the surface and subsurface flow systems. The couple model results prove that the hydrological constituents in the Big Darby watershed are adequately represented for the entire simulation. Based on these results, the model can be used to predict stream flows and groundwater levels for long-term management scenarios.

Table 5.12. Statistics of groundwater level residuals.

Year	1992	1993	1994	1995	1996	1997	1998	1999
Mean	4.27	5.35	-1.12	6.18	8.8	10.2	4.75	6.9
Standard Error	0.67	0.75	0.66	0.59	0.87	0.71	0.53	0.53
Standard deviation	5.08	6.16	7.05	5.76	6.52	6.8	7.2	95.1

All values in units of meters

Table 5.13. Predicted hydrologic budget using Baseflow separation, TOPNET and the coupled TOPNET –MODFLOW.

Hydrologic component	TOPNET		TOPNET–MODFLOW		Baseflow separation	
	1992-96	1997-02	1992-96	1997-02	1992-96	1997-02
Precipitation	768	960	768	960	768	960
Surface runoff	57	104	37	63	59	134
Baseflow	10	47	18	34	12	67
Evapotranspiration	700	809	713	863	697	758

All values in mm

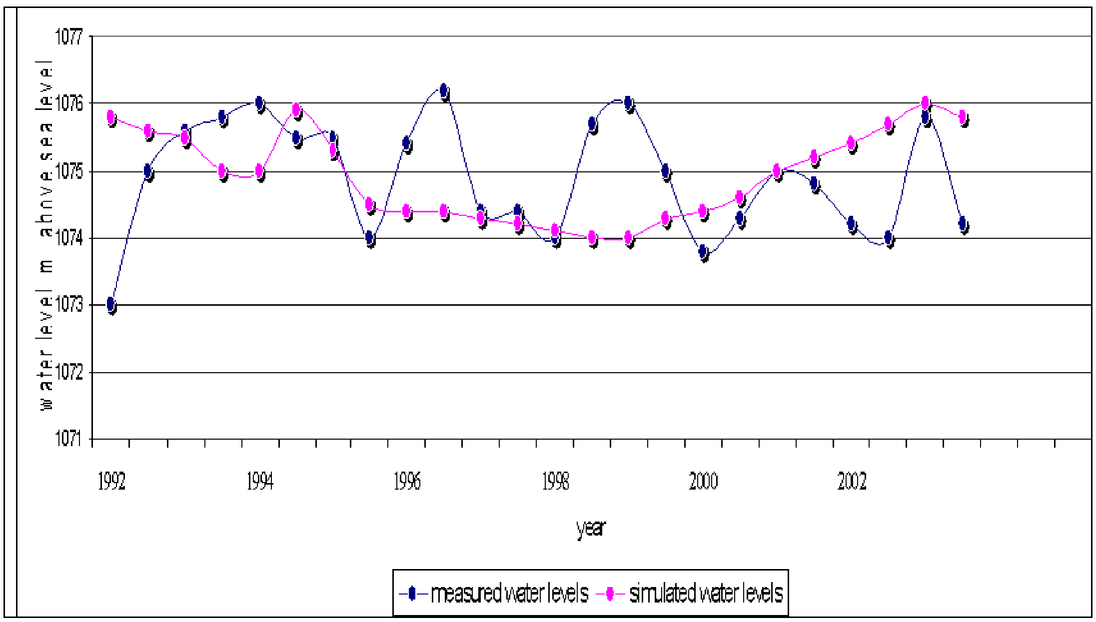


Figure 5.24. Measured and simulated groundwater heads at observation well PK-4.

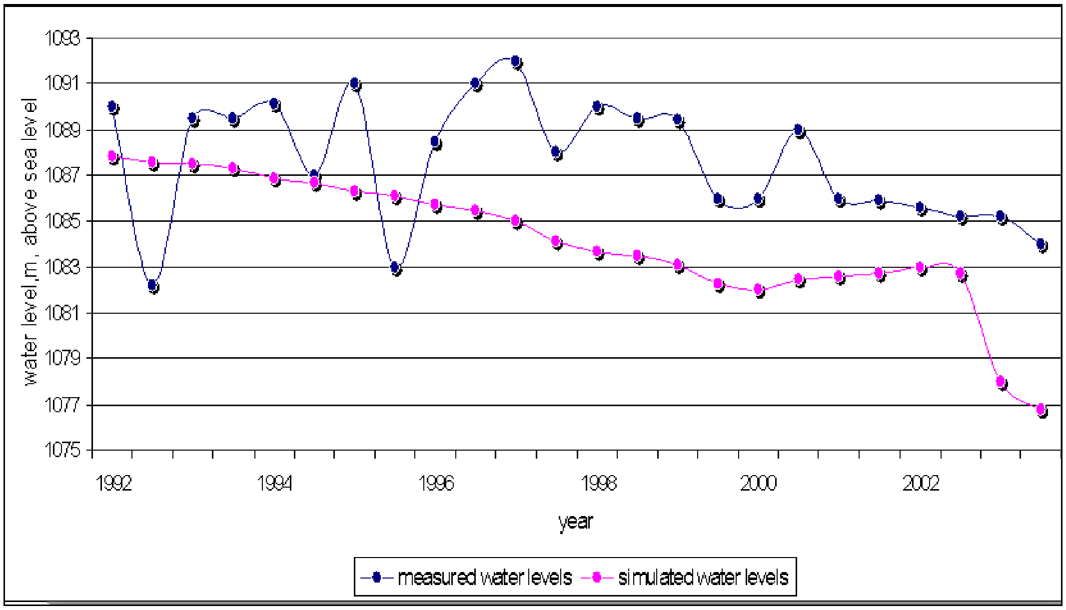


Figure 5.25. Measured and simulated groundwater heads at observation well PK-7.

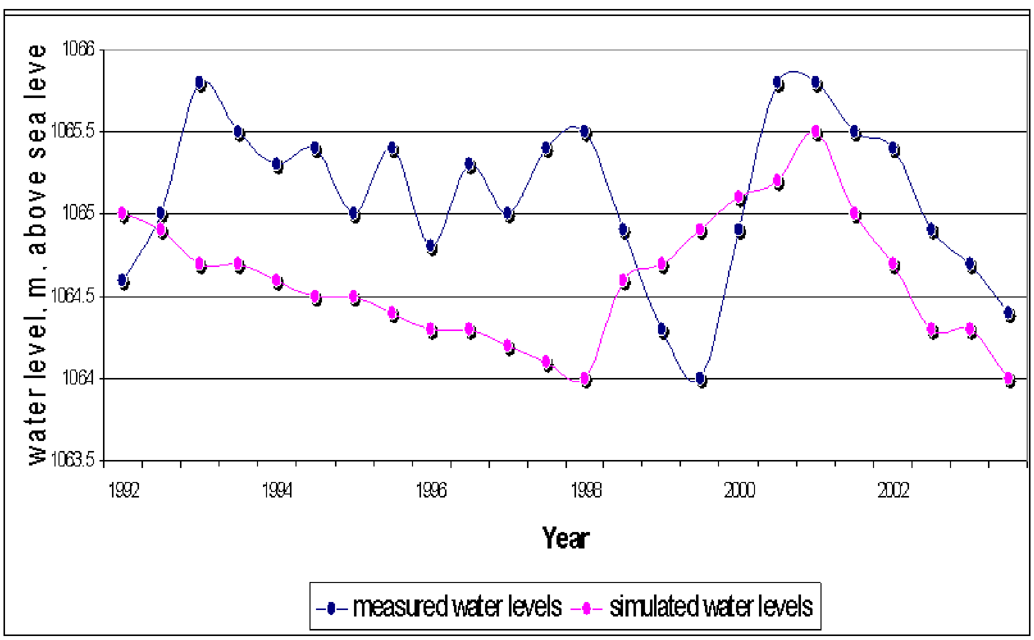


Figure 5.26. Measured and simulated groundwater heads at observation well M-2.

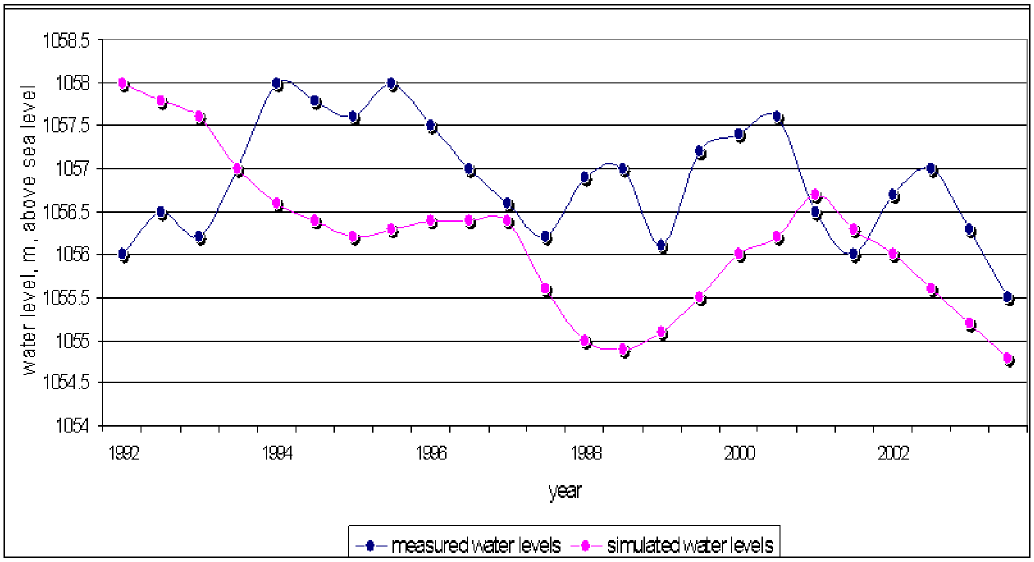


Figure 5.27. Measured and simulated groundwater heads at observation well M-4.

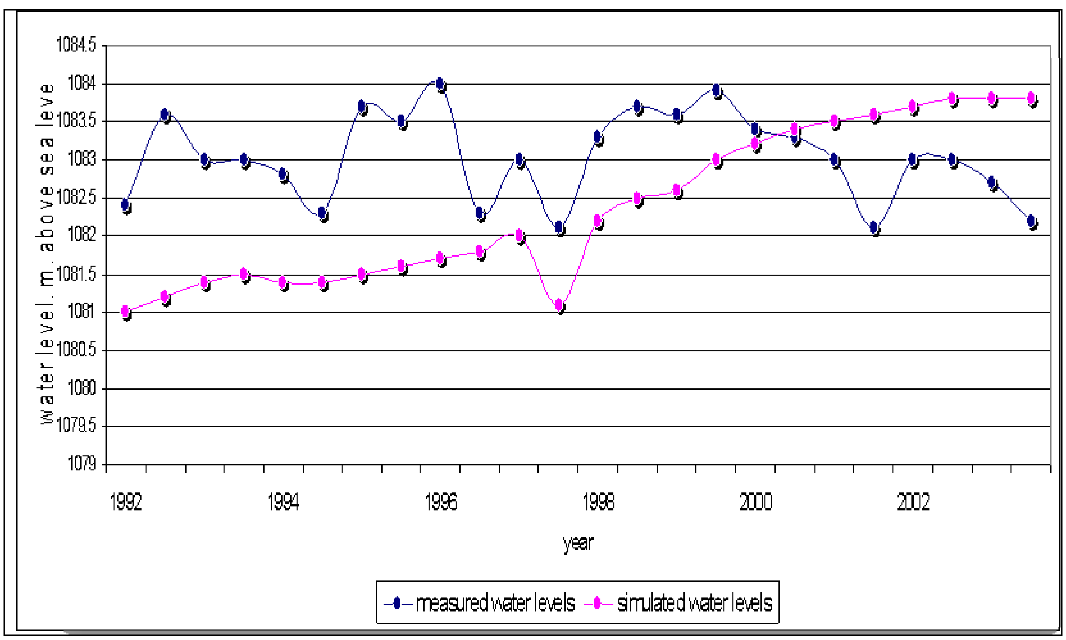


Figure 5.28. Measured and simulated groundwater heads at observation well FR-19.

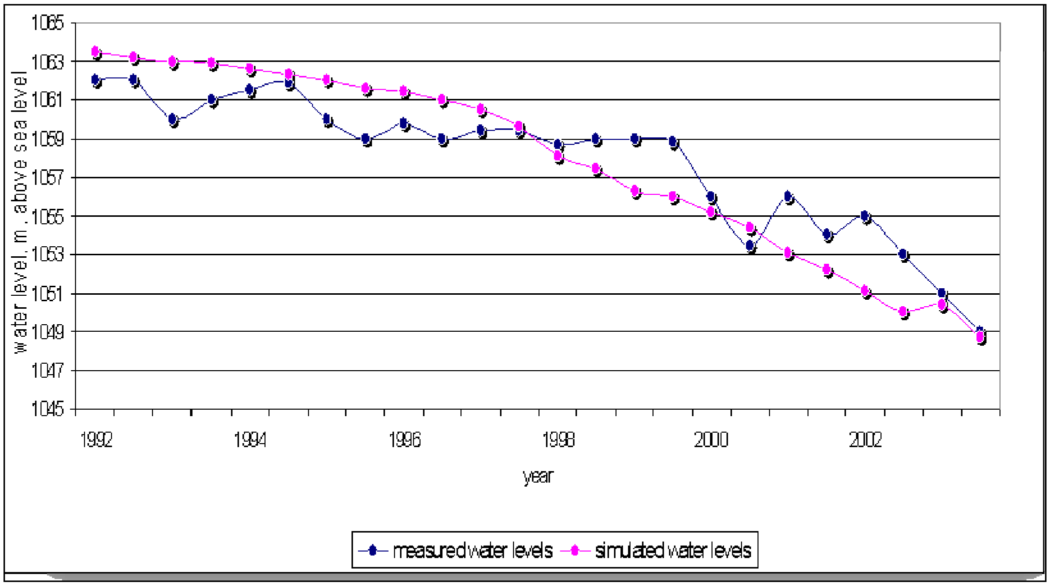


Figure 5.29. Measured and simulated groundwater heads at observation well FR-10.

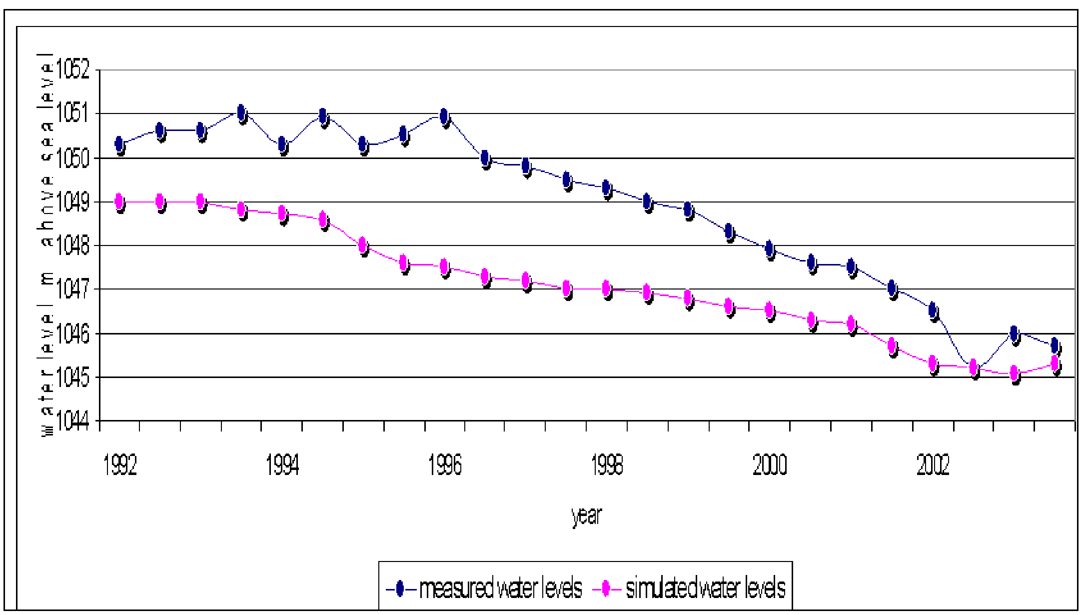


Figure 5.30. Measured and simulated groundwater heads at observation well U-4.

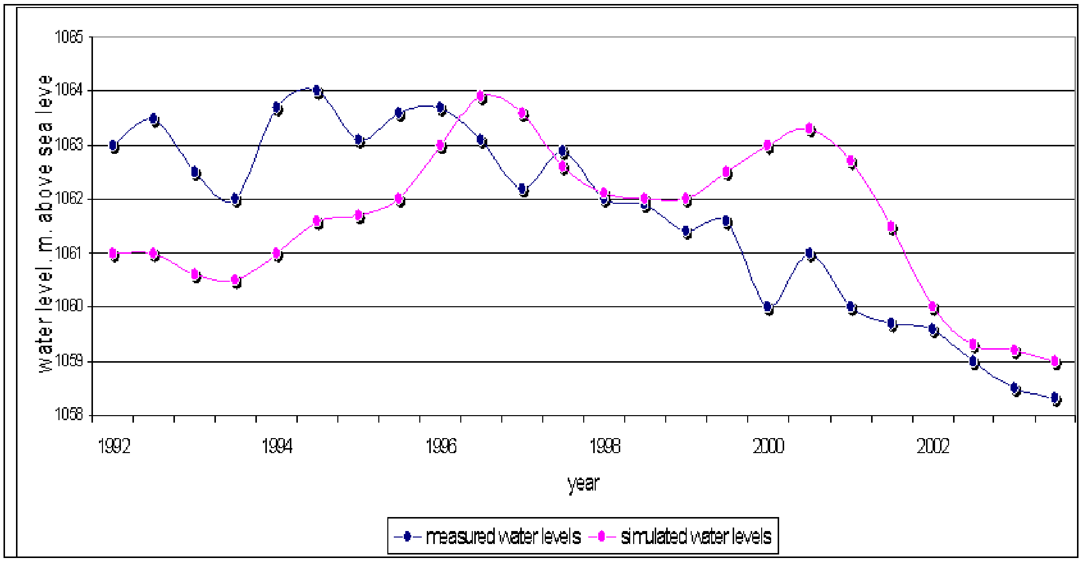


Figure 5.31. Measured and simulated groundwater heads at observation well LO-3.

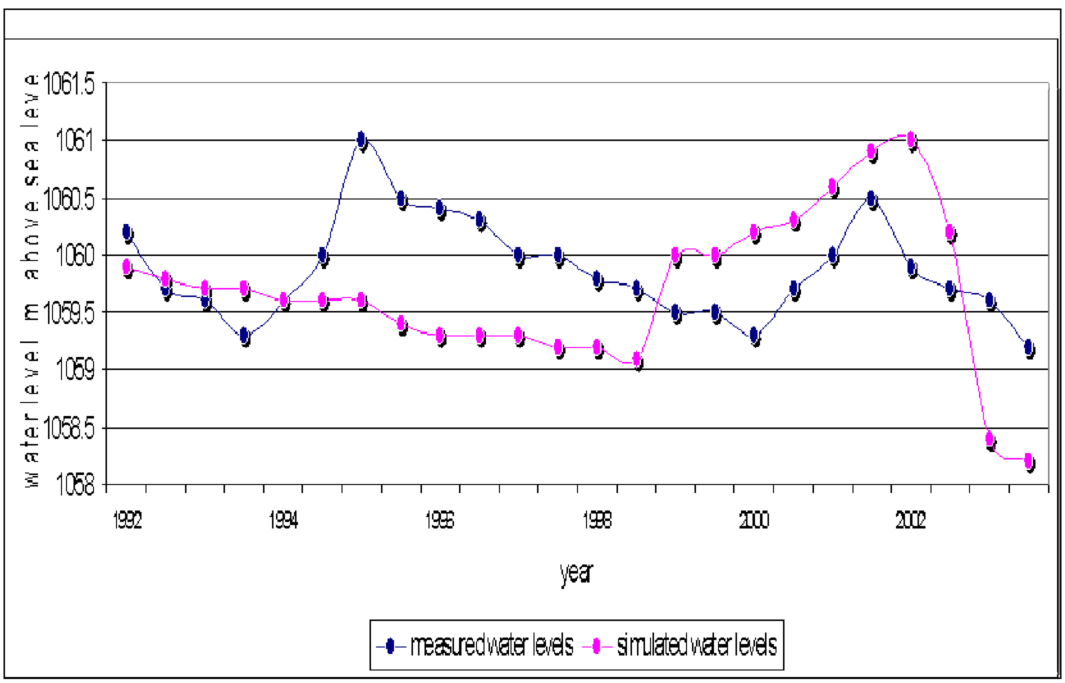


Figure 5.32. Measured and simulated groundwater heads at observation well PK-9.

<u>Effects of Water Withdrawals on Simulated Stream Flow</u>

The coupled model for the Big Darby basin was used as a tool to evaluate the effects of different ground water withdrawal scenarios on stream flow. The model was applied only to the Little Darby creek watershed with stream flow measured at the West Jefferson gaging station.

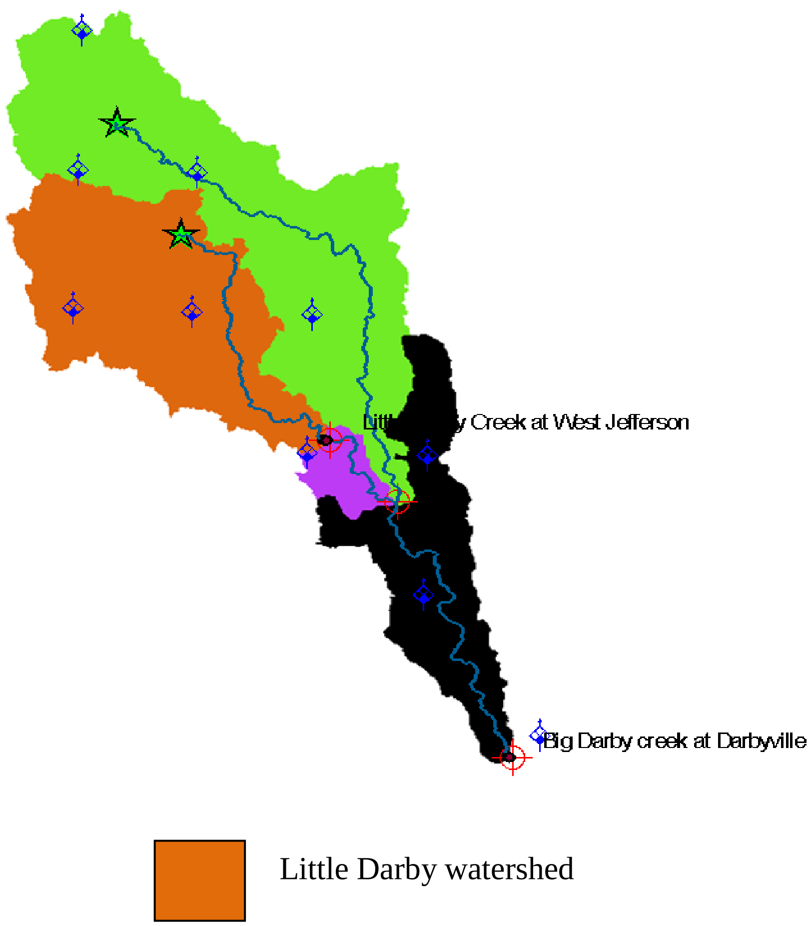


Figure 5.33. Little Darby watershed.

Results of the simulation of these hypothetical scenarios will help show how the coupled model can be used as a water resources management tool. Five hypothetical scenarios were identified and these are:

1. Stopping all withdrawals during the 1992–94 period,
2. Only ground-water withdrawals for the 1992-1994 period,
3. Only surface-water withdrawals for the 1992-1994 period,
4. Simulate stream flows for 1994-1998 with 1994 land-use conditions, and no withdrawals,
5. Simulate stream flows in response to average 1994–98 water withdrawals.

The 1992-1994 period was chosen since it was a period with low stream flows indicating below normal rainfall while the 1994-1998 period was chosen because it was a period with average to above average rainfall. The variation from the 1991 land use conditions used for the first three scenarios to the 1994 land use conditions was chosen to understand whether the coupled model can capture the influence of land use changes on hydrological balances of a watershed. The scenarios required modification of the land use input files in TOPNET and the water withdrawal input file in MODFLOW. Water withdrawals were estimated for periods of missing data for each reach by averaging daily withdrawals for each month that data were available and generating a similar daily withdrawal for each month for periods of no record. The simulation results obtained for each of the defined scenarios provide relative differences between stream flows under different water withdrawals and land uses.

1992-1994 simulation period

Flow-duration curves for simulations of the 1992-1994 period are shown in Figure 5.34 for simulations with (1) no withdrawals, (2) only surface-water withdrawals, and (3) only groundwater withdrawals.

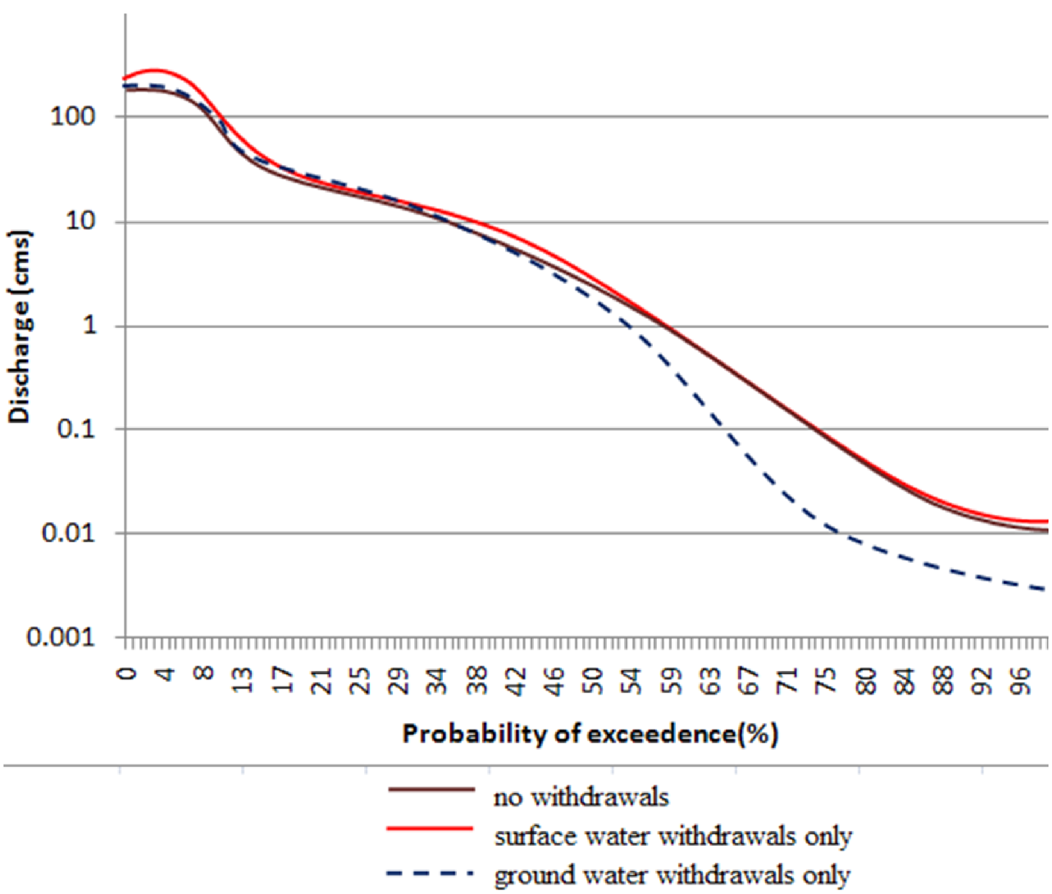


Figure 5.34. Flow-duration curves developed from simulated daily flows for three scenarios- no withdrawals, only surface-water withdrawals, and only ground-water withdrawals at the West Jefferson station, 1992-1994.

The flow-duration curves are almost similar at the West Jefferson station for simulations with no withdrawals and simulations with only surface water withdrawals at the gaging station. The two curves show a marked differences at the 99% exceedence probability. This indicates that surface water withdrawals have little effect on the duration and frequency of low flows. This also shows that groundwater withdrawals have a marked influences on the magnitude, duration, and frequency of low flows. Generally in the Big Darby watershed, more than 95% of water is from ground water sources. That is why any surface water withdrawals have a little impact as shown on Figure 5.34.

The differences between curves diminish as the exceedence probability decreases; little difference between curves is indicated below the 50 percent exceedence probability. This indicates that water withdrawals have little effect on high and medium flows at either station. As mentioned earlier on, in the entire Big Darby watershed all counties depend mostly on ground water for domestic, industrial and agricultural purposes. There is very little surface water use.

1994-1998 simulations

The 1994-1998 simulations shown in Figure 5.35 indicate that the differences in stream flow between scenarios with no withdrawals and those with average water withdrawals are similar to the differences in stream flow for the 1992–94 simulations for similar types of scenarios. The flow-duration curve for 1991 land use with no withdrawals is similar to the flow-duration curve with no withdrawals with 1994 land-use conditions. Small differences can be noted in the flow-duration curves between simulations with 1991 land use with no withdrawals and simulations with no withdrawals with 1994 land-use conditions for medium- to high-flow conditions at the gaging station. The lack of difference in duration curves might reflect greater lower zone evapotranspiration between forested and open land which offsets any gains in base flow because there is less developed land use.

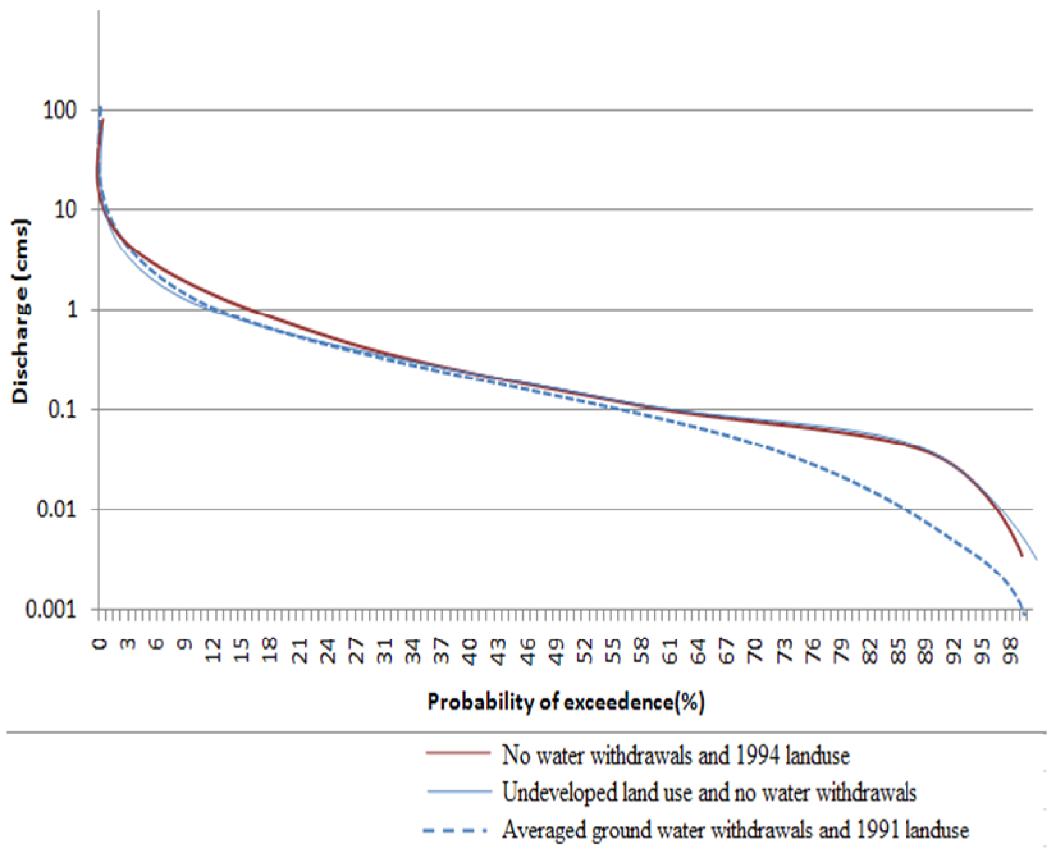


Figure 5.35. Flow-duration curves for average water withdrawals with 1994 land-use conditions, no withdrawals with 1994 land-use conditions, and no withdrawals with undeveloped land-use conditions, for the Little Darby creek at West Jefferson for the period (1994-98).

Minimum daily flows for simulations with no withdrawals with 1994 land-use conditions and no withdrawals with undeveloped land-use conditions were comparable. At West Jefferson station, flows ranged from 5.8 and 5.5 m^3/s at the 100-year recurrence interval to 23 m^3/s and 21 m^3/s at about the 1-year recurrence interval for simulations with (1) no withdrawals with 1991 land-use conditions and (2) no withdrawals with undeveloped land-use conditions, respectively. Simulations with no withdrawals with 1994 land-use conditions and those with 1991 land use conditions indicate that 1991 land use conditions resulted in decreased discharge at the gaging station.

This may mean that the 1991 land use condition resulted in lower flow through evapotranspiration. Less impervious surfaces in 1991 led to less runoff and stream flow compared to 1994. Minimum daily flows for simulations with average 1992–94 withdrawals were considerably less than the minimum daily flows for simulations with no withdrawals. Flows with water withdrawals ranged from 0.84 m^3/s to 13 m^3/s.

Potential Uses of the Model

Having developed the coupled model and applied it to the two case studies, the logical question is to explain the purpose of the coupled model. How does it help improve or what is the contribution to the science of watershed modeling?

In order to understand and decide whether there is need for a coupled surface water–groundwater model, one needs to first define the problem one wishes to address. If the goal is simply to model stream flow at a particular location in a watershed, one may not need to use an integrated model. Also, if the goal is simply to model heads in an aquifer, then a simple lumped parameter model of groundwater flow might suffice.

However, if one is interested in evaluating the effects of groundwater or surface water withdrawals in relation to climate variability on the entire resource of a watershed or basin, then the only means of reasonably doing such an analysis is with integrated models. Thus, the model developed in this research becomes important.

A TOPMODEL model works well for evaluating the total flow out of the Tenmile Creek watershed in Washington State. There are few observation wells and the rocks are, at best, marginally permeable. Yet, the TOPMODEL simulation is incapable of simulating the decadal changes in baseflow (fall flows) caused by climate variability because TOPMODEL cannot simulate the much longer transient times of deeper groundwater flow, which only can be simulated when we add multiple layers to MODFLOW in an integrated model that connects TOPMODEL with MODFLOW. Furthermore, there is no way to evaluate transient effects of groundwater withdrawals on stream low and soil zone storage when using TOPMODEL and MODFLOW. The effects of groundwater withdrawals on stream flow can only be determined by ignoring the timing of snowmelt, ET, and soil-zone storage. Thus, many assumptions are made when effects of groundwater are simplified in TOPMODEL and effects of precipitation, ET, and runoff are simplified in MODFLOW. However, if one wants to evaluate complex problems associated with changing climate, land use, and withdrawals on the combined resource (both surface and ground waters), then an integrated model will provide the best answers.

The reason for developing the coupled model and the capability of simulating unsaturated flow from land surface to the water table is that groundwater models neglected surface effects on the flow and storage above the water table and surface water models neglected complexities in groundwater flow. That is, recharge in most MODFLOW simulations was assumed at some constant rate approximated externally without consideration of whether the aquifer was completely saturated (saturation excess or rejected recharge). Similarly, most surface water models, including TOPMODEL, ignored many of the processes including flow and storage in unsaturated zones, and flow and storage in multiple aquifers that affected the rate and timing of baseflow to streams and interactions with lakes, and certainly no surface water model is capable of simulating the effects of groundwater withdrawals, other than with extremely simplistic calculations.

The developed coupled model also has varying recharge rates in each time step and for each sub basin. It is highly likely that flow through the unsaturated zone results in non-uniformly distributed recharge which changes in space and time simply because of the thickness of the unsaturated zone and the quantity of water that can be stored in it. Such a simulation results in a completely different conceptualization of groundwater interaction with surface water and in the timing of stream flows, lake stages, and the formation of intermittent wetlands. Superimposing groundwater withdrawals on such a simulation results in a different effect than would be predicted if recharge were assumed to vary uniformly across the modeled area, including areas where the head in the shallow aquifer was above land surface.

It is also important to mention why the couple model is a better tool than, for example, just using MODFLOW to simulate surface and subsurface dynamics, such as using the evapotranspiration package to simulate evapotranspiration. MODFLOW only simulates ET (evapotranspiration) losses from groundwater. This loss is computed on the basis of a linear function (or a series of segmented linear functions) between the base of the rooting depth and, typically, land surface. MODFLOW does not compute a potential evapotranspiration rate on the basis of energy and, thus, any computations on potential evapotranspiration must be done externally to the simulation, so it is impossible to evaluate temperature changes on ET in MODFLOW without first computing it externally. TOPNET gives a much more accurate simulation of evapotranspiration. Prior to the development of the Unsaturated Zone Flow Package (UZF), soil moisture was simply ignored in MODFLOW.

Up until now, with the development of Coupled Ground-Water and Surface-Water Flow (GSFLOW) model which couples MODFLOW and Precipitation-Runoff Modeling System (PRMS), the simulation of surface water processes with MODFLOW was limited to saturated interactions with stream channels and lakes, and lakes could only fill because of groundwater discharge into the lake (a major limitation of the Lake Package). One could not simulate overland runoff when precipitation exceeded the hydraulic conductivity of the soil nor was there any means of simulating overland runoff from saturation excess when the water table rose to or above land surface.

Although springs, wetlands, and riparian areas could be simulated (springs could be simulated as drains or as stream reaches, and wetlands and riparian areas could be simulated as specified fluxes, as ET, or drains), they had to be specified prior to the simulation and could only change if specified in later stress periods. There was no mechanism that allowed a spring, wetland, and riparian area to form and disappear simply on the basis of the water table in relation to land surface without specifying its location in the model. This is clearly defined in TOPNET and it makes the coupling of TOPNET and MODFLOW a worthwhile scientific venture.

In summary the developed coupled model can be a useful tool in addressing the following hydrological questions:

1. What are the effects of a rising/falling water table on surface water processes?
2. What are the dynamics of surface water and groundwater interaction in springs, wetlands, and riparian areas?
3. What are the effects of different climate scenarios (e.g., floods and droughts) on a surface water and groundwater system?
4. What are the effects of different management scenarios (e.g., conjunctive use, urbanization, and irrigation) on the surface and groundwater system?

These are fundamental problems that are key to sustainable water resources management. Despite the modeling results being quite close to those obtained using TOPNET alone on stream flow simulation, for example, the exercise is a worthwhile attempt for reasons outlined earlier on in this section.

Addressing the ever-increasing range and complexity of environmental resource management and policy development requires interdisciplinary and adaptive approaches that build on existing science and technology. Also, these approaches must provide mechanisms for modeling over different spatial and temporal scales, and provide for the integration of science and management objectives. These are the objectives of the developed coupled model.

CHAPTER 6

SUMMARY, CONCLUSIONS AND RECOMMENDATIONS

Summary and Conclusions

This dissertation has outlined the development and use of a model coupling tool which was used to successfully integrate two separate hydrological models, TOPMODEL and MODFLOW. Past research has shown that major challenges to model coupling are: 1) disparate time scales of surface water behavior and groundwater behavior which are difficult to resolve, 2) the spatial discretization schemes for surface and sub surface models are dissimilar. The InCouple methodology used an approach where interfaces were utilized to enable the models to communicate with each other as they run. The communication occurs by way of a third program that acts as a coupler. Prior to performing coupled simulations with the integrated model, the surface water and groundwater models were independently developed and calibrated to the extent possible.

Interaction between the two models goes beyond a simple feed-forward approach in which one model is run to completion and then its results are directed into another model as input. Data from the models are transmitted at the completion of each time increment. In this way, the results of each model can immediately influence the functioning of the other. Synchronization of the models is maintained by the interprocess communication functions.

Differences in the discretization of time in the two models are handled by integrating the shorter time steps of the TOPMODEL model to correspond to

MODFLOW's longer stress periods. Differences in the spatial discretization schemes of the models are accommodated by performing spatial conversions at each time step.

TOPMODEL simulates discharge and daily groundwater recharge at a high spatial resolution. Using the latter as input, MODFLOW calculates groundwater levels and baseflow, which are then returned to TOPMODEL. The basic approach of having an intermediary program interposed between the models being coupled allows pre-existing modeling systems to be linked without major restructuring. Each model operates and communicates only with reference to its own spatial and temporal domain, and the coupler performs all tasks related to reconciling the different spatial segmentation schemes and time scales. Models representing other related processes, (e.g., a receiving water quality or a model of unsaturated zone transport), could be readily incorporated into the system as well. Then, since each model in the system is aware only of the coupler and contains no functions specific to any other model, different combinations of models could easily be applied to a particular application as the project's needs dictate.

The obtained results show the inadequacies of the coupled models, and their imitations. The developed coupled model for the Big Darby watershed provides a watershed scale simulation of surface water and groundwater flow in the study area. As with all mathematical models of natural systems, the simplifications and assumptions incorporated into the models result in limitations to their appropriate uses and to the interpretations that may be made of simulation results.

The groundwater model component simulates groundwater flow, water levels, and the interaction with surface water features at the watershed scale. Hydrologic processes and spatial variability in hydraulic properties and stresses are simplified and approximated to a degree consistent with this scale. The model calibration also represents the best fit to estimates and observations made throughout the watershed. Thus, the agreement between simulated water levels or stream baseflow in specific areas of the flow system may not be adequate to support local-scale model applications for smaller subbasins.

The effects of temporal and spatial discretization also impose limitations on model use. Hydrologic processes and hydraulic stresses were represented in the transient models as monthly averages. Simulation results are monthly average groundwater levels and flows. The surface water model, however, uses a daily time step. Despite attempts to synchronize the time steps, this spatial disparity may also have a great influence on the modeled results. The spatial resolution of the simulation results was limited by the area of the 500m x 500m grid cell.

In both case studies, the groundwater models were discretized as single layer aquifers; therefore, groundwater flow through the other aquifer layers is not directly simulated in the models. Groundwater in fractured bedrock can have a widely variable area of recharge and natural discharge. Thus, even though water withdrawals from bedrock aquifers may be simulated indirectly, the effects of these withdrawals on the groundwater–surface water system may not be appropriately addressed with such models. In the absence of any information about flow rates or pathways through bedrock, where flow through bedrock is substantial, the spatial or temporal distribution of simulated flows may not be adequately represented.

Another model limitation exists due to differences in the temporal discretization of the coupled models. A time step of one day was implemented in TOPMODEL and

TOPNET, while a monthly time step was most feasible in MODFLOW. This was due to the differences in the flow rates in the surface and subsurface systems. In the surface water systems, a substantial amount of flow can occur from one location to another during a day while, for the groundwater system, the change in flow rates is negligible over a period of one day. The model is, thus, able to adequately represent stream aquifer interactions on a long term basis, and does not represent the day to day variations of this phenomenon

In spite of adequate data in the Big Darby watershed, there are still significant gaps which must be bridged by interpolation or assumptions based in part on educated guesses. The need to interpolate data is not exceptional in groundwater modeling, but in a regional model, the results of interpolation are even less reliable because of the large distances between observation points. Often the most basic requirements needed to interpolate meaningfully between two data points are not fulfilled. On the other hand, for some parameters, interpolation is the only way to parameterize the large number of cells.

Recommendations

Fundamentally, the study was conducted to evaluate the potential of using coupling interfaces for model integration and to evaluate the extent of surface water– groundwater interactions in the case study watersheds.

Integration of the models provides more than just an improvement in estimates of stream flow at the basin outlet. In the Big Darby case study, the coupled model takes advantage of TOPNET which can be calibrated to stream flow (and other data) measured within the basin to improve the spatial distribution of flows and storages in addition to the overall stream flow at the basin outlet. Also, spatial variations in surface-water groundwater interaction are simulated such that management issues can be analyzed with the coupled model, such as conjunctive water use, stream depletion, etc. The main expectation is for the coupled model to be used for applications beyond predicting the stream flow at the outlet of a basin, which can be achieved reasonably well using an independent watershed runoff model.

A general recommendation is that future studies should consider a variety of improvements to the existing models, including additional data collection, different conceptual model design, and other factors. This will encompass the following specific aspects for further exploration and evaluation to generate greater confidence in the developed coupled modeling system:

1. A post-audit comparison of the predictive abilities of the TOPMODEL–MODFLOW coupling with those of other coupled models like MIKE-SHE to help assess their relative merits.
2. Rigorously calibrate and refine the groundwater and surface water modeling components using both models in conjunction.
3. Run the models based on specified management and develop output data sets for analysis scenarios.
4. It is also important to ensure that the rainfall runoff model component considers changes in land use over time. Therefore, an additional modeling component that will cater to land use changes is highly recommended so that the rainfall runoff model succinctly captures and accurately describes runoff and stream flow generation.

5. A more accurate assessment of the coupled model performance can be attained by testing the model with the inclusion of TOPNET water management option.
6. Synoptic measurements of groundwater levels in wells would improve knowledge of the water table configuration and provide better groundwater level data for model adjustments. Data from the twelve wells used in this study may be insufficient. Therefore, there is need to establish more monitoring wells. Continuous monitoring of water levels in wells near streams can provide a record of the transient response to natural and anthropogenic events.
7. Further investigations of the impacts of the surface and groundwater interactions on soil moisture, and surface water and energy fluxes over larger areas under both wet and dry climate conditions are necessary.

While the main reason for developing the coupled model was to build a tool to quantify and predict spatial and temporal variability of interdependent surface and subsurface hydrologic fluxes of precipitation, evapotranspiration, runoff, infiltration, recharge, and discharge, the structures of the component models used allow for integration with other scientific disciplines and environmental processes. Thus, it is recommended that more work can be done to expand the model to include geochemical and water quality components.

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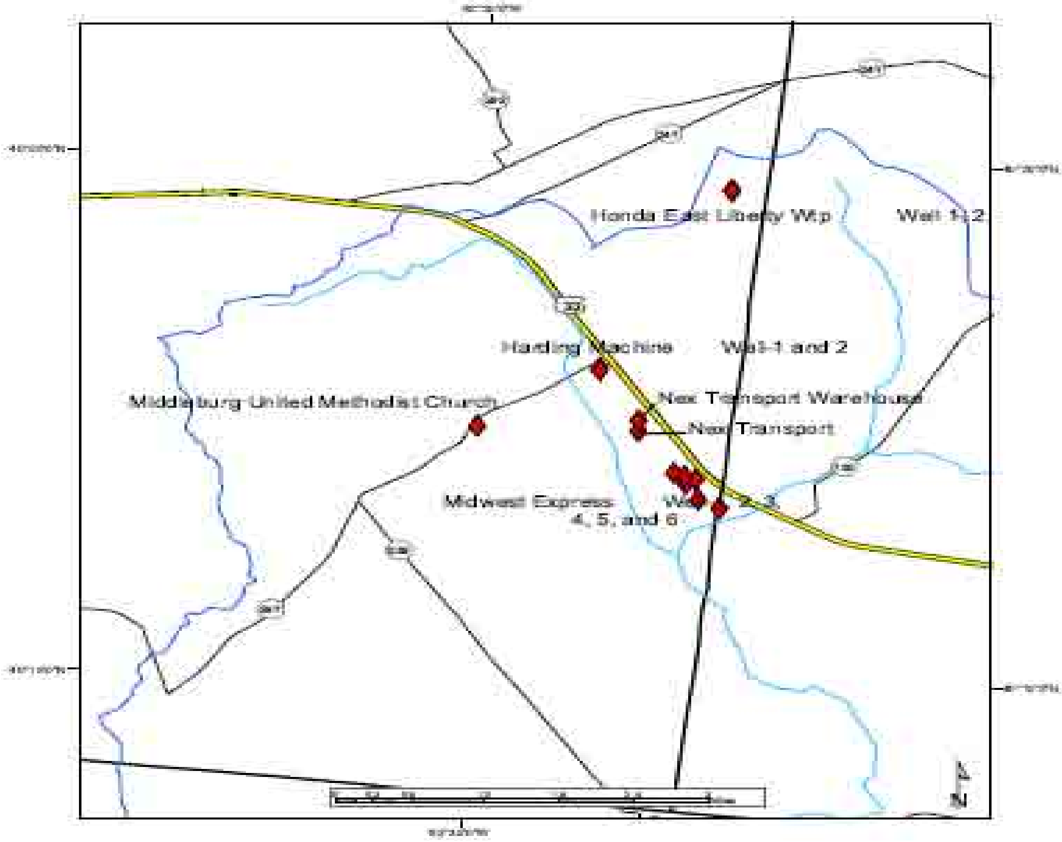
APPENDICES

Appendix A: Data Mapping Script

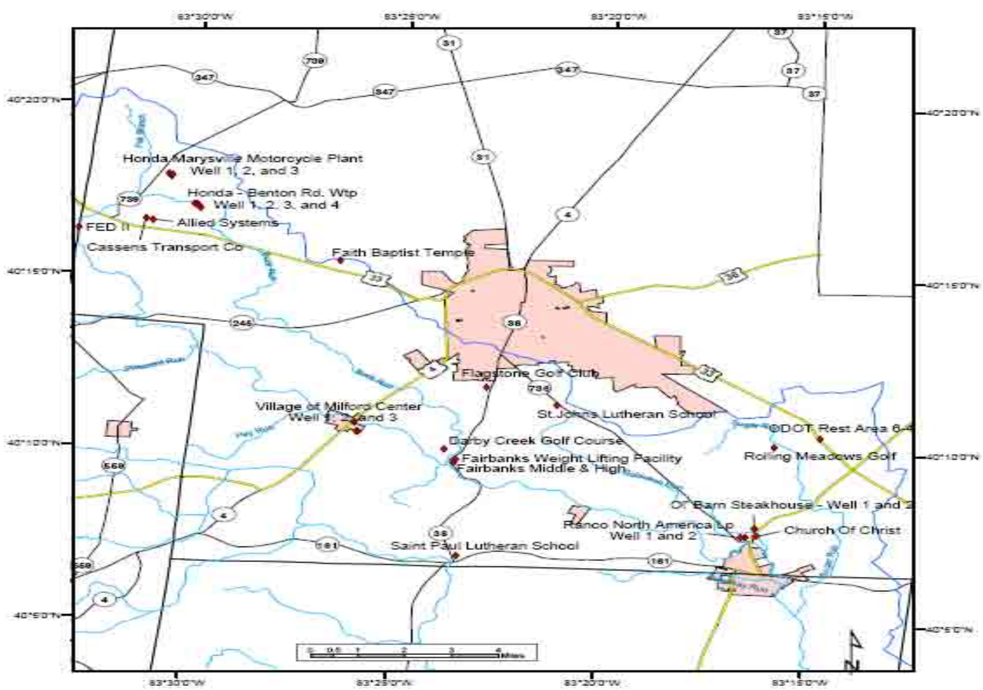
```
Public Sub createDataMapping()
Dim pMxDoc As IMxDocument, pMap As IMap, pFilter As IQueryFilter
Dim pIntersectLayer As IFeatureLayer, pOverlayLayer As IFeatureLayer
Dim pOverlayFCursor As IFeatureCursor, pIntersectFCursor As IFeatureCursor
Dim pIntersectTopo As ITopologicalOperator As Double, theProportion As Double
Dim pIntersectFeature As IFeature, pOverlayFeature As IFeature
Dim pSpatialFilter As ISpatialFilter, pOverlayArea As IArea, newArea As IArea
Dim pIntersectFClass,pOverlayFClass As IFeatureClass
Set pMxDoc = Application.Document
Set pMap = pMxDoc.FocusMap
Set pActiveView = pMap
If Not TypeOf pMap.Layer(0) Is IFeatureLayer Then Exit Sub
If Not TypeOf pMap.Layer(1) Is IFeatureLayer Then Exit Sub
Set pIntersectLayer = pMap.Layer(0)
Set pOverlayLayer = pMap.Layer (1)
Open "data_map.txt" For Output As #1
headerLine = Format (Now, "mm/dd/yyyy")
Print #1, "Created by ArcMap on " & headerLine
Set pIntersectFClass = pIntersectLayer.FeatureClass
Set pOverlayFClass = pOverlayLayer.FeatureClass
Set pFilter = New QueryFilter
pFilter.WhereClause = ""
Set pIntersectFCursor = pIntersectLayer.Search(pFilter, False)
Set pIntersectFeature = pIntersectFCursor.NextFeature
While Not pIntersectFeature Is Nothing
Set pIntersectTopo = pIntersectFeature.Shape
Set pSpatialFilter = New SpatialFilter
pSpatialFilter.GeometryField = pIntersectFClass.shapeFieldName
Set pSpatialFilter.Geometry = pIntersectFeature.Shape
pSpatialFilter.SpatialRel = esriSpatialRelIntersects
Set pOverlayFCursor = pOverlayFClass.Search(pSpatialFilter, False)
Set pOverlayFeature = pOverlayFCursor.NextFeature
While Not pOverlayFeature Is Nothing
Set pOverlayArea = pOverlayFeature.Shape
Set newGeometry = pIntersectTopo.Intersect(pOverlayFeature.Shape,
pIntersectFeature.Shape.Dimension)
Set newArea = newGeometry
theProportion = newArea.Area / pOverlayArea.Area
iModel = pIntersectFeature.Value(pIntersectFeature.Fields.FindField("Model"))
iCoupleID =
pIntersectFeature.Value(pIntersectFeature.Fields.FindField("CoupleID"))
oModel = pOverlayFeature.Value(pOverlayFeature.Fields.FindField("Model"))
```

```
oCoupleID = pOverlayFeature.Value(pOverlayFeature.Fields.FindField("CoupleID"))
Print #1, iModel & ":" & iCoupleID & " " & oModel & ":" & oCoupleID & " " &
theProportion
Set pOverlayFeature = pOverlayFCursor.NextFeature
Wend
Set pIntersectFeature = pIntersectFCursor.NextFeature
Wend
Close #1
End Sub
```

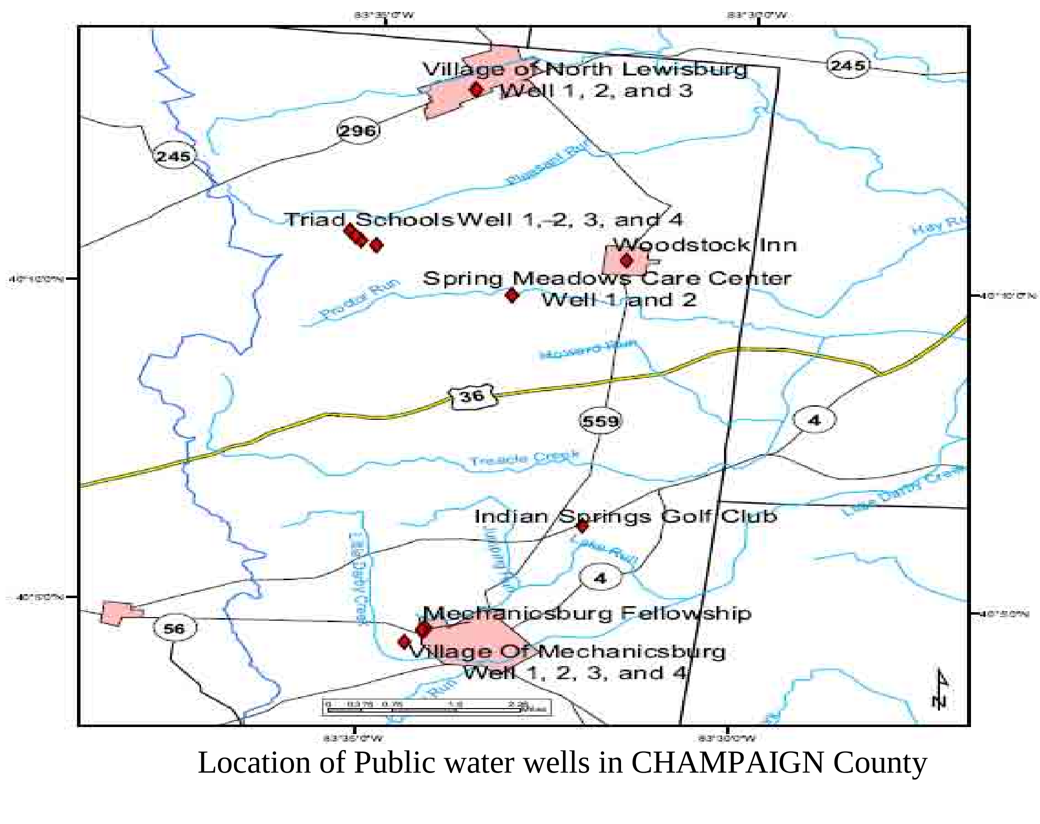
Appendix B: Location of Public Wells in the Big Darby Watershed



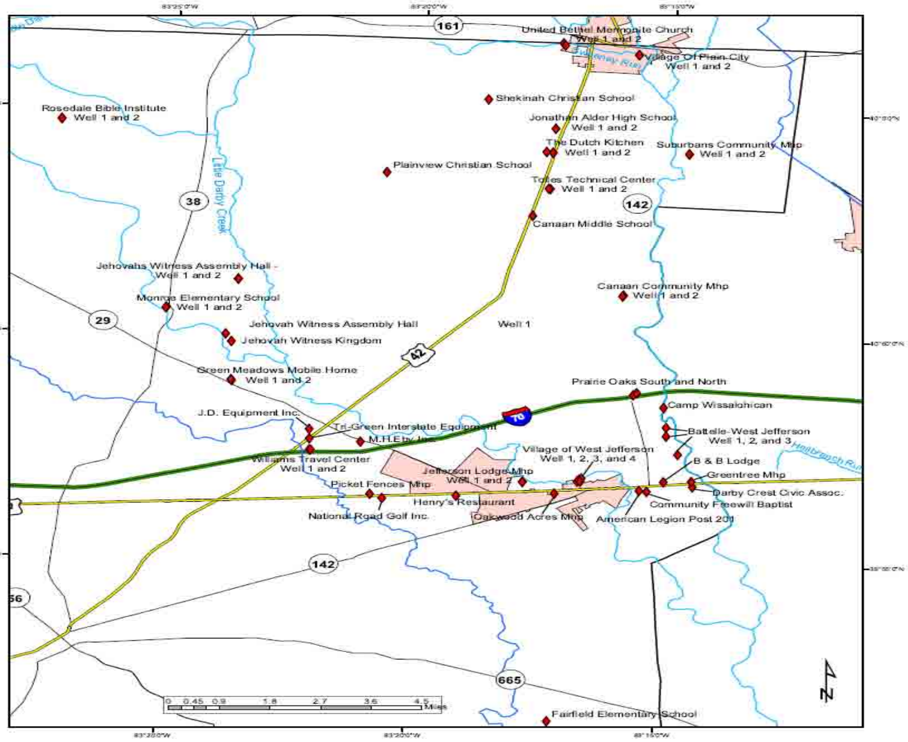
Location of Public water wells in LOGAN County



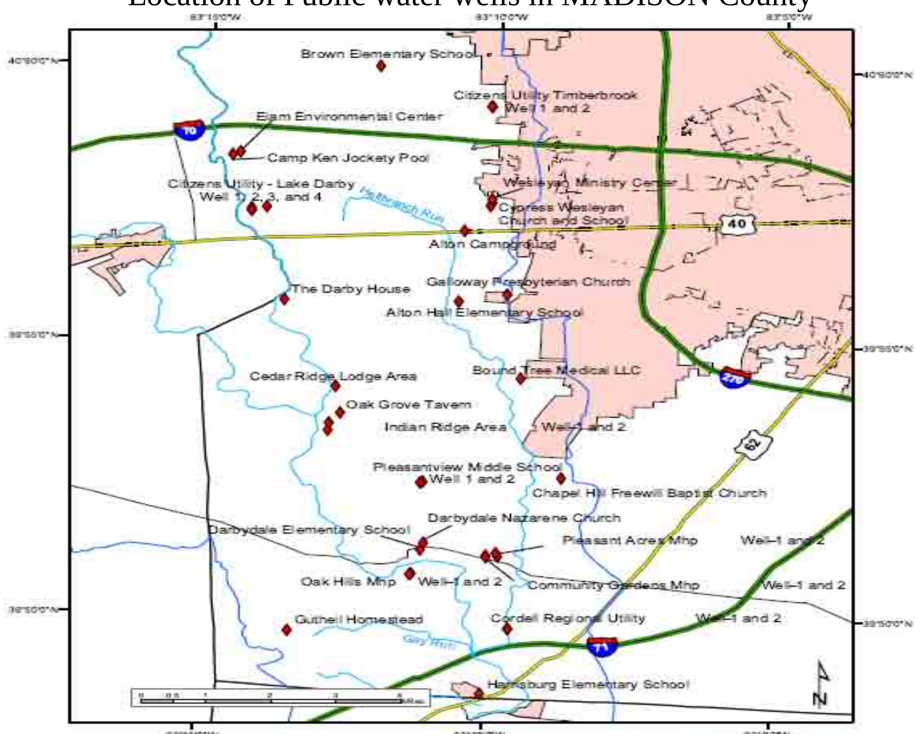
Location of Public water wells in UNION County



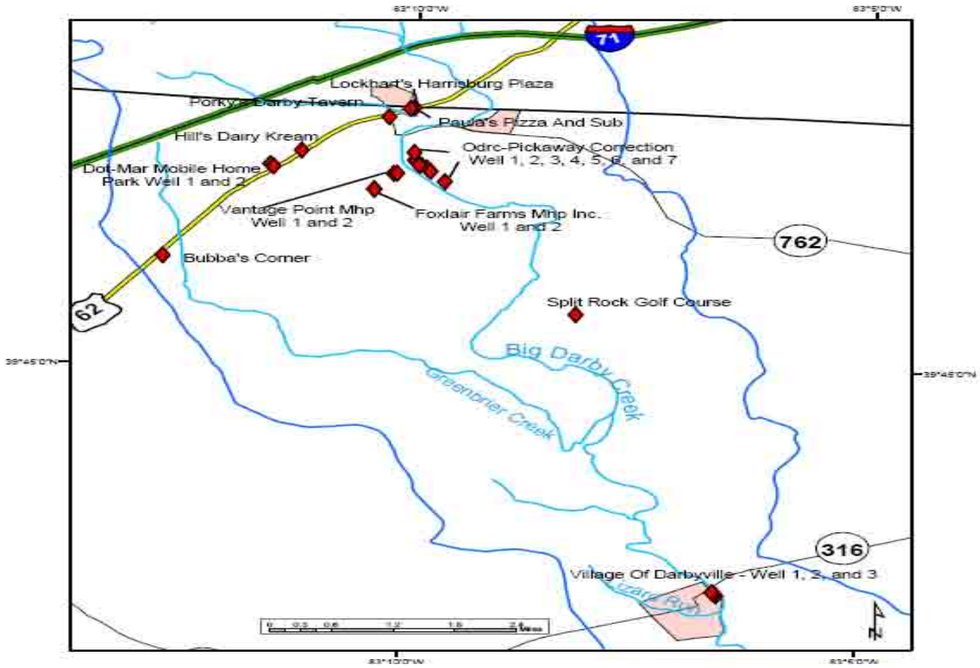
Location of Public water wells in CHAMPAIGN County



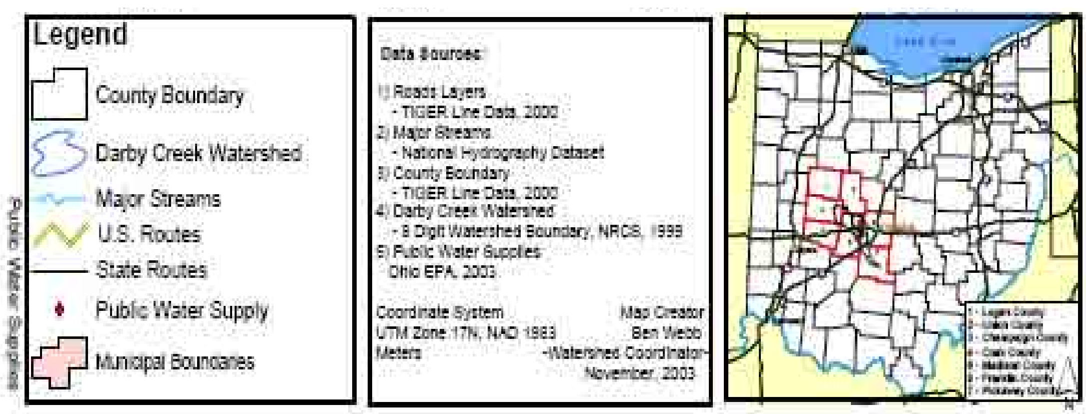
Location of Public water wells in MADISON County



Location of Public water wells in FRANKLIN County



Location of Public water wells in PICKAWAY County

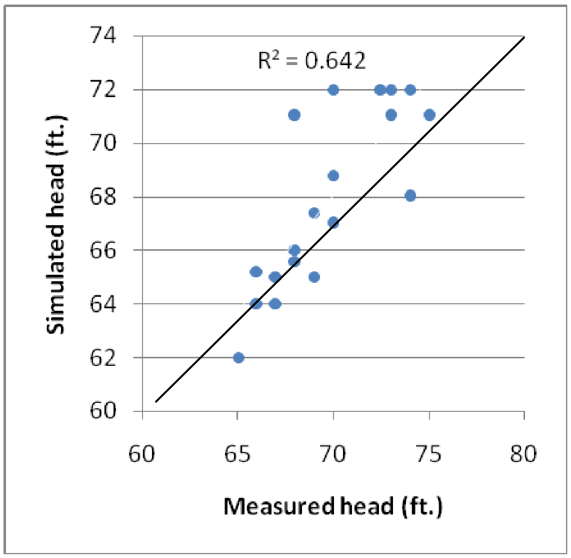


Appendix C: Monthly Precipitation Data from the Columbus Weather Stations

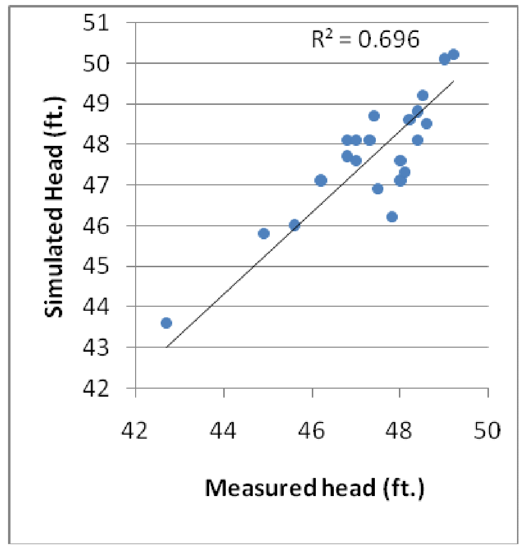
	Monthly Precipitation (in)									
Year	Circleville	Columbus VLY Crossing	Columbus WSO Airport	Delaware	Irwin	London	Marysville	Washington Court House	Westerville	Average
Jan-90	2.46	2.86	**2.43**	N/A	1.37	2.06	1.45	2.56	2.37	2.20
Feb-90	3.44	3.92	**5.15**	2.8	3.43	4.8	4.16	3.57	5.09	4.04
Mar-90	2.16	1.61	**1.32**	1.4	1.62	2.14	1.58	3.39	1.54	1.86
Apr-90	3.13	3.21	**2.82**	2.35	2.57	2.53	2.37	3.1	2.9	2.78
May-90	10.46	7.22	**7.01**	7.49	7.09	7.65	7.52	10.56	8.7	8.19
Jun-90	2.78	4.24	**5.25**	3.44	3.93	5.18	6.72	3.01	6.07	4.51
Jul-90	6.38	5.41	**8**	7.58	6.13	N/A	7.03	N/A	9.6	7.16
Aug-90	3.38	2.13	**1.86**	1.73	1.79	1.99	2.72	4.21	1.97	2.42
Sep-90	3.36	3.56	**5.26**	3.25	4.83	6.08	3.97	3.02	3.35	4.08
Oct-90	N/A	N/A	**N/A**	N/A	N/A	N/A	N/A	N/A	N/A	N/A
Nov-90	1.79	1.79	**2.03**	N/A	1.86	1.74	1.79	1.74	1.72	1.81
Dec-90	7.58	7.52	**6.98**	8.9	8.18	7.66	8.74	7.99	7.25	7.87
Jan-91	2.14	1.85	**1.97**	N/A	1.79	2.28	1.75	1.81	1.91	1.94
Feb-91	2.77	2.27	**2.3**	1.74	2.11	2.17	1.4	2.58	2.07	2.16
Mar-91	5.11	3.91	**3.97**	3.19	3.47	4.59	3.38	5.18	3.63	4.05
Apr-91	3.91	4.67	**4.15**	3.09	3.64	3.76	3.19	4.3	3.24	3.77
May-91	1.24	2.45	**2.47**	2.41	3.15	2.75	2.89	2.76	2.1	2.47
Jun-91	1.22	2.98	**2.81**	2.35	3.09	2.57	1.83	0.27	0.94	2.01
Jan-00	1.42	1.42	**2.14**	1.68	1.22	1.27	2.89	N/A	1.73	1.72
Aug-91	2.3	1.71	**2.02**	0.84	2.44	2.97	0.99	3.79	2.11	2.13
Sep-91	4.24	3.79	**4.05**	1.5	3.91	1.65	1.24	2.4	1.27	2.67
Oct-91	1.2	1.47	**1.76**	2.28	1.25	1.42	1.14	1.17	1.78	1.50
Nov-91	0.71	1.92	**1.31**	1.51	1.15	1.04	1.07	1.01	1.31	1.23
Dec-91	4.15	4.14	**3.79**	N/A	1.32	3.59	2.31	3.75	2.99	3.26
Jan-92	1.59	2.8	**1.79**	2.62	2.61	2.56	2.07	2.19	2.03	2.25
Feb-92	1.25	1.76	**0.85**	1.17	0.64	0.88	0.76	1.32	0.88	1.06
Mar-92	3.91	4.07	**3.4**	4.4	3.59	3.65	3.45	3.6	3.57	3.74
Apr-92	2.24	3.36	**2.83**	3.27	4.28	3.77	3.73	2.86	3.24	3.29
May-92	5.75	3.69	**3.4**	2.63	3.87	2.73	3.98	4.05	2.23	3.59
Jun-92	3.1	1.96	**2.33**	2.81	3.03	2.89	2.6	2.22	3.59	2.73
Jul-92	8.24	10.44	**12.36**	11.54	10.3	6.84	9.97	9.21	8.37	9.70
Aug-92	6.25	6.24	**3.75**	3.05	2.94	3.38	4.22	4.59	2.73	4.13
Sep-92	1.89	2.14	**2.14**	2.46	1.64	2.27	2.59	1.34	4.24	2.30
Oct-92	1.03	2.09	**1.4**	1.77	1.89	1.69	2.03	1.18	1.19	1.59
Nov-92	3.81	4.58	**4.03**	5.06	4.85	4.62	5.32	N/A	1.99	4.28
Dec-92	1.11	1.01	**1.32**	1.12	0.00	1.2	0.00	0.94	1.10	1.15

Jan-93	2.98	2.74	**4.14**	4.81	4.03	4.16	3.91	4.22	4.35	3.93
Feb-93	2.37	2.37	**1.82**	1.75	2.73	1.87	1.85	2.53	1.87	2.13
Mar-93	2.52	2.44	**3.5**	3.25	2.54	2.68	2.84	2.81	3.05	2.85
Apr-93	3.91	3.18	**4.49**	4.98	3.52	4.57	3.4	4.12	4.99	4.13
May-93	2.4	3.69	**2.47**	1.85	2.02	1.93	2.31	2.53	2.26	2.38
Jun-93	3.99	4.62	**3.33**	5.34	5.7	3.12	7.5	3.77	5.3	4.74
Jul-93	5.75	5.33	**5.95**	6.18	8.15	6.92	8.36	4.26	5.95	6.32
Aug-93	1.24	2.99	**0.74**	0.84	2.3	0.48	1.26	0.39	1.07	1.26
Sep-93	2.6	2.47	**1.75**	4.08	4.59	2.62	4.35	2.02	3.1	3.06
Oct-93	4.97	2.81	**3.05**	2.94	2.32	3.2	2.04	4.12	2.49	3.10
Nov-93	5.38	3.74	**4.45**	5.23	6.14	4.82	4.86	4.6	4.88	4.90
Dec-93	1.64	1.58	**2.16**	2.38	1.75	1.96	2.36	1.6	2.62	2.01
Jan-94	3.62	3.03	**3.79**	3.17	3.2	3.4	3.63	3.5	3.29	3.40
Feb-94	2.05	0.86	**1.56**	1.41	0.88	1.58	1.46	1.96	1.56	1.48
Mar-94	2.93	1.62	**1.94**	1.49	1.74	1.75	1.37	2.64	1.94	1.94
Apr-94	5.46	6.22	**3.64**	2.76	4.22	2.96	2.66	5.87	3.47	4.14
May-94	2.57	1.64	**1.69**	3	1.92	3.38	2.16	2.58	2.52	2.38
Jun-94	3.55	4.43	**1.93**	5.75	5	3.54	4.28	4.21	4.48	4.13
Jul-94	5.39	5.86	**6.02**	3.96	2.09	2.23	3.45	2.13	3.7	3.87
Aug-94	2.77	5.24	**3.29**	4.7	3.95	3.36	4.48	4.03	4.72	4.06
Sep-94	3.35	2.16	**1.68**	0.76	0.92	1.12	0.83	1.01	0.84	1.41
Oct-94	0.79	1.07	**0.92**	0.76	0.83	0.67	0.63	0.64	0.78	0.79
Nov-94	2.6	3.31	**2.94**	3.73	2.92	3.77	2.94	3.42	3.21	3.20
Dec-94	2.74	2.61	**2.22**	2.18		2.18	2.17	2.18	2.35	2.33
Jan-95	4.36	5.8	**4.54**	5.18	3.75	4.08	3.34	3.8	4.85	4.41
Feb-95	1.53	1.82	**1.64**	0.88	0.65	1.32	0.89	1.47	1.5	1.30
Mar-95	1.47	1.97	**1.61**	1.45	1.84	1.62	1.32	2.2	1.47	1.66
Apr-95	3.12	3.44	**3.17**	3.03	4.4	3.79	4.1	3.41	3.56	3.56
May-95	7.32	7.16	**4.86**	5.95	7.55	6.34	6.72	7.02	6.29	6.58
Jun-95	4.39	3.85	**5.3**	8.01	4.84	7.75	7.94	2.86	7.93	5.87
Jul-95	1.82	4.92	**6.99**	5.52	3.94	6.39	3.96	1.46	5.13	4.46
Aug-95	3	5.7	**7.56**	5.44	4.83	5.59	4.71	4.62	5.19	5.18
Sep-95	0.99	0.47	**1.15**	1.1	1.27	1.61	0.83	1.92	0.78	1.12
Oct-95	3.93	4.66	**4.04**	4.45	3.78	4.46	3.98	3.63	4.24	4.13
Nov-95	1.94	2.33	**2.47**	3.03	1.74	2.53	2.09	1.91	2.27	2.26
Dec-95	2.57	2.24	**1.97**	1.66	2.66	2.97	2.05	2.54	1.87	2.28
Summation	**231.51**	**241.36**	**235.22**	**224.43**	**226.52**	**225.09**	**230.74**	**213.65**	**230.73**	**228.81**

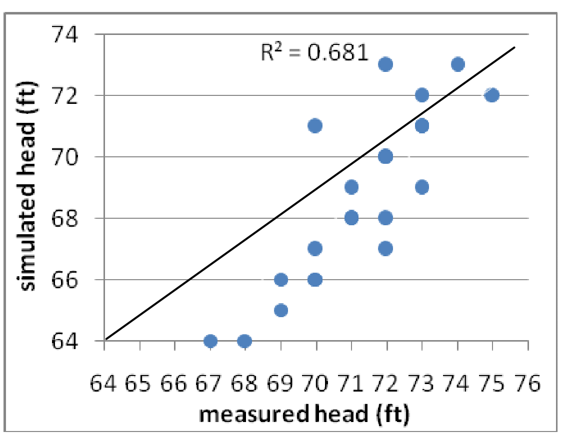
Appendix D: Hydraulic head correlation for the Tenmile calibration wells.



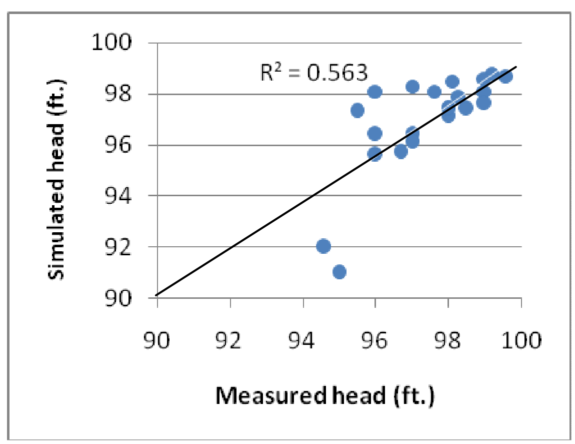
(a) Well 15



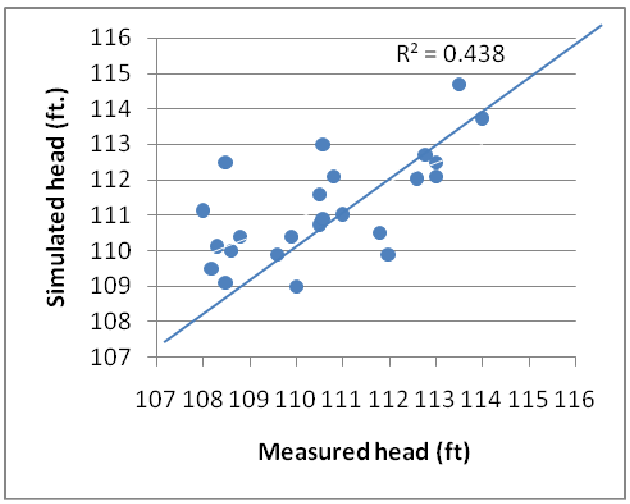
(b) Well 17



(c) Well 18



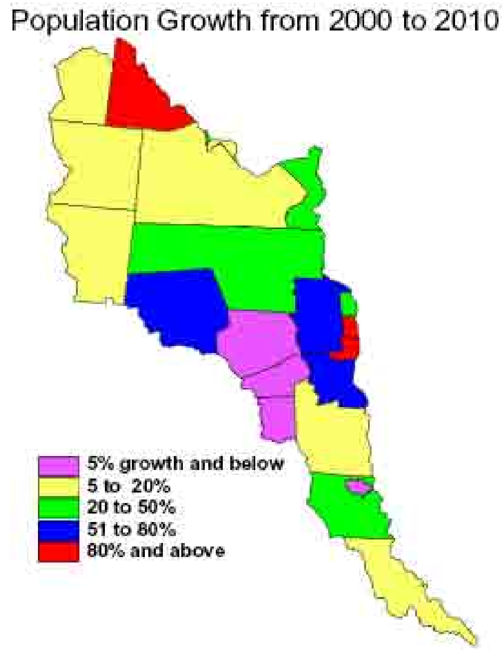
(d) Well 20

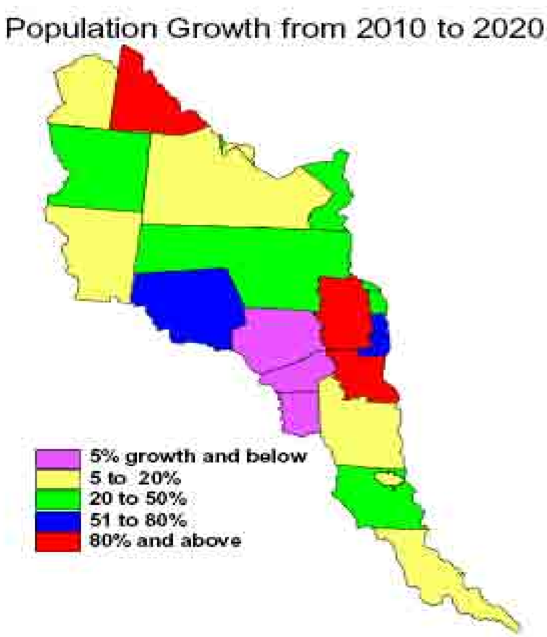


R² = 0.438
116
115
114
113
112
111
110
109
108
107
Simulated head (ft.)
107 108 109 110 111 112 113 114 115 116
Measured head (ft)

(e) Well 21

Appendix E: Projected population increases in the Big Darby watershed





Source: Gordon, Steve, et. al., 2001. Development and Change in Big Darby Watershed. Department of City and Regional Planning, OSU. pg. 47.

CURRICULUM VITAE

Alphonce C. Guzha

EDUCATION

2003-2008	Utah State University PhD Water Resources Engineering
1994-1996	Sokoine University of Agriculture MSc Agricultural Engineering (Irrigation Engineering)
1991-1993	University of Zimbabwe BSc Soil Science and Agricultural Engineering

ACADEMIC AWARDS

Ph.D.: William J. Fulbright Scholarship

Dean's List: 2003 and 2004 academic years. Utah State University

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BSc: University book prize-best student - 1992 and 1993 academic years

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