2 Background

2.1 The Hydrologic Cycle

The hydrologic cycle describes the pathways that water takes on, in all its phases, as it moves between the atmosphere, land surface, subsurface, and open water. These pathways, hydrological processes, result from the interaction of the meteorological, geological, and vegetative conditions. A conceptual diagram of the hydrologic cycle is presented in Figure 2.1

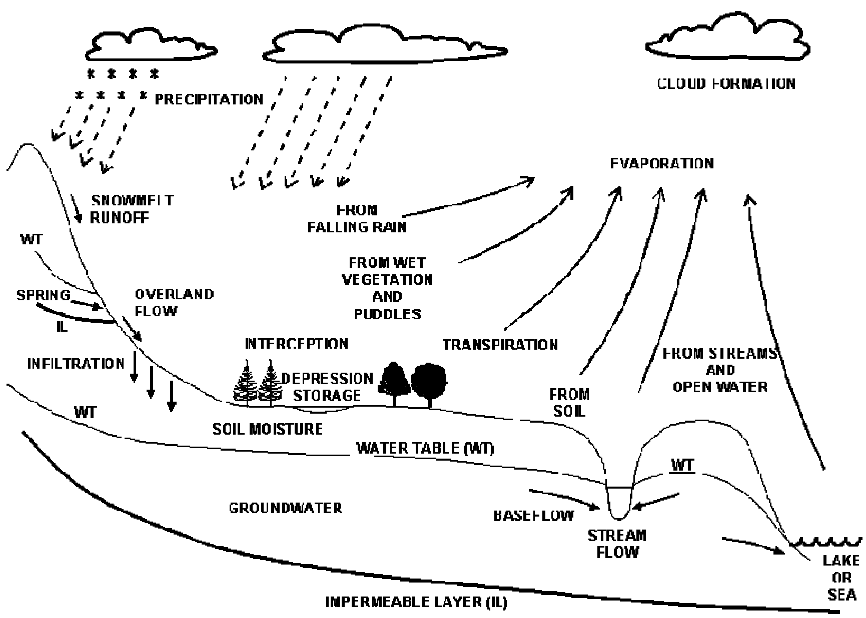

Figure 2.1 Conceptual Model of the Hydrologic Cycle (after Brutsaert, 2005)

The hydrologic cycle is comprised on several components that are all inter-connected. There is no start or end in the hydrologic cycle. It is the continuous movement of water over, on, and below the Earth's surface. To describe the hydrologic cycle we will begin with precipitation which is the driving force behind the movement of water over and through the land. The accumulation and condensation of water vapor in the atmosphere forms precipitation which delivers fresh water (or snow) to the land and oceans.

Precipitation that occurs over land will fall over the land surface or onto surface waters which will return the water to the ocean through the surface water drainage network. In some cases water in the river network may infiltrate into the ground before discharging to the surface again downgradient. Precipitation that is not directly captured by the surface water network will fall onto the land surface or may be intercepted by vegetation. Some of the intercepted precipitation transfers to the ground via throughflow and stemflow while remainder is allowed to evaporate.

The duration and intensity of the precipitation event in combination with the surficial geology will determine whether the precipitation that reaches the ground will infiltrate into the subsurface or flow over land. Overland flow occurs when the vadose zone becomes saturated from below (i.e., the water table rises to ground surface) or if the rate of precipitation is greater than the rate if infiltration, saturation from below. Overland flow may occur as run-on, whereby the water flows over land to become part of depression storage or infiltrate into the subsurface at a down gradient location, or as runoff, which is overland at reaches a surface water body.

The infiltrating water can be captured by plant uptake within and just below the root zone and is transpired through the vegetation canopy. The water that migrates beyond the zone of influence of evaporation and root uptake is referred to as percolation and will eventually reach the water table, providing recharge to the groundwater flow system. Where the water table intersects the ground surface the groundwater discharges to surface water. For large surface water bodies this is the shoreline line and for tributaries the discharging groundwater provides base flow to the streams, which varies throughout the year.

Finally, water is returned from the land and oceans to the atmosphere via evaporation and sublimation. Evaporation occurs from water on than and surface (e.g. ponded water, lakes, rivers, and oceans) and from water in soil pores near the land surface. A limited amount of water can also evaporate from precipitation before reaching the land or oceans.

Historically, for simplicity, the sciences of groundwater and surface water were treated as separate entities but from the description of the hydrologic cycle it is apparent that they are intimately connected and in fact ought to be viewed as a single resource. When discussing the hydrologic cycle one must bear in mind the temporal and spatial scales at which hydrologic components are being described (Winter et al., 1998). The various spatial and temporal scales that the hydrologic cycle operates on are illustrated in Figure 2.2.

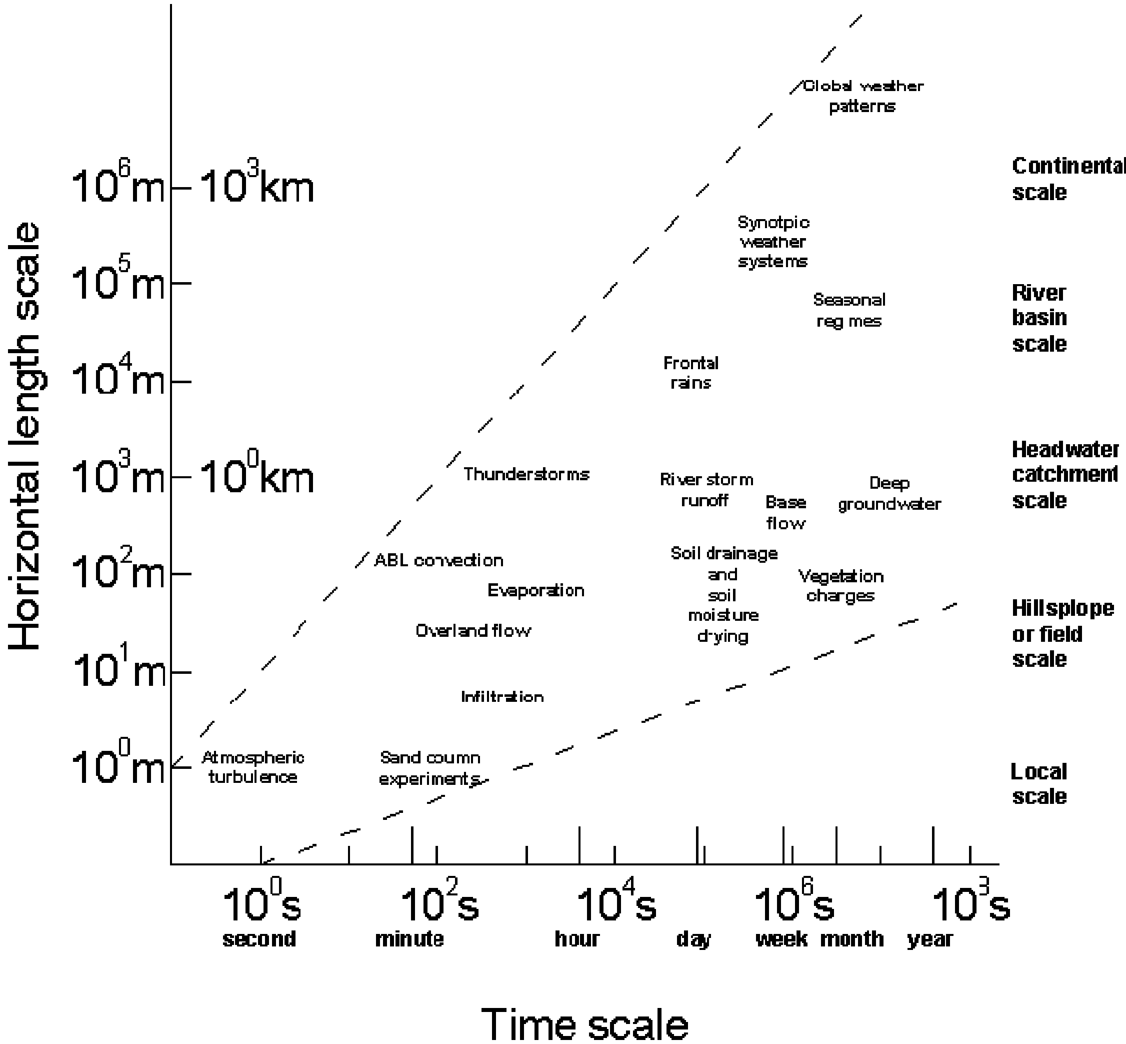

Figure 2.2 Spatial and Temporal Scales of the Hydrologic Cycle (after Brutsaert, 2005)

2.2 Simulation of the Hydrologic Cycle

When all the components of the hydrologic cycle are accounted for as inputs, outputs, and storage mechanisms the hydrologic cycle can be idealized a closed system that conforms to the principal of conservation of mass (of water in all phases). Since the system is closed it can quantitatively be viewed with the concept of a water budget where the movement and storage of each component is tracked and accounted for each component within the system.

All the hydrological processes are inter-related and dependant on each other. Whether a hydrological process is considered an input or an output is dependant on the point of view. With respect to groundwater systems the balance between the inputs and the outputs to the system can be expressed as:

$$R = P - I - ET \pm O \pm \Delta S \qquad \text{Equation 1}$$

where: R is the recharge reaching the aquifer (percolation);
P is the precipitation;
I is the interception by the vegetation;
ET is the sum of evaporation and transpiration;
O is the lateral overland flow; and
ΔS is the change in water storage in the unsaturated and saturated zones;

The components of the hydrologic cycle must be accounted for in a deterministic, physically-based manner. To this end, Freeze and Harlan (1969) drafted a conceptual framework for the numerical modeling of the mechanisms that describes the movement and storage of water within a closed system, referred to as the “Blueprint”. The major limitations of physically-based approaches include issues of scale, parameterization, and computational intensity.

In contrast to the physically-based approach conceptualized in the Blueprint are systems based (also referred to as empirical or “black box”) approaches. In this approach the physics involved in the hydrologic processes are not considered, rather a mathematical relationship is derived between inputs (e.g., precipitation) and outputs (e.g., stream flow). This type of approach is often referred to a as a “black box” approach since the mechanisms and the functional relationships of the hydrologic processes are not represented. There are two major limitations to the systems based approach. First, parameters are often lumped and lack physical meaning implying that the conditions may need to be characterized based on the modelers experiences and biases. The second limitation is that these models cannot be used as a predictive tool for stress conditions (e.g., climate change, urban development) that are outside its calibration.

A third classification of a hydrologic response model, referred to as a “grey box” by Brutsaerrt (2005), is one that is an intermediate approach to the physically-based and systems based approaches. In these types of models a Hydrologic Response Unit (HRU) is defined based on unique combinations of areas that contain similar properties (e.g., soil, slope, vegetation characteristics). The relationship between the inputs and outputs are based on simplifications that retain some physical meaning but lack the rigorous descriptions of the relevant physical processes.

In order for the hydrologic response model to be an effective tool that can produce meaningful results it must be able to simulate all of the significant hydrologic process operating within the area of study. If the model is to be relied on in a predictive capacity it must not only be able to represent historical and current conditions but also future scenarios under new and different stress conditions. .

2.3 The Physically-Based Modeling Approach

Freeze (1969) eloquently and concisely provides a description of what physically-based modeling is:

> "In a physically-based mathematical model, the component, time-dependant hydrologic processes are represented by a set of partial differential equations, interrelated by the concepts of continuity of mass and of momentum. These equations, together with the boundary conditions that define the shape and boundary properties of the basin, comprise the composite boundary value problem that is the hydrologic response model. A boundary value problem of this complexity must, by its very nature, be solved by numerical techniques and a digital computer."

Typically, modeling the hydrologic response of a catchment is done with a surface water model or a groundwater model. Surface water models tend to oversimplify subsurface processes and vice-versa, groundwater models oversimplify surface water processes. This simplification is typically treated by adding a source/sink component to the model to account hydrologic processes occurring outside the model domain of interest. For example, in surface water models infiltration into the subsurface may be simulated as a sink with no further provisions given for routing the water. Similarly, in groundwater models boundary conditions are specified to represent aquifer communication with surface water features. In either case, as water moves from one domain to another it is considered to be "lost" from the system.

In order to simulate the hydrologic response as presented by Freeze and Harlan (1969) the groundwater and surface water domains need to be 'coupled' or 'integrated' into a single framework. The communication between the models can be done in a 'sequentially coupled' or 'fully-coupled' approach.

The sequentially-coupled approach can be implemented by either 'externally' or 'internally' coupling the hydrologic models at the water table. With externally coupled models the simulation result from one model is applied as a boundary condition to another. Examples of this approach are Jyrkama et al. (2002), and Scibek and Allen (2006) who used HELP3 (Schroeder et. al., 1994a) to simulate the processes of infiltration, evapotranspiration, and runoff to estimate the recharge reaching the watertable, which is used as a boundary condition to a three-dimensional groundwater flow model, MODFLOW (e.g. McDonald and Harbaugh, 1996). In this case, there is no feedback from the groundwater model to the HELP3 model. The groundwater and surface water models can also be internally coupled as is done in MIKE-SKE (Abbot et al., 1986a/b). In doing so, the model iterates between the solutions provided by the groundwater model (i.e., determining the location of the water table) and the fluxes from the unsaturated and overland flow model to the water table; completing the feedback loop is between the position of the water table and the fluxes to the water table are completed.

A fully-coupled, or fully-integrated, groundwater-surface water modeling approach links the subsurface and surface domains at the land surface boundary through first-order flux relationships. This provides a robust and physically-based simulation approach as all the principal mechanisms for generating stream flow are accounted for, groundwater discharge providing base flow, subsurface storm flow, Dunne overland flow (saturation from below), and Horton overland flow (saturation from above). Examples of integrated models in include Integrated Hydrologic Model (InHM) (VanderKwaak and Loague, 2001), MODHMS (Panday and Huyakorn, 2004), and HydroSphere (Therrien et al., 2005). In a study comparing sequential (internally-coupled) and fully-coupled models Fairbanks et. al., (2001) concluded that the fully coupled approach is robust and provides reliable solutions while sequentially-coupled models perform well when the interaction fluxes between the surface and subsurface domains are low.

2.3.1 Advantages of the Physically-Based Modeling Approach

The numerical representations of the hydrological process are based on mimicking the mechanics and properties of the processes involved whose parameters can be measured and have a physical meaning. The use of physically-based, numerical models to simulate water flow (pathways and velocities) allows for further numerical modeling to be carried-out that is dependant the water flow path such as sediment, solute, and reactive transport.

Since the simulation results are directly dependant on distribution of parameters, such an approach is well suited to assess the potential impacts for many types of diverse water resource problems facing the environment and society today through the use of “what-if” scenarios. The parameter distributions of the numerical model can be modified to reflect the properties of future conditions and the numerical model can be re-run to simulate the outcome of such conditions. Issues that impact hydrological processes that can be evaluated using the “what-if” approach include (but are not limited to) evaluating: changes to land-use practices (e.g., urbanization, deforestation, agricultural development, etc.), sustainable water resource development, impacts to groundwater and surface water exploitation and contamination, and the impacts of climate change and salt water intrusion.

Given that the numerical model is based on the discretization of partial differential equations describing the relevant physical processes with physically meaningful parameter values the model can be applied different geographic locations. To further this point, Bathurst and O'Connell (1993) raised the point that physically-based models can be applied with more confidence at different locations and under different conditions than those they were validated for because of the physical laws they embody.

Furthermore, these models can be used to provide the user with feed back as to the spatial distribution of the sensitivity of model input parameters. This information can be used to guide efforts to collect field information in locations where it will provide the most information and to aid in the interpretation of the simulation results (e.g., Gillham and Farvolden, 1974; Sykes, 1985).

2.3.2 Disadvantages of the Physically-Based Modeling Approach

The advantages of physically-based modeling are appealing however interpretation of the simulation results must be done with an appreciation for limitations of the particular physically-based model(s) employed. The limitations of models that follow the "Blueprint" stem from the assumptions made in the formulation of the mathematical expression representing the hydrological processes, the techniques employed by numerical schemes to find "well-behaved" numerical solutions to the mathematical expressions, and the issue of scale in hydrology. In addition to these, due to the complexity involved, especially when considering integrated groundwater–surface water modeling, requires the modeler to have a broad and sophisticated knowledge base.

Distributed, physically-based modeling requires that the modeler have a specialized skill set which includes understanding the hydrological response mechanisms of the basin under study, parameterization of the properties and boundary conditions, model calibration and validation, and understanding for the relationship between the conceptual, mathematical, and numerical models (Bathurst and O'Connell, 1993). In fact, distributed, physically-based models have become so detailed and complex that the numerical model can be a problem in its own right (Bear et al., 1992).

Since the parameters have a physical meaning they can not be arbitrarily adjusted to fit the performance measure dataset but rather, may be calibrated to within an acceptable range of values. That is, the model must producing accurate results for the right reasons (Klemes, 1986). This is particularly challenging given the disparity in scales between hydrologic models, hydrologic processes, and the measurement of materials properties.

In a critical review of numerical models, Oreskes et al. (1994) contend that the verification and validation of numerical models is impossible because in reality natural systems are never be truly closed and model results are always non-unique. Though a model can simulate and confirm limited field observations they must be viewed with skepticism and should not be held as absolute truth. Oreskes et al. (1994) insist that the primary value of a model is a heuristic tool and that they should be utilized as an aid in the decision making process. Essentially, one must bear in mind that though physically-based, distributed models have inherent advantages over empirical models, all models are incomplete abstractions from the natural system and increased sophistication does not guarantee more accurate results. The state of the science (and art) of hydrologic modeling requires that models be used as a tool in the decision making process (e.g., Bear et al., 1992; Bredehoft, 2003; Neuman and Wierenga, 2003).

2.3.3 Parameterization Requirements

Distributed modeling implies that the discretization of its spatial and temporal increments must be finer than the material distributions and process being modeled (Vieux, 1993). The degree of parameterization required is directly related to the scale of discretization. As the discretization becomes more refined the complexity of the problem increases as it necessitates that a greater number of parameter values are specified.

Specifying the all the necessary parameter values for each computational node/element requires consideration be given as to how the data is distributed over the model discetization. In the case of sparse datasets (e.g., point precipitation measurements, geologic boreholes) the information needs to be distributed in a realistic and continuous manner across the model domain. A variety of algorithms exist to interpolate and extrapolate sparse datasets, each having its own advantages and disadvantages. Ideally, sparse datasets should be distributed in a physically-based, meaningful manner which entails giving consideration to the physical processes that give rise to the distributions that is being estimated by the algorithm. The kriging method is often used as it honors the input data, inherently considers the spatial structure of the dataset, and provides an estimate of the uncertainty. However, this method is computationally intensive and can be highly sensitive to selection of the proper semi-variogram (e.g., spherical) and the geostatistical method employed (e.g., ordinary kriging) all of which can have a significant impact on the estimated distribution (Zimmerman et al., 1998 and Patriarche et al., 2005).

The detail and complexity of physically-based, distributed models results in intensive parameterization requirements. The effective values of all the model parameters can not be known for the entire area of study, making physically-based models vulnerable to the criticism that the model's capability exceeds data availability (Vieux, 1993). The concern of over-parameterization was recognized by Freeze (1971) at the advent of physically-based distributed modeling to which he replied that if the deterministic approach can be shown to have greater value relative to the empirical approach (that requires less data) it would encourage the measurement of necessary data.

The uncertainty associated with the model structure, model parameters, and their distributions leads to the problem of *Equifinality* (Beven, 1993). The problem of equifinality is that multiple parameter sets and model structures can produce equally acceptable fits to the observed data. At its simplest, the non-uniqueness of the saturated groundwater flow problem is most readily identified by recognizing the relationship between hydraulic conductivity and recharge. As more hydrological processes are accounted for there are more parameters available for the modeler to adjust.

Concerns relating to parameterization requirements are not limited to over-parameterization but also the ability to measure effective parameter values. Beven (1993) states that physically-based models, by their nature, are designed to have parameters that are physically measurable. This is not always the case; mathematical descriptions used to simulate the hydrological processes of infiltration (e.g., Brooks-Corey, 1964) and evapotranspiration models (e.g., Kristensen and Jensen, 1980) employed by MODHMS and HydroSphere, for example, requires for specification of parameters that have little or no physical meaning. In this regard, some aspects of deterministic, physically-based modeling have not improved on empirical approaches.

2.3.4 Issues of Scale

The issue of scale is one that affects all the hydrologic processes represented in groundwater and surface water models. The various hydrologic processes (identified in Figure 2.1) occur at different temporal and spatial scales (as shown in Figure 2.2). For example, precipitation events occur over large areas for short durations, overland flow occurs over short distances and is rapid, while groundwater flow occurs over large areas and is relatively slow. Even within a given hydrologic process the effects of scale can be significant, for example, surface tension of water and presence of macropores affecting the processes of infiltration or inundation heights of roughness elements affecting overland flow. In numerical models, the spatial effects of scale result from a mismatch between the size of the model elements and the heterogeneity and structure of the physical system. In an analogous manner, scale effects also occur due to a mismatch in temporal scales.

The spatial effects of scale are manifested in two manners. The first case is the ability of the model's discretization to capture the heterogeneity within the catchment and the second is in representing the variability within the model elements themselves (Bathurst and O'Connell, 1993). In addressing the first point, avoiding down-scaling could potentially be addressed by taking measurements for model parameters for every element however this not practical nor do measurement techniques exist to collect subsurface data at the element scale (Beven, 1993). With respect to up-scaling material and boundary condition properties on the sub-computational level to the scale of the model element no standard methods exist. For example, in a review of the scales of processes occurring within the vadose zone (i.e., the effective representation of unsaturated hydraulic conductivity values) Harter and Hopkins (2004) find that many of the approaches taken by researchers to address the issues of scale have had some measure of success but that the assumptions that govern the use of these techniques may be exceedingly restrictive and their conclusions echo the concerns of Klemes (1986) that there is no universal solution to the problem posed by the hierarchy of significant process scales. Even in a 'relatively' homogeneous media detailed studies have shown the effects of scale, for example, Sudicky (1986) finds that values of saturated hydraulic conductivity can vary by orders of magnitude over very short distances in the clean, homogenous sand of the Borden aquifer. The scale dependence of hydraulic conductivity with respect to the measurement technique is well known in the field of hydrogeology (e.g., Bradbury and Muldoon, 1989; Hart et. al., 2006). In fact, all model parameters are subject to the effects of scale.

The traditional approach taken when characterizing large areas with limited data is that of 'zonation' (Jenson and Mantoglou, 1993). In this approach homogenous model parameters are assigned to areas conceptualized to have similar hydrologic response or material properties (e.g., urban vs pasture). These zones simplify the model calibration by effectively reducing the degrees of freedom. The zonation approach results in a lumping of parameters and material properties which, degrades the robustness of the physical-basis of the numerical model and may result in the misrepresentation of hydrological processes due to inappropriate parameter value selection.

2.4 Challenges of Fully-Integrated Groundwater-Surface Water Models

Brustsaert (2005) summarizes the major concerns associated with distributed physically-based modeling as:

i) never being able to accurately measure the properties of natural catchments, and

ii) that the solutions to the numerical implementation of the mathematical expressions representing the hydrologic processes can only be obtained for grossly idealized conditions, which are coarse approximates of the dynamics of field processes.

The limitations of i) have already been discussed, the challenges of parameter value distribution and scale when implementing a distributed, physically-based approach. With respect to ii), the limitations of primary of concern are in the representation of the unsaturated zone as it links the subsurface and surface domains thereby exerting controlling influence over the interaction between precipitation, recharge to groundwater, and surface runoff. The physically-based modeling of the overland flow and surface waters as part of a fully-coupled model also poses significant challenges in simulating event-scale hydrologic processes in a complete and rigorous as presented in the Blueprint. The issues relating to the numerical approaches to simulating variably saturated and overland flow are discussed.

2.4.1 Flow in the Unsaturated Zone

Proper representation of the unsaturated zone is critical for conjugative groundwater–surface water modeling. The unsaturated zone links the subsurface and surface domains and controls the partitioning of precipitation between infiltration and overland flow. Flow in the unsaturated zone is more complicated than the saturated zone because there are two fluids present, water and air. In this two-phase system water is the wetting fluid, meaning that it preferentially coats the soil grains, and air is the non-wetting fluid. As water migrates through the unsaturated zone it displaces the air within the pore space.

Unlike the saturated zone where moisture content equals the soil porosity and hydraulic conductivity is constant for the given soil texture, in the unsaturated zone both moisture content and hydraulic conductivity are dependant on the pressure head. This gives rise to what is often referred to as characteristic curves such as those in Figure 2.3.

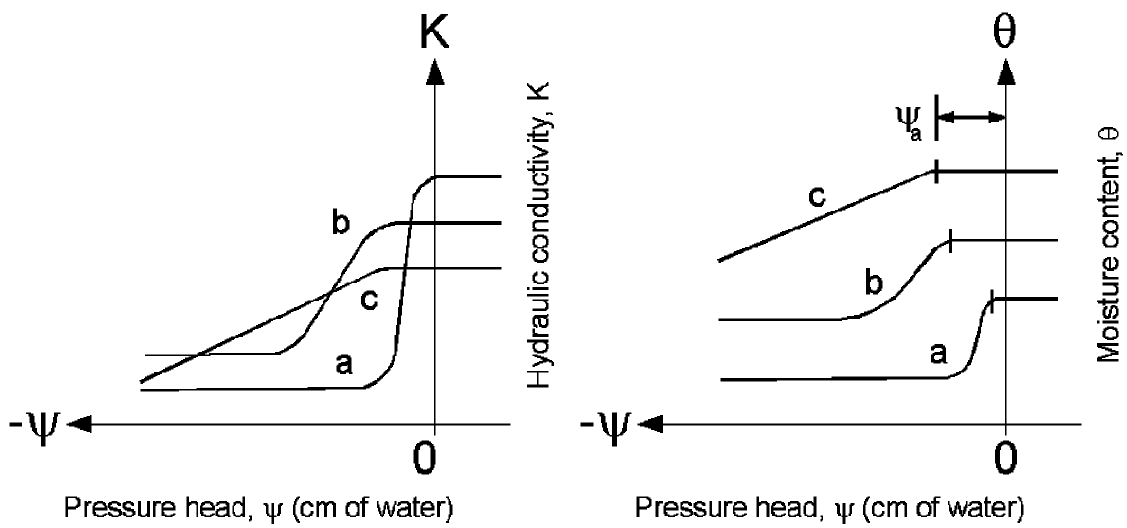

Figure 2.3 Characteristic curves illustrating the relationships between pressure head (ψ), hydraulic conductivity (K), and moisture content (θ) (after Freeze and Cherry, 1979)

Due to surface tension of water and air the hydraulic head in the unsaturated zone is less than atmospheric (i.e., pressure head is negative), also referred to as capillary pressure. Smaller pore throats, typical of silt and clay materials, create more surface tension than larger ones, such as sand. Thus, for a given pressure head a coarse-grained material will have less moisture content than a fine-grained material. As the soil moisture increases progressively larger pore spaces (with lower surface tensions) fill with water. This relationship between negative pressure head and soil moisture, referred to as the water-retention curve, is non-linear. Furthermore, the shape of the curve is dependant on whether the soil is wetting or drying. This occurs because drainage is controlled by the smallest pores, whereas wetting is controlled by the largest. The scanning curves between the main wetting and drainage curves illustrate the hysteretic effect of antecedent moisture conditions which can play a significant role in the response of a soil type to an advancing wetting front (Freeze, 1971b).

Functional relationships for soil water retention characteristics and their relation to hydraulic conductivity, as shown in Figure 2.3, have been developed by Brooks and Corey (1964), Mualem (1976), and van Genuchten (1980). The primary difference between the Brooks-Corey model and the Mualem and van Genuchten models is that the Mualem and van Genuchten models are continuous functions for the entire range of pressure heads while the Brooks-Corey model specifies a minimum displacement pressure which is the displacement pressure of air that water must overcome before infiltration can take place. The values of the parameters used in the functional relationships for various soil textures are based on fitting curves to observed data. These parameters may be given physical descriptions (e.g., lp is pore connectivity) but none of these parameters can actually be measured and have little or no physical meaning.

2.4.2 Numerical Simulation of Flow in the Unsaturated Zone

The equation describing the movement of water in the unsaturated zone is given by Richards (1931), referred to as the Richards' Equation. The Richards' Equation is derived by applying the Darcy Equation for multi-phase flow (i.e., assuming water and air are the only phases present) with the principle of conservation of mass equation, while making some simplifying assumptions.

The equation is developed for a representative elementary volume (REV) where the properties of the medium are assumed to be constant (Bear, 1972); the scale of the REV is on the order of 10^{-2} m to 100 m (Harter and Hopmans, 2004). The concept of a REV is not used to discretize the model domain for groundwater modeling on the catchment scale (or greater) and thus is subject to the effects of scale.

The Richards' Equation incorporates many assumptions that place restrictions on the appropriate use this equation, which typically include: laminar flow, that inertial forces, velocity heads, temperature gradients, osmotic gradients, and chemical concentration gradients are all negligible, that the porous medium is non-deformable, that water is incompressible, and that the air phase of infinitely mobile (Freeze, 1971a).

$$\frac{\partial}{\partial x_i}\left[\frac{k_{ij}^0 k_{rw}}{\mu_w}\left(\frac{\partial p_w}{\partial x_j}+\rho_w g\frac{\partial z}{\partial x_j}\right)\right]\pm\Gamma = n\frac{\partial S_w}{\partial t} \qquad \text{Equation 2}$$

where:

k_{ij}^0	are the components of the intrinsic permeability tensor;
k_{rw}	is the relative permeability;
μ_w	is the absolute viscosity;
p_w	is the fluid pressure;
ρ_w	is the density;
g	is the gravitational constant;
z	is the elevation relative to a reference datum.
Γ	is used to represent sources and sinks (e.g., evapotranspiration)
n	is porosity; and
S_w	is the storativity.

In the case of saturated groundwater flow the k_{rw} term becomes unity and Equation 2 becomes a linear PDE which is much less computationally intensive to solve. Generally speaking, the assumptions noted for the development of the Richards' Equation are reasonable and allow for its use in a wide range of applications. However, since the effects of temperature are precluded from its development it does not allow for the simulation of the hydrologic response at freezing temperatures.   This poses a restriction on its application as a tool to simulate long-term, continuous hydrologic response for geographic locations that experience freezing temperatures that are seasonally prevalent. When the water in soil pores freeze the void space becomes increasingly restricted, causing increased tortuosity and a decrease in hydraulic conductivity, hence reducing infiltration and promoting overland flow. As the soil water freezes it expands to become ice which deforms the soil matrix, violating the premise that the medium is non-deformable. The impact of frozen soil during the winter also has implications when the soil thaws during the spring, increasing the hydraulic conductivity and allowing for the rapid infiltration of snowmelt (Jyrkama, 1999).

2.4.3 Vertical Discretization of the Unsaturated Zone

Due to the relationships captured by the characteristic curves, between pressure head – soil moisture and pressure head – hydraulic conductivity, the Richards' Equation is a highly nonlinear partial differential equation.

Both the finite difference (e.g., Panday and Huaykorn, 2004) and finite element (e.g., Therrien et. al., 2005) discretization approaches are used to solve the Richards' Equation numerically. Typically, for reasons of stability, an implicit temporal discretization scheme is employed when solving the equation numerically (Paniconi and Putti, 1994). An intermediate step is required to iteratively resolve the non-linearity presented by relative hydraulic conductivity, which is relationship expressed in the characteristic curves, such that a numerical solution can be obtained (Paniconi et al., 1991). The intermediate iteration is necessary to solve the dependant variable (i.e., pressure head) with the nonlinear terms (i.e., moisture content and hydraulic conductivity). Various iterative and non-iterative strategies for the numerical solution to the Richards' Equation are investigated by Paniconi et al. (1991) and Paniconi and Putti (1994). Some strategies are more efficient and robust than others however they are all computationally burdensome relative to solving the Richards' Equation under saturated conditions (i.e., a constant hydraulic conductivity values).

Ultimately, the speed at which the wetting front infiltrates into the soil is controlled by the hydraulic conductivity of the uppermost cells. As can be seen in Figure 2.3, a small change in moisture content can result in a large change in pressure head and hence, hydraulic conductivity. The rate at which the hydraulic conductivity increases (from initially dry conditions) is dependant on the rate at which the soil moisture content increases which in turn is a function of the negative pressure head. Thus, if the near surface vertical discretization is too coarse then a small amount of water entering will not sufficiently increase the moisture content of the entire cell and the hydraulic conductivity will remain low. In this case, the hydraulic conductivity remains low and the other concurrently acting hydrological process (i.e., evapotranspiration and runoff) will deplete the available water before it can infiltrate into the subsurface. Thus, for initially dry conditions infiltration will be underestimated while evapotranspiration will be overestimated. Consequently, even if the numerical techniques employed by the chosen model can rigorously account for the initiation of both Hortonian and Dunne overland flows, if the discetization is not adequate the simulation result will not be physically correct.

The appropriate level of vertical discretization of the Richards' Equation required for simulating infiltration, evapotranspiration, and initiation of runoff can be determined through the analysis of a spatial convergence study as demonstrated by Downer and Ogden (2004). The outcome of a spatial convergence study reveals the resolution (i.e., cell size) required to achieve a solution that accurately represents the system. In order to obtain a meaningful, physically correct solution to the Richards' Equation under variably saturated conditions a very fine vertical discretization is required to capture the non-linear response of the vadose as illustrated by the soil characteristic curves (Downer and Ogden, 2004). The results of their investigations show that a near surface vertical discretization coarser than 2 cm can result in significant misrepresentation of the hydrological process, (i.e., infiltration, evapotranspiration, and runoff). Downer and Ogden (2004) conclude that employing too coarse of a vertical discretization, especially at the ground surface; can result in a model that does not respond in an accurate, physically correct manner. In fact, this necessarily implies that too coarse of a discretization will result in the specification of physically unrealistic parameter values in order to achieve a solution that is consistent with observations.

El-Kadi and Ling (1993) have proposed using the Courant and Peclet numbers as criteria for estimating the necessary level of spatial and temporal discretization required to find an accurate numerical solution to the Richards' Equation. The Courant number is the ratio of travel by advection (in the vertical direction) to the cell size (i.e., Δz) and the Peclet number is ratio of advection to dispersion. With respect to the Richards' Equation the soil moisture diffusivity term, the product of the unsaturated hydraulic conductivity and the soil characteristic curve (Rolston, 2007), is used to represent the dispersion in the Peclet number. Numerical experiments were conducted for three soil types with a wide range of material properties (e.g., capillary height and saturated hydraulic conductivity). The results show that the Courant and Peclet numbers are highly dependant on the material type and their criteria vary by orders of magnitude for achieving an optimal solution. In general, the upper range for the Courant and Peclet numbers was found to be about 2 and 0.5, respectively. This translates to quite stringent space and time increments restrictions of 1 to 2.5 cm and on the order of seconds to minutes, respectively. Relaxing these criteria likely leads to numerical dispersion and overshooting (El-Kadi and Ling, 1993).

The computational requirements (for both storage and processing speed) are directly proportional to the level of discretization employed in the model. This signifies that the vertical discretization required to achieve spatial convergence, which is necessary to properly represent near surface hydrological processes, may be too burdensome for the current state of available desktop computing technology when simulating the hydrologic response at the watershed scale. Indeed, Harper and Hopmans (2004) points out that the current transient modeling of the three-dimensional Richards' equation that be solved in a reasonable timeframe is limited to approximately 10^6 nodes while the application of the Richards' Equation at a discretization level consistent with the REV scale it is derived (i.e., 10-2 m and 100 m or the vertical and horizontal directions, respectively) for a watershed application (e.g., 100 km2) with a 1 m thick root zone requires 10^{10} nodes. When further considering the discretization required for the remaining thickness of the subsurface (both overburden and bedrock units) and possible integration with surface water equations for overland flow (with even finer time-stepping requirements) this problem becomes daunting. At the discretization level of a REV the computational effort required is impractical and the parameterization effort is impossible.

To alleviate the numerical burden imposed by the nonlinear relationships of the vadose zone, captured by the characteristic curves, an upscaling approach has been applied whereby the response of the vadose zone at the REV scale is translated to its effective response to the scale of the vertical discretization. Harter and Hopmans (2004) provide a detailed review of the various upscaling approaches that have been take by researchers. A variety of analytical and numerical models have been developed for steady-state or transient conditions, using vertically lumped soil texture or heterogeneous soil columns. Despite the research efforts to date, Harter and Hopman (2004) identify that more research still needs to be done before the upscaling approach may utilized the modeling practitioner.

The Richards' Equation mathematically describes the rate at which water can migrate through a variable saturated porous medium. However, when considering the process of infiltration it is important to bear in mind that it is driven by precipitation events. Dunne et al. (1991) report that for soils that do not form seals the infiltration rate is positively correlated with rainfall intensity. Essentially, with increasing rainfall intensity the tendency to exceed to the saturated hydraulic conductivity for larger portions of soil increases thereby increasing the average hydraulic conductivity. There is a secondary effect here as well; as soil becomes saturated and the overland flow progressively inundates a larger portion of the downslope area where the runon can distribute the water to areas of soil moisture deficit. The portion of overland flow that does not infiltration is captured by surface water channels and runoff.

2.4.4 The Occurrence of Overland Flow

Overland flow is controlled by the response of the vadose zone and can be triggered by two mechanisms. The first is by infiltration excess, formalized by Horton (1933), whereby overland flow occurs when the rainfall intensity exceeds the infiltration capacity of the soil, essentially saturation from above. The second mechanism is saturation from below, also referred to as saturation excess, whereby overland flow can only occur once the underlying soil is saturated (Dunne, 1970). In this case infiltration causes the water table to rise until it reaches the ground surface. As such, as an increasing area of soil becomes saturated a greater surface area is available to contribute to runoff; this is known as partial contributing area or variable source area.

The theory of overland flow initiated by infiltration excess is applicable to semi-arid environments but is a rare occurrence in vegetated, humid conditions. In the latter conditions overland flow is most prominently generated by saturation excess and is a rare occurrence (Freeze, 1972a). Freeze (1972b) summarizes the necessary criteria for ponding to occur, as demonstrated by Rubbin (1966), as requiring a rainfall rate to be in excess of the saturated hydraulic conductivity of the surface soil and for the rainfall duration to be greater than the time required for the soil to become saturated. The rarity of overland flow can be recognized with consideration for the saturated hydraulic conductivity and typical rainfall intensities (e.g., less than the 10-yr return storm for Boston, Mass, US area as exemplified by Freez (1972b)). The rise of the water table to generate overland can be rapid if the capillary fringe extends to ground surface (Gillham, 1984). As such antecedent moisture conditions may play an important role in the generation of overland especially in areas of low topography (e.g., rivers).

2.4.5 Resistance to Overland Flow

In general, the pattern of flow over land is the result of gravity acting on the water in the downstream direction which is being counteracted by the internal (i.e., the viscosity of the fluid) and the external (i.e., those forces imposed by obstacles) resistance forces. All roughness elements on a soil surface contribute to the external resistance encountered by the overland flow (Rauws, 1988). This additive nature of the resistance to overland flow has been expressed as a composite, or effective, roughness (Abrahams et al., 1995). Qualitatively this represents the sum of the resistance presented by soil grains, microtopography, ground surface cover, and the standing vegetation. The resistance offered by each roughness element can be characterized by it is individual shape and size, as well as its spacing and pattern relative to the surround roughness elements (Abrahams and Parsons, 1994).

Experiments conducted in the field and laboratory generally express the effect of roughness on overland flow as the relationship between the Darcy-Weisbach friction factor and the Reynolds number. The Reynolds Number is the dimensionless ratio of inertial to viscous forces. Viscosity is dependant on temperature which can be considered constant the event scale. As such, the viscosity of the water can also be considered constant and the Reynolds Number is effectively a measure of the overland flow velocity. There are many relationships between the friction factor and the Reynolds number (e.g., Abrahams and Parsons, 1994). The typical relationships between the friction factor and the Reynolds number are a convex upward and a negative slope (Rauws, 1988), as illustrated on Figure 2.4, though others have are reported (e.g.Abrahams and Parsons, 1994). This figure highlights some of the key complexities in quantifying resistance to overland flow. It illustrates the additive nature of the forces opposing flow, in this case the contributions to simulated soil roughness from grain and form resistance. Grain resistance represents contributions from soil particles and micro-aggregates. Form resistance is the resistance offered by macro-roughness elements.

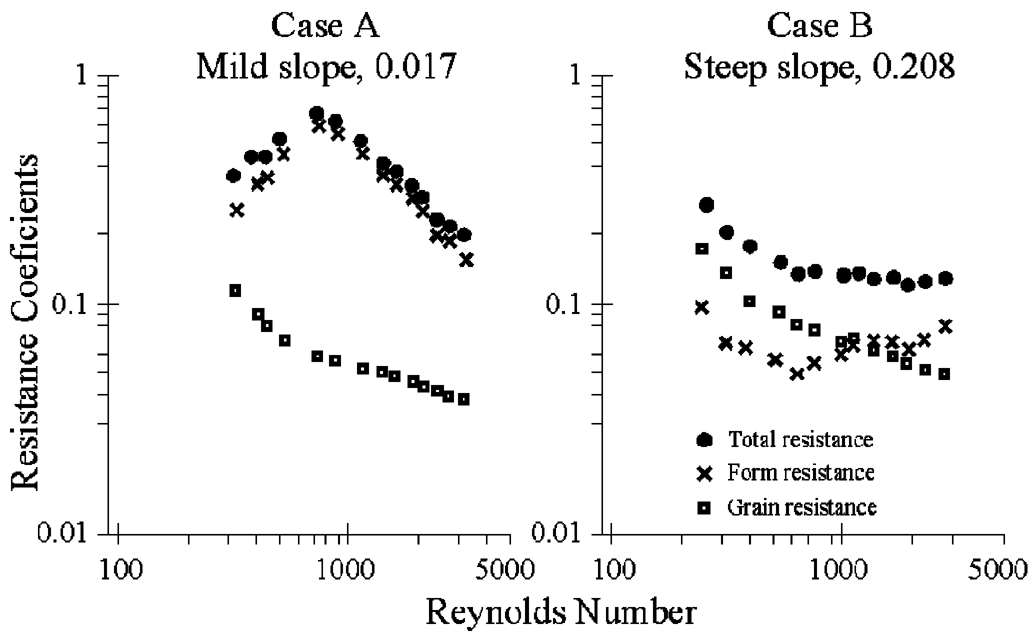

Figure 2.4 Contribution of form and grain resistance to total resistance as a function of the Reyonlds Number for mild and steep slopes (after Rauws, 1988)

The figure also clearly shows the non-linearity of roughness and its dependence on external factors, such as, in this case, slope which, limits the application of a friction factor value to a narrow range of conditions. The convex upward relationship is typically observed on mild slopes. In this situation the grain resistance has a lesser influence than the form resistance. As the depth of overland flow increases the roughness element becomes progressively inundated, increasing the wetted upstream-projected area, resulting in increased resistance. When the roughness element is submerged its resistance decreases (Rauws, 1988; Abrahams and Parsons, 1994; Lawerence, 1997). At steeper gradients a negative sloping relationship has been observed. This is attributed to the form friction no longer having an over-riding influence over grain friction and hence resistance continually decreases with increasing depth (Rauws, 1988).

The effect of vegetation on the resistance to overland flow requires due consideration. Vegetation plays an important role through retarding the overland flow velocity, thereby providing greater detention time, and allowing for greater infiltration (Dunne et. al., 1991). Vegetation is also (one of several) mechanisms by which marcopores are generated (Walker et al., 2002) which provide preferential pathways for infiltration and in so doing decrease overland flow.

Another characteristic that is particular to vegetative resistance is that it is subject to deformation with increasing depth of flow. This was studied in detail by Fathi-Moghadam and Kouwen (1997) who utilized physically-based parameters for the calculation of resistance (expressed as Manning’s n) presented by vegetation. Their experiments show that the calculated roughness values are within the acceptable ranges, as reported by Chow (1953) and Arcement and Schneider (1984), but show that the Manning's n roughness value increases proportionally to the square root of depth and is inversely proportional to the mean velocity under marginally inundated conditions. On a similar note, overland flow concentrated in rills gives rise sediment mobilization and erosion. As such, the land surface is continually being deformed, creating ever-changing preferential pathways, and altering the roughness characteristics of the land surface.

2.4.6 Impact of Human Activity to Surface Water Flow Pathways

In rural and urban areas, along agricultural plots and roadways, surface water routing features have been constructed to channel overland flow. In an urban setting this is typically accomplished with curbs and storm sewers. In an agricultural setting runoff is channeled by furrows to the edges of plots where it is diverted to ditches which may also be receiving flow fromm tile drains. Alongside roadways, road runoff is channeled to ditches and through culverts which ultimately carry the water to a streams or surface water detention ponds. In other cases roads are raised creating barriers to overland flow, creating depressions that are disconnected from the catchment outlet, reducing the effective drainage area.

The preferential pathways of human development on the land surface are typically at a finer resolution than the information (e.g., Ontario Base Maps (OBM) or a Digital Elevation Model (DEM) derived from remote sensing technology used to parameterize in regional scale models. Agricultural plots, roadways, and their associated drainage networks are not necessarily oriented coincident with the slope or aspect of the land surface limiting the application of the DEM to define the drainage on scales of the stream catchment or greater.

Duke et. al. (2003, 2006) present a methodology for downscaling DEM data with the ancillary information. The secondary information, road elevation, ditch depth, irrigation channels, and culvert location, are indirectly incorporated with the DEM when it is processed to derive the drainage pattern. A Road Enforcement Algorithm (REA) and a Channel Enforcement Algorithm (CEA) were developed to reroute overland flow paths from the well known D8, steepest decent, algorithm (O'Callaghan and Mark, 1984). The REA and CEA use the ancillary information to enforce drainage barriers and preferential surface water pathways. The effect on overland flow paths can be quite pronounced and shown in Figure 2.5 (Duke et. al., 2006).

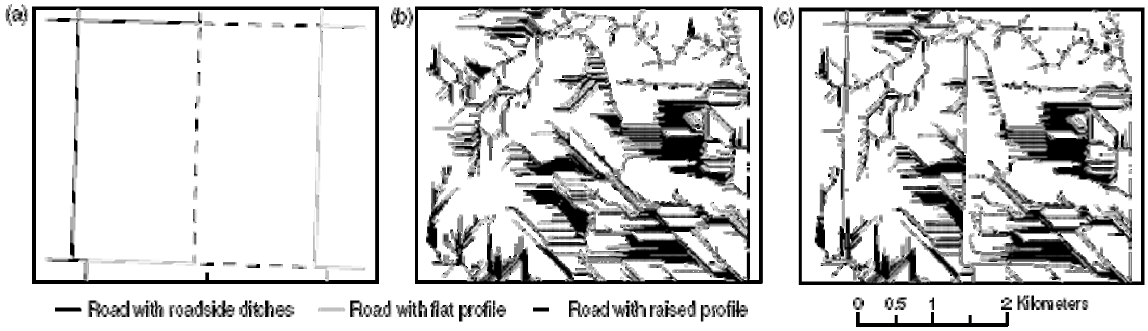

Figure 2.5 Overland flow-path patterns resulting from the REA (after Duke et. al. (2003)): (a) a typical road network; (b) D8-derived drainage pattern showing the grid cells with a runoff contributing area greater than 5000 m^2, in black; (c) REA-derived drainage.

The application of this methodology to the Piyami Drain watershed, Alberta indicated that up to 49% of the watershed area is disconnected from the surface water drainage outlet (after Duke et. al., 2006). This can have a significant impact on the dynamics of groundwater-surface water interaction; the distribution of increased surface water drainage by preferential pathways of ditches and culverts and the increased localized infiltration resulting from drainage barriers.

2.4.7 Numerical Simulation of Overland Flow

Shallow overland flow is expressed in a mathematical framework by the set of Saint Vennant equations which couple the continuity equation with the conservation of momentum equations in the x- and y-directions. This formulation of overland flow is referred to as the fully dynamic wave model. Two common simplifications to this model are the diffusion wave equation (DWE) and the kinematic wave equation models (KWE). All three levels of representation of overland flow are non-linear with the DWE and KWE bring less intensive to solve numerically.

The conditions under which the KWE is valid approximation of the fully dynamic wave model for shallow overland flow is quantified by the *kinematic number* (Woolhiser and Ligget, 1967). It is found that the KWE is suitable under many natural conditions but may be a crude approximation under smooth urban conditions. For most conditions the KWE equation has been found to be a suitable approximation of shallow overland flow with the recognition that the diffusion or dynamic wave models may be a dominant in some cases and short lived (Singh, 2002).

The KWE approximation assumes the flow be parallel to the land surface and to be in the direction of maximum slope. The underlying and limiting assumption in this level of simplification is that the downstream boundary conditions do not an effect on the overland flow, and as a result, this precludes the simulation of backwater effects. For this reason, the DWE approximation is applicable over a wider range of conditions and is employed to simulate shallow overland flow in rigorous physically-based models such as InHM, MODMHS, and HS. For further information regarding the approximations of the Saint Vennant equations the reader is referred to Vieira (1983).

The two-dimensional diffusion wave approximation for shallow overland flow is presented as (after Gottardi and Venutelli 1993):

$$\frac{\partial h}{\partial t}+\frac{\partial}{\partial x}\left(K_x H\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_y H\frac{\partial h}{\partial y}\right)=Q_e \qquad \text{Equation 3}$$

where:

	K_x, K_y	are the surface conductance terms slope in the x- and y-directions which, are dependant on the choice of relationship used to approximate the friction (e.g., Maning's *n*)
	h	is the hydraulic head (i.e., water depth plus elevation);
	H	is the water depth; and
	Q_e	is a source term (positive if entering the system)

The assumptions associated with the DWE is that the surface water flow is gradually varying with depth averaged velocities and a hydrostatic pressure distribution in the vertical direction. Furthermore, Equation 3 assumes mild surface slope, dominant bottom shear stresses and that the Manning's equation can be used to approximate a valid frictional resistance.

When combining the groundwater and surface water interactions in a fully integrated framework the horizontal discretization for the groundwater and surface water models is necessarily coincident. This brings to mind the question of what size control volume is appropriate for modeling overland flow. For groundwater flow the concept of a REV is well known and is the basis for deriving the flow and transport equations for a porous medium. In an analogous manner, the concept of a Representative Elementary Area (REA) is investigated by Wood et al. (1988) and is defined as: "A critical area at which implicit continuum assumptions can be used without knowledge of the patterns of parameter values, although some knowledge of the underlying distributions may still be necessary."

The study concluded that the notion of a REA does not exist and is strongly influenced by the catchment's topography and to a lesser extent the variability of the soil and rainfall parameters. The non-existence of a REA raises the question of what is a suitable level of discretization for simulating overland flow.

2.4.8 Issues Relating to Simulating Resistance to Overland Flow

The resistance to overland flow is non-linear and strongly dependant on slope, as shown in Figure 2.4, limiting the applicability of a single “effective” friction value; Freeze (1972b) notes that it is the “weakest link in the deterministic chain” of coupled groundwater – surface water modeling. When using a term such as Manning’s *n* to quantify roughness the constituents of the composite roughness are lumped into a single term. Due to scale discrepancy between roughness elements in a natural setting and the scale of the model discretization MODHMS and HydroSphere employee addition parameters to help capture the dynamics of overland flow that are neglected through by using constant, composite roughness. These parameters include surface flow domain porosity, and obstruction and rill storage, for details, see Therrrien et. al. (2005). Though these terms all for better characterizing the effects of the environment’s geometry defining appropriate values to these parameters remains a challenge.

As a direct result of the scale discrepancy, sheet flow is simulated over model element faces resulting from the gradient in surface water depth between at surface water elevations at model nodes; as already discussed, in reality, this mode of overland flow is a rare occurrence as observed by Emmett (1970). The simulation of overland flow occurring only as sheet flow necessitates an overestimation of the roughness parameter. In the numerical model the elemental area is contributing to overland flow where in reality the occurrence of overland flow is highly variable, ranging from concentrated threads to sheet flow. This results in an underestimation of local the surface water velocities (having implications for transport of contaminants) even though the timing and peak of the stream hydrograph may be reproduced by the simulation.

Though it may be possible to simulate the correct stream flow hydrograph by calibration of parameters (e.g., Vieux, 1993; VanderKwaak and Loague, 2001) the parameter values themselves may not be characteristic of the natural properties they are representing but are rather applied as fitting parameters. This is due in large part to the scale discrepancy between the model discretization, the scale the hydrological processes operate on, and the heterogeneity of natural systems. The effects of scale are carried forward when the flow solution is used to drive advection-dispersion models to simulate the transport of contaminants impacting estimates of first arrival times of contaminants and pathogens.

2.4.9 Temporal Discretization Requirements

Then considering groundwater–surface water modeling, the modeler must take into consideration that the velocity of groundwater and surface water are orders of magnitude apart and hence, the temporal discretization becomes a critical component in deterministic, physically-based groundwater–surface water modeling.

In relation to groundwater the velocities of overland and channel surface water flows are much greater. Dunne and Black (1970) estimates overland flow velocities to be on the order of 100 to 500 times that of groundwater velocities. When considering channel flow the difference in flow velocities is even greater. Dunne and Black (1970) report that field measurements show that channel velocities are on the order of 1000 to 2500 ft/s while overland velocities from banks areas is on the order of 20 to 200 ft/s and that subsurface velocities are on the order of 1 ft/hr or less. This necessitates very small timesteps that are orders of magnitude smaller than those typically encountered in groundwater modeling.

Timesteps for routing of overland flows must be very fine during precipitation events, for example, Downer and Ogden (2004) used a maximum time-step of approximately 1 min which was relaxed to a maximum time-step of 1 hour once overland flow was completed. When consideration is given to channel flows even finer timesteps are required to produce stable and accurate numerical solutions to the DWE. With current computing power this limits the length of time that can be simulated to event-based or steady-state applications. The combination of highly refined spatial and temporal discretization effectively precludes rigorous, fully-integrated models from being able to simulate the possible effects of climate change as the simulation length required to make an assessment of the potential impacts is on the order of years to decades to centuries.

VanderKwaak and Loague (2001) report optimistic results using InHM to simulate the groundwater-surface water dynamics of the small, approximately 0.1 km^2, R-5 catchment near Chickasha, Oklahoma, at the event scale. The small cathcment area allowed for a fine spatial discretization to capture the dynamics of the groundwater–surface water interaction. Despite their success, VanderKwaak and Loague (2001) identified that numerical modeling of groundwater-surface water interactions on the event-based scale requires much more detailed information, for example initial soil moisture conditions and the position of the watertable than traditional groundwater modeling.

With the sophisticated, fully-integrated groundwater-surface water models discussed herein detailed meteorological inputs are required in order to properly simulate the timing and routing of surface water, especially when considering event-based modeling. In studies conducted by Singh (1997, 2005) it was found that storm direction and velocity and rainfall duration and intensity had significant impacts of storm runoff, infiltration, and the discharge hydrograph. Thus, unless the dynamics of storm events are captured the overall calibration of the model will suffer as other parameter values will adjusted (in err to compensate for unrepresented storm dynamics) to calibrate the model output to observations. This level of detailed information is not typically available.

These models show that the equations used to describe the movement of water within a closed system can be solved numerically and are based on a deterministic, physically-based approach. However, this does not necessarily imply that the simulated result in an adequate representation of reality. Signh (2002) makes the following point, referring to physically based modeling of overland flow, which is quite relevant with respect to the state of the science (and art) of hydrologic modeling:

> "Our knowledge about the validity of these descriptions and the physical meaning and measurability of the parameters contained in them is woefully inadequate. A close examination of these descriptions suggests that the so-called physical descriptions are not really physical after all, for we cannot determine their parameters beforehand and therefore a lot of fitting is to be undertaken."

2.5 A Coupled Modeling Approach

The coupled modeling approach is not as computationally intensive and shares many of the advantages of a fully-integrated approach however it is not a rigorous in its accounting of the dynamics of groundwater-surface water interaction. There is a variety of groundwater and surface water codes that can be coupled to investigate the interactions of groundwater and surface water. In this investigation, HELP3 linked with HydroSphere by providing the transient recharge boundary condition.

HELP3 uses a water balance approach to abstract the surface (runoff, surface storage, snowmelt), near-surface (interception, evapotranspiration), and vadose zone (soil moisture storage) hydrological process from daily precipitation to simulate the movement of water in a quasi-two-dimensional soil column. A benefit of HELP3 is that it uses empirically derived relationships to reflect the effects of temperature (e.g., reduced infiltration and increased runoff under freezing conditions) and vegetative growth (e.g., increased evapotranspiration, and preferential drainage by roots). The ability for

HELP3 to simulate hydrologicial processes under freezing is particularly powerful as it allows for continuous simulation in geographic areas where freezing temperatures may be seasonally prevalent. A more complete and detailed description of the HELP3 model is proved by Schroeder et. al. (1994b).

The HELP3 code has been extensively tested by its developers (Schroeder et. al., 1994b). HELP3 has been used to simulate percolation through a clay liner overlaying mine tailings and was found to be in good agreement with field observation (e.g., Woyshnet and Yanful, 1995). The HELP3 model also compares well to other more rigorous methods (i.e., application of Richards' Equation) for simulating flow in unsaturated porous media. A study by Fleenor and King (1995) found that the two models were in good agreement under humid conditions but that the HELP3 tended to estimate greater downward fluxes in arid environments. Gogolev (2002) compared recharge estimates generated by HELP3 and VS2DT (Lappala et. al., 1987) for the Waterloo Moraine. VS2DT is a code used to calculate flow and transport in the unsaturated zone using the Richards' Equation. Their study showed that there is no significant gain in determining recharge estimates between the two codes. In all cases the differences between the two estimates was less than 8% except for one case for which the difference was 12.4%. It is also worth nothing that HELP3 performed the 100 year simulation period in less than 15 minutes while for the VS2DT required 2.5-7 hours. In comparing the HELP3 estimate to field estimates (made using the tritium profile method) the HELP3 estimates were within 4% of measured values. Gogolev (2002) concludes the study with the following comment that is worth reiterating, "It is considered that the HELP technology has all necessary qualities to become a core for computational technology for assessing groundwater recharge rates."

The applicability of the HELP3 technology as a tool to assess recharge to large-scale groundwater models has been shown by Jyrkama et. al. (2002). In this application HELP3 was used to estimate a detailed transient, spatially distributed estimate of percolation to be used as the recharge boundary condition for a large (~138 km^2) groundwater flow model. The application of the detailed boundary condition improved model calibration; at observation points the simulated heads were within 0.5 m of observed values, while the best agreement achieved using a uniform recharge boundary condition was 2 m. This study also highlights the importance of the combination that precipitation and temperature play in the distribution, quantity, and timing of recharge

2.5.1 Overland Flow and Runoff Considerations

The rarity of overland flow in humid vegetative conditions is an additional consideration for simplifying the surface water routing component of the numerical model, thereby reducing the computation burden and making the numerical integration of groundwater and surface water systems a manage problem. In HELP3 runoff is accounted for using the SCS method (NRCS, 1986) for the following reasons: it is widely accepted, it is computationally efficient, the required input is generally available, and it can conveniently handle a variety of soil types, land uses and management practices (Schroeder et. al., 1994b). The user chooses the most appropriate Curve Number (CN) that is reflective of soil drainage characteristics, land use classification, and vegetative cover. The CN is applied and updated by HELP3 internally to account for soil moisture conditions.

HELP3 allows for the CN to be adjusted based physical factors such as specified slope and length. Hundreds of runoff estimates were generated for combinations of slope, length, soil type, level of vegetation, and rainfall characteristics using the KINEROS model (Woolhiser et al., 1990), which uses a kinematic wave model used in the evaluation of erosion and sediment transport. Relationships between the KINEROS simulation results were established by regressing CNs to reflect slope, length, and CN based on soil type and vegetation characteristics. Despite these efforts to modify the CN to have a better physical basis, the HELP3 daily runoff estimates cannot be expected to yield accurate runoff for individual storm events. However, since the SCS rainfall-runoff relationship is based on considerable daily field data it is expected that long-term runoff estimates are reasonable and consistent with respect to the daily temporal scale (Schroeder et. al., 1994b).

2.5.2 Limitations of the Coupled Approach

Major weakness in this coupled approach is that there is no feedback from the water table to the soil column. This results in an over-estimation of recharge when the depth to the water table is greater than the length of the soil column and vice-versa. There are no internal checks to ensure agreement with the flux output from HELP3 and the rising and falling of the water table from the groundwater code. This results in a less rigorous accounting the interaction of the groundwater and surface water dynamics as would be simulated in an internally coupled (e.g., MIKE-SHE) and fully-integrated (e.g., HydroSphere) models.

In groundwater models surface water features are typically represented by prescribed Head or Cauchy boundary conditions, which oversimplify the surface water systems their relationship. A coupled approach could be used to provide a greater degree of realism between the tributaries and the groundwater system (e..g, Scibek and Allen, 2007) however the approach may require simulation results to be scaled between the two models.

Each of the process simulated in HELP3 has its own assumptions and limitations which may not be reasonable under all circumstances depending on the application of the model (Schroeder et al., 1994b). Most of the limitations in HELP3 are from the empirical relationships used to describe processes. Though this is not an ideal approach, considerations for the effects of these processes generally enhance the model as more physically based processes are accounted for. These are viewed as short comings that do not limit the application of the model. The major assumptions used in the HELP3 model are summarized:

<u>Freezing Conditions</u>

- precipitation on days for freezing temperatures is assumed to occur as snow
- prediction of frozen soil conditions, snowmelt, and snow accumulation are based on empirical relationships and antecedent air temperatures

<u>Infiltration</u>

- it is assumed that all flow in the soil column is vertical, thus not allowing for interflow or lateral flow on clay lenses
- effect of macropores from rooting channels is incorporated by using empirical relationships to modify the hydraulic conductivity

<u>Evapotranspiration</u>

- based on average annual wind speeds and quarterly humidity values
- humidity is assumed to be 100% on days with precipitation
- the start and end dates of the growing season and maximum rooting depths are constant for the duration of the simulation
- leaf area index is limited to a range from 0 (bare ground) to 5 (excellent stand of grass)

<u>Runoff</u>

- since the SCS rainfall-runoff relationship is based on considerable daily field data it is expected that long-term runoff estimates are reasonable and consistent with respect to the daily temporal scale

- runoff is most strongly dependant on the selection of appropriate CN, introduces bias
- SCS method does not explicitly consider surface topography and vegetative effects that can both enhance or retard overland flow velocities
- temporal distribution of a storm event (i.e., duration and intensity) is lumped to daily timestep thus impacting infiltration and runoff rates. It is expected that runoff is underestimated for individual high intensity, short duration storm events
- the timestep is limited to 1 day, which precludes the model to be used as a tool to evaluate the hydrologic processes occurring at the event-scale (i.e., storm)

Though the coupled approach has its limitations it offers much flexibility and advantages to the computationally intensive fully-integrated approach. The savings for a computationally less intensive approach can be well spent on constructing detailed, continuous, long-term simulations to be used as a tool to asses the potential effects of climate change, deforestation, and urban sprawl. Another key advantage is the ability to incorporate empirical relationships to estimate the effects of vegetative growth, freezing temperatures, snow accumulation, and snow melt.