Evaluating Temporal and Spatial Variations in Recharge and Streamflow Using the Integrated Landscape Hydrology Model (ILHM)

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Projections of climate and land use changes suggest that there will be significant alterations to the hydrology of the Upper Midwest. Forecasting those changes at regional scales requires new modeling tools that take advantage of increases in computational power and the latest GIS and remote-sensing datasets. Because of the need to resolve fine-scale processes, fully coupled numerical simulations of regional watersheds are still prohibitive. Although semi-distributed lumped-parameter models are an alternative, they are often not able to accurately forecast across a broad range of hydrologic conditions such as those associated with climate and land use changes.

We have developed a loosely coupled suite of hydrologic codes called the Integrated Landscape Hydrology Model (ILHM), which combines readily available numerical and energy- and mass-balance modeling codes with novel routines. In this paper, the ILHM is used to predict hydrologic fluxes through a 130 km² portion of the Muskegon River Watershed in northern-lower Michigan. We combine GIS maps of the land cover, soils, and sediments with a variety of gaged and remotely sensed data for this watershed to simulate evapotranspiration, groundwater recharge, and stream discharge from 1990–2004. These estimates are compared to measured stream discharge data to demonstrate the capability of the ILHM to provide reasonable predictions of groundwater recharge with minimal calibration. The results begin to illustrate critical differences in hydrologic processes due to land cover and climate variability, including a demonstration that approximately 75% of precipitation becomes recharge during leaf-off periods while almost no recharge occurs during the growing season.

INTRODUCTION

Land use and climate changes are expected to alter the spatial and temporal distribution of groundwater recharge over the next century [Bourari et al., 1999; Houghton et al.; IPCC, 2001]. In the humid midwest, these changes could have far reaching consequences because recharge maintains groundwater supplies that are used as primary drinking water sources, and is critical to stream ecosystem health as groundwater is the main source of streamflow during dry periods. Despite the clear importance of groundwater recharge, its spatial and temporal distribution is generally poorly understood in humid regions. Many hydrologic modeling studies ignore both spatial and temporal variations in recharge rates, either because limited measurements of critical parameters
are available, or because existing modeling methods are not adequate to accurately evaluate these variations at the scales of interest. Integration of available hydrologic and landscape data can help improve estimates of historic recharge rates, and can then provide the basis for evaluating the range of impacts of anthropogenic alterations of the landscape and climate on future hydrological and ecological conditions.

A range of approaches have been developed to estimate recharge rates based on relatively simple analysis of flows and levels in surface water, the unsaturated zone, and the saturated zone, as reviewed by Scanlon et al. [2002]. These methods include analyzing baseflows or tracer concentrations, developing estimates based on changes in groundwater levels (reviewed by Healy and Cook [2002]), and evaluating recharge through the unsaturated zone with lysimeters or well-instrumented field sites. A variety of empirical models have also been developed to estimate recharge across a range of scales (e.g., Bogena et al. [2005]), which can provide estimates with varying degrees of reliability and spatial extent depending on the types, quality, and density of the input data [Scanlon et al., 2002].

Numerical models provide a powerful framework to integrate different data types for recharge estimation. Such models can be broadly categorized as lumped parameter models or process-based models. Lumped-parameter semi-distributed models, such as SWAT [Arnold et al., 1993] and TOPMODEL [Bevan and Kirkby, 1979] have parameters that can be adjusted to fit measurements but cannot necessarily be independently measured. As a result, such models tend to have difficulty predicting flow in a new system without independent calibration, or projecting likely changes in a currently modeled system due to changes in factors including climate and land cover.

Process-based codes such as MODFLOW for groundwater flow are based on fully distributed parameters such as hydraulic conductivity, which can be independently measured based on laboratory analyses (e.g., Zhao et al. [2003]) or using field evaluations such as pump or slug tests. Unfortunately most groundwater codes are not designed to estimate recharge rates because they do not incorporate important landscape and unsaturated zone processes that are critical to redistribution of precipitation from the soil surface and the vegetation canopy.

To address this limitation, several codes have previously been developed to link MODFLOW or other groundwater codes to landscape or watershed codes that incorporate aspects of the hydrologic cycle beyond groundwater flow. For example, MODFLOW has been linked with SWAT [Sophocleous et al., 2000] and HSPF [Said et al., 2005], which are both lumped-parameter codes. A new Variably Saturated Flow (VSF) package was developed as a MODFLOW module [Thoms et al., 2006] to add unsaturated zone and overland flow processes to groundwater flow simulations, but the data and computational requirements appear to be too great for large watershed simulations based on our analysis of this code for the watershed presented in this paper.

A variety of process-based models have also been developed to simulate fully coupled surface water, and variably saturated subsurface flow. Such codes, which include SUTRA3D [Voss and Provost, 2002], Mike-SHE [DHI, 1993], WASH123D [Yeh and Huang, 2003], MODHMS [HydroGeoLogic Inc., 2003; Panday and Huyakorn, 2004], and InHM [VanderKwaak, 1999; VanderKwaak and Loague, 2001], provide powerful tools to examine complex interactions between flow and transport across the range of natural conditions observed in the surface and subsurface. Unfortunately, the data requirements and significant computational demands have generally limited the use of these codes to simulate flows through fairly small domains.

In this paper, we present a new Integrated Landscape Hydrology Model (ILHM) to integrate widely available hydrologic and landscape data in a synergistic and computationally efficient manner to assess temporal and spatial changes in important hydrologic processes. Since the focus of this monograph is data integration in hydrology, we begin by describing the watershed that we chose for testing and development of the code along with the available hydrologic and landscape data used in this simulation. This is followed by a detailed description of the model development and results.

METHODS

Study Region: Cedar Creek Watershed

The Cedar Creek Watershed, in southwestern Michigan (Figure 1), was chosen as a site to test the ILHM because it is one of our main field sites in an ongoing ecohydrological monitoring and modeling study. Cedar Creek flows through the lower half of the Muskegon River watershed (7,052 km²), where urbanization of previously agricultural and forested landscapes is projected to increase runoff volumes and the associated solute transport over the next 35 years based on an empirical model [Tung et al., 2005]. The spatial distribution of land uses within the Cedar Creek Watershed facilitates evaluation of differences in recharge associated with land cover types because the upstream portion of this area is dominated by agriculture while the downstream portion is predominantly forested (Figure 2a). The quaternary geology ranges from medium and coarse-textured glacial tills that drape the northern watershed, to glacial outwash and lacustrine sand and gravel in the central and southern watershed (Figure 2b).

The groundwater source area, which we call a groundwatershed, of Cedar Creek was delineated using a two-layer ground-
water model of the region encompassing the Muskegon River Watershed (Figure 1). The groundwatershed (~130 km$^2$) was used in addition to the surface watershed (~100 km$^2$) for this study because regional modeling of the Grand Traverse Bay Watershed in Michigan by Boutt et al. [2001] indicated that surface- and groundwatersheds can differ significantly. The regional Muskegon River groundwater model was developed by expanding the watershed boundaries to significant hydrologic features (i.e., the next large stream or lake beyond the surface watershed) to avoid this issue at regional scales (Figure 1). The groundwatershed boundary does fluctuate somewhat with both seasonal and long term climatic variations, but for simplicity in this study we have defined the groundwatershed using the steady-state model.

**DATA COLLECTION AND ANALYSIS**

Before constructing the groundwater model for the expanded Muskegon River Watershed (MRW) region to define the Cedar Creek groundwatershed, we assembled the available landscape, hydrology, and climate data for the region into a geodatabase. In many parts of the world, the types of data used for this analysis are commonly available as a free download from internet sites. However, supplementary data such as flows and water levels beyond those available from the US Geological Survey will often need to be collected for model calibration or optimization.

**Hydrologic Data**

Two pressure transducers installed in Cedar Creek recorded stream stage at hourly to sub-hourly intervals [Wiley and Richards, unpublished data] from mid to late 2002 through 2004. These surface water levels provide critical information for this study. One transducer was installed in the northern, agricultural portion of the watershed, while the other was installed in forested land near the watershed outlet (See Figure 1). Stream discharge measurements [Wiley and Richards, unpublished data] were used to construct rating curves between stage and discharge. The stage discharge relationships were developed between measured streamflows and concurrent water levels from the transducers. For the upper watershed site 19 stage discharge pairs were used, while 23 pairs were used for the lower watershed site.

Groundwater levels for this region were collected from the Michigan Department of Environmental Quality (MDEQ) residential well database, as no monitoring well data are available in this region except at our surface water-groundwater interaction site adjacent to Cedar Creek. Unfortunately, the wells at this site are too close to the stream to provide useful

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Figure 1. On the left is a map for the Cedar Creek watershed (shaded) along with the groundwater contributing area to Cedar Creek (dashed outline). Also displayed on this map are locations of two stream gages, nine measurement cross sections, and residential drinking water wells located within the watershed. On the right, a map of the lower peninsula of the state of Michigan shows the Cedar Creek watershed within the greater Muskegon River watershed. The boundary of a regional groundwater model of the Muskegon River is shown in bold on the state map. The precipitation and climate gage locations are Hesperia (H), and Fremont (F).
information about groundwater levels for this watershed-scale model testing. Observations were available from 99 wells installed across the watershed during the simulation period in the MDEQ database. For each well, one static water level measurement was taken by the well driller at the time of installation. The static water level measurements were used in a preliminary calibration of an early version of the Cedar Creek groundwater model, but there is a significant amount of error associated with these water level measurements as a variety of methods were used by well drillers to identify the location of the well, the elevation of the ground surface, and the depth to water. Residential wells were not included in the model as extraction wells because most of the extracted water is assumed to return via septic systems and the remainder is assumed to be a very small component of the water budget for this region. There are no known irrigated agricultural areas within the Cedar Creek watershed.

**GEOLOGIC, LANDSCAPE, AND REMOTE SENSING DATA**

We established a GIS database for the Cedar Creek region with topography, land use, hydrography, and hydrogeology characteristics. These GIS datasets were compiled from the Michigan Geographic Data Library, established by the Michigan Department of Environmental Quality. These datasets are assumed to be static for the purposes of this analysis.

Land surface elevations (see Figure 2d), which were defined based on the National Elevation Dataset 26.5 m digital elevation model (DEM), were used for a range of model inputs. Flow direction and gradients for the overland flow and near surface soil moisture redistribution modules were calculated from the full-resolution DEM and then upscaled. This 4x upscaled DEM was used to set the land surface elevation for the soil water balance model. The drainage network and lake boundaries were defined based on the Michigan Framework GIS dataset, with stream crossing elevations manually extracted from a digitized version of the USGS 1:24000 topography quadrangles. The watershed does not contain any large lakes that are connected to the stream system, thus lakes are not separately considered in the groundwater portion of the integrated model.

The land cover distribution across the Cedar Creek model area (Figure 2) is taken from the Integrated Forest Monitoring Assessment and Prescription (IFMAP) coverage, which is a statewide digital land cover map with 30 m resolution derived from 1997–2000 LANDSAT data [MDNR, 2001]. This watershed is dominated by forested/openland/wetland land covers (60%), while 36% of the area is agricultural and the remaining 4% is urban. The IFMAP coverage provides inputs that are used in calculating evapotranspiration and overland flow. Land cover types are associated with transpiration estimates through a variety of terms including stomatal conductance, canopy height, and root depth, and with evaporation estimates through changes in canopy interception, wind speed, and interception of incoming radiation. Overland flow is also associated with land cover through Manning’s roughness coefficients. The details of these connections are described in the modeling sub-sections below.
Hydrogeologic zones were parameterized according to a Quaternary Geology coverage of Farrand and Bell [1982] for the groundwater model along with the Soil Survey Geographic (SSURGO) database (see Figure 2), which was then mapped into saturated and unsaturated zone parameters according to lookup tables based on literature values. The geometry of the aquifer base was interpolated between measured bedrock elevations by de-clustered and polynomial-detrended simple kriging of the elevations of the drift/bedrock contact from oil and gas wells across the entire expanded Muskegon River model domain. Initial hydraulic conductivity values were assigned to the geologic zones based on an optimization of these parameters for the nearby Grand Traverse Bay watershed that has the same geologic zones [Boutt et al., 2001].

GIS grids of leaf area index (LAI), the ratio of one-sided green leaf area to ground area [Myneni et al., 2002], were also used in calculations of potential evapotranspiration (PET), canopy interception, and solar radiation interception. For this study, we used remotely sensed LAI measurements from NASA’s Moderate Resolution Imaging Spectroradiometer (MODIS) eight-day averaged product. Spatially averaged LAI for both forest and agricultural land-use types are plotted for 2003 and 2004 in Figure 3a. In cases where these data were not available (i.e., prior to 2000), we average all LAI grids for each Julian day and apply these multi-year averages to the earlier periods. This will have little effect on our results because we are only comparing simulated and observed flows for mid 2001 through 2004 when all datasets are available. However, it is important to spin up the model using realistic data inputs because we found that it takes between two and three years before the model results are independent of the starting conditions.

CLIMATE DATA

Precipitation data was obtained from the NOAA gage at Hesperia, MI approximately 20 km NNW from the center of the Cedar Creek watershed (see locator map in Figure 1). This gage was chosen because lake effect precipitation is an important meteorological phenomenon in this area, and this gage lies at relatively the same distance from the Lake Michigan shoreline as the Cedar Creek watershed. NOAA data (shown in Figure 3b) included hourly precipitation totals, as well as daily measurements of new snowfall and snow pack depth.

Other climate data, including hourly temperature, relative humidity, wind speed, and incoming solar radiation (Figures 3c–f), were extracted from the Fremont, MI station of the Michigan Automated Weather Network (MAWN) (see Figure 1 for location). This climate network is operated by the Michigan State University Extension, the Michigan Agricultural Experiment Station and the Michigan Department of Agriculture. Since the MAWN data did not exist prior to 1996, from 1990–1995 we used the Julian-day average of the available data.

COMPONENTS OF THE INTEGRATED LANDSCAPE HYDROLOGY MODEL (ILHM)

Figure 4 illustrates our conceptual model of the most important hydrologic processes in the Cedar Creek watershed, and diagrams the linkages between input datasets, ILHM modules, and model outputs. As mentioned earlier, this version of ILHM was developed by linking a novel landscape water balance model with a simple linear -delay unsaturated zone model and MODFLOW-2000 [Harbaugh et al., 2000], the most commonly used groundwater flow code. The landscape and near-surface portion of the ILHM combines several existing codes with a set of new modules, in order to speed development and to incorporate the full range of hydrologic processes. The canopy water balance model
is based on equations published in Chen et al. [2005]. The surface hydrology model, including infiltration and runoff routing, are modified from the Distributed model for Runoff, Evapotranspiration, and Antecedent soil Moisture (DREAM) model by Manfreda et al. [2005]. The snow pack is simulated using the UEB Snow Model by Tarboton and Luce [1996]. Soil moisture accounting along with near-surface flows are handled by a set of codes we developed based on common unsaturated zone flow modeling methods.

The ILHM suite calculates each term in the full water balance equation:

\[ \Delta S = P - T - E - Pc + Tr - Ex - R \]  

where \( \Delta S \) is the change in soil moisture storage in the biologically active soil zone, \( P \) is watershed available precipitation, \( T \) is transpiration, \( E \) is evaporation, \( Pc \) is deep percolation beneath the biologically active soil zone, \( Tr \) is lateral near-surface unsaturated flow called throughflow, and \( Ex \) is the exfiltration from each cell, and \( R \) is precipitation excess runoff. For Equation 1 and the detailed water balance equations presented in Appendix A, terms are in units of meters per unit time unless otherwise specified.

The landscape portion of our model sequentially calculates the water balance along the paths water takes as it is redistributed from precipitation to various subsurface and surface pathways. Incoming rainfall is first subjected to canopy interception, while snow is routed directly to the snow pack model. Next, canopy throughfall and snowmelt are applied to the soil surface. These new inputs are then combined with any water stored in surface depressions and allowed to infiltrate into the soil. Any excess water at this point enters surface depression storage up to the available capacity.

Infiltrated moisture is added to the existing surficial soil layer budget, where it can then percolate downward under the influence of hydraulic gradients. Any moisture within the
first soil layer is then available for evaporation, along with any transpiration that may occur in any of the biologically-active soil layers. Subsurface lateral throughflow is then calculated, which may cause moisture in down-gradient cells to exceed saturation. At this point, moisture in the lowest biologically-active soil layer may then percolate into the sediments beneath, where it becomes deep percolation. Remaining moisture in excess of saturation is exfiltrated back toward the surface where it also enters depression storage.

Deep percolation is then delayed as a linear function of the thickness of the unsaturated zone, which is estimated based on a steady-state run of the regional MRW groundwater model. The delayed percolation then becomes recharge to the three-dimensional transient groundwater flow model when it crosses the water table. Water stored in surface depressions is then subjected to direct evaporation. If depression storage capacity is exceeded, the excess water becomes surface runoff. Baseflow discharge from the groundwater model is then combined with the surface runoff and throughflow to produce the complete simulated stream hydrograph.

The landscape hydrology components of the model for Cedar Creek are simulated with a 177×153 grid at 106.3 m resolution. As shown in Figure 1, while the watersheds and groundwatersheds overlap for most of the modeling domain, some locations contribute only surface water or groundwater to Cedar Creek. To account for this, the landscape hydrology model is run for the entire domain while the unsaturated zone model only allows groundwater recharge in active cells of the saturated groundwater model, and the stream routing module only includes areas within the surface-watershed of Cedar Creek.

The following sections describe the details of each set of processes simulated in the ILHM, with specific assumptions that were made for the Cedar Creek case. The mathematical descriptions and details about parameters are included in Appendix A.

Precipitation and Snowmelt

Watershed available precipitation \( P \) is the sum of liquid rainfall and snowmelt. From late December through mid March, precipitation falls predominantly as snow in the Cedar Creek watershed. To model the storage and release of snow we used the UEB Snowmelt Model by Tarboton and Luce [1996], which is an explicit energy and water balance model designed to track three state variables: snow water equivalent, energy deficit (i.e., how much energy would be required to return the snow pack and soil layer to the 0 degree C reference condition), and the snow surface age. This model is computationally efficient because it assumes no temperature gradient within the snow pack and the layer of soil with which it interacts. For ease of integration with the rest of the ILHM model suite, we ported the FORTRAN version of this snowmelt code into MATLAB.

The full UEB model requires air temperature, wind speed, relative humidity, and solar insolation. The adjustable parameters in this model component include the density of the snowpack, the thermal conductance of the snow, the liquid water holding capacity of the snowpack, and the depth of soil with which the snow thermally interacts. Preliminary calibration to snow depth data from a single year of record provided the parameter values shown in Table 1.

The current version of the ILHM only runs the UEB model if either the air temperature during a precipitation event is below freezing, or the snow water equivalent of the snowpack is greater than 0.01 mm. Any water remaining in the snowpack below this amount of moisture is then applied to the surface as additional snowmelt. All available snowmelt calculated by the UEB model is then added to any liquid precipitation for each time step to become watershed available precipitation \( P \).

Evaporation and Transpiration

The evaporation \( E \) and transpiration \( T \) terms of the water balance equation first require calculation of potential evaporation and transpiration. All evaporation and transpiration potentials, (canopy, depression, soil, and transpiration) are calculated using the modified Penman-Monteith equation [Monteith, 1965] presented by Chen et al. [2005] and shown as Equation A1. For each separate potential, the aerodynamic resistance and resistance to vapor transport terms are modified as described in the Appendix.

Evaporation and transpiration rates vary temporally according to land cover types through 8-day LAI scenes, a stomatal conductance coverage, and soil texture. For the Cedar Creek watershed, which is small and has relatively brief surface water residence times, we assume that there is no open water evaporation. Work is ongoing to explicitly model this component of evaporation.

Incoming rainfall is first subjected to interception up to the water holding capacity of the canopy, which is related to LAI in the cell. Water storage in the canopy is simulated as a

\[
\text{Parameter} \quad \text{Units} \quad \text{Value} \\
\hline
\text{Snow Density} & \text{kg m}^{-3} & 200 \\
\text{Liquid Holding Capacity} & - & 0.15 \\
\text{Thermally Active Soil Depth} & \text{m} & 1.0 \\
\text{Snow Thermal Conductance} & \text{m hr}^{-1} & 0.2 \\
\hline
\]
separate storage layer, with losses only from evaporation. We assume that the largely deciduous canopy does not intercept snowfall. The canopy is also assumed to intercept a portion of the incoming solar radiation based on LAI.

The evaporation of moisture in depression storage is calculated after any infiltration and exfiltration (described below) in a given time step have occurred. Total depression storage capacity is determined by land use, soil texture, and slope class; a tabular reference of storage capacities can be found in Manfreda et al. [2005]. Depression evaporation occurs at the potential rate until depression storage is depleted. Evaporation from depressions and directly from soil is allowed only from the proportion of soil that is exposed to solar radiation, thus assuming no soil evaporation from the portion shaded by canopy or covered with snow or ice.

Direct soil evaporation is allowed only from the first soil layer, which is a reasonable assumption in this relatively humid region. Soil evaporation occurs at the lesser of calculated potential rate or the soil exfiltration depth (discussed further in the Appendix following Equation A16). We are exploring alternative strategies for calculating evaporation using the model of Ritchie [1972]. The total transpiration in each cell $T$ is calculated to be the sum of the root water uptake from each biologically active soil layer. We assume that transpiration only occurs above a dormant threshold temperature, which was chosen to be 40 degrees F for this study. Stomatal conductance is also assumed to be constant for this case, although we plan to incorporate variations due to changes in temperature, carbon dioxide concentrations, and soil moisture as described in Chen et al. [2005].

**Infiltration, Percolation, Throughflow, and Exfiltration**

For this study we assume the biologically active soil can be described by two layers, with a total thickness calculated according to Equation A14 as the depth above which 90% of the root mass lies. The first soil layer, from which evaporation occurs and that controls infiltration capacity, is on the order of several centimeters thick. Infiltration capacity is calculated as the greater of either the soil-texture dependent saturated infiltration capacity, $i_{sat}$, or the first layer moisture deficit from saturation. We chose the maximum infiltration rate to be $(2 \cdot i_{sat})$ which determines the choice of the first soil layer thickness. This formulation produces similar results to empirical infiltration rate descriptions, and has the advantage that it does not require storm event tracking or single storm event modeling.

Here we assume that $i_{sat}$ can take either its nominal soil-texture dependent value taken from literature values (see Appendix), or $i_{sat} = 0$ if the soil is frozen, which we only allow to occur in agricultural soils based on Schaeztl and Tomczak [2001]. The soil is assumed to be frozen if its temperature is below $-0.25^\circ$C, measured by the MAWN station in Fremont at a depth of 10 cm.

In addition, we assume that no infiltration occurs in cells classified as permanent water features. Although the water features in this watershed often cover only a small portion of each land-use cell, our assumption may nevertheless be fairly realistic. The true physical system process is more accurately described as rapid percolation to a shallow water table followed by equally rapid rise and subsequent relaxation of the water table. The effect is a temporary increase in groundwater discharge that generally does not modify what is more traditionally called stream baseflow. Our stream gage data indicate that the characteristic time of this response may be on the order of twice the surface runoff response, or perhaps 5–7 days. This does not allow for significant losses due to evaportranspiration, thus the combined increase in stream discharge due to percolation to the near-stream water table and direct overland flow would be nearly equal to that expected from an assumption of zero infiltration. Streamflows in our model would thus be expected to peak higher and return to baseflow levels more rapidly than observed.

Throughflow, defined here as lateral subsurface flow within the biologically active soil zone is calculated using a simplified Richards equation model. For a full development of the Richards equation, see [Hillel, 1980]. For purposes of computational efficiency, and in order to assure that the subsurface redistribution of moisture occurs in only one dimension, we assume that flow only occurs parallel to the dip of the slope and thus cannot flow uphill on the ~100 meter scale of our model cells. For environments where these assumptions may be invalid, alternate two-dimensional formulations could be substituted for this ILHM module given adequate computational resources. The van Genuchten model was used to calculate all soil-moisture dependent properties [van Genuchten, 1980] with parameter values given in the Appendix. The downgradient cell is determined via the D8 flow direction function [ESRI, 2003]. However, each cell can have more than one upgradient cell, thus throughflow is the summation of the shallow subsurface flow out of all upgradient cells.

** Unsaturated Zone Delay **

Deep percolation beneath the biologically-active soil layer is delayed prior to becoming recharged as a linear function of the depth to the water table. The slope of this delay in units of days/meter, was determined from wells installed in the nearby Grand Traverse Bay Watershed, and was fixed at 2.5 for this study. It is important to note that this would not be the same as a solute transport time through the unsaturated
zone. Delayed deep percolation then becomes recharge once it reaches the water table. The depth of the water table is assumed temporally invariant for the current version of the unsaturated zone delay module for ease of implementation. By fixing the depth of the water table for this purpose, we do not account for seasonal or trending differences in water travel times through the unsaturated zone. The seasonal differences in the depth of the water table are small (typically <1 meter) relative to average depths to water over most of the model domain. Locations very close to surface water features are an exception, but these areas comprise a small fraction of the total watershed area and thus will not significantly affect the dynamics of modeled recharge.

**Groundwater Model**

The groundwater model for the expanded MRW was developed using a suite of MATLAB utilities that we developed to create input files from GIS layers for MODFLOW. A regular grid of 1798 x 1865 cells (106.3 meters on a side) was used so that each cell directly overlies 16 Digital Elevation Model (DEM) cells. The vertical domain of this saturated zone model was then subdivided into two layers with approximately equal saturated thicknesses based on the simulated water table in a single layer model. Automated parameter estimation routines were applied to an early version of the Cedar Creek groundwater and soil balance model to estimate hydraulic conductivity values for aquifer sediments in geologic zones parameterized using a digital map created from Farrand and Bell [1982].

The Cedar Creek model is a single layer with dimensions of 184 x 162 and cell size of 100 meters. In this case, a single vertical layer provided an adequate description of flows through the Cedar Creek watershed, because it does not have significant vertical relief, extensive low-permeability subsurface layers, high-capacity pumping wells, or other features that tend to induce significant vertical head gradients. Groundwater discharge to streams is calculated with the Stream Flow Routing (SFR) package in MODFLOW that routes water via the kinematic wave equation [Prudic et al., 2004].

**Runoff and Stream Routing**

Surface runoff is routed to the streams using an approach modified from that presented by Manfreda et al. [2003]. In this version of the code, we assume that runoff cannot re-infiltrate once it is generated. It is routed overland and through streams according to the D8 flowdirection algorithm in ARC [ESRI, 2003] with runoff times given by the velocities in each cell along the flowpath. Runoff is assumed to travel overland at a velocity given by the Kerby time of concentration equation [Kerby, 1959]. Once the runoff enters the stream channel its velocity is calculated using Manning’s Equation.

For this study, the hydraulic radius, \( r \), for Manning’s Equation was determined as a function of discharge using low-flow channel geometry measurements in and around the Cedar Creek watershed along with geometries reported by the USGS for their stream gages. Wetted perimeter was assumed equal to \( 2 \times \text{depth} + \text{width} \), while area was simply \( \text{depth} \times \text{width} \). A power law fit to these data produced the empirical relationship for this watershed:

\[
 r = 0.9046 \cdot Q^{0.283}
\]

with a correlation coefficient \( R^2 \) of 0.77 (see Figure 5). \( Q \) in the above equation is the measured stream discharge in m\(^3\)/s.

While streamflow velocity can be dynamically calculated, for simplicity we have assumed a temporally constant \( v_{\text{stream}} \) for each stream cell. To calculate \( v_{\text{stream}} \), we assume that each cell in the Cedar Creek watershed contributes a unit of runoff which is then routed to the gages using ARC’s flowaccumulation function. To rescale the output to match a typical discharge event in a stream cell \( (Q_i) \), we multiply the measured \( Q \) at the outlet by the ratio of the flowaccumulation value in each cell \( (i,j) \) to that of the outlet:

\[
 Q_i = Q_{\text{outlet}} \left( \frac{\text{flowaccumulation}_i}{\text{flowaccumulation}_{\text{outlet}}} \right)
\]

Once \( v_{\text{stream}} \) has been calculated, the flow time from each grid cell to the outlet(s) can be calculated using ARC’s flow-length function, which calculates the cost-weighted distance.

![Figure 5](image-url)
from each cell to the outlet. Here the weighting is the inverse of the velocity in seconds/meter. ARC then multiplies this "cost" by the distance traveled through each cell along the entire flowpath and outputs to the travel time to the outlet. The travel time grid is then used to transform the precipitation excess grid into a runoff hydrograph.

RESULTS

The results of the ILHM simulation for the Cedar Creek watershed are discussed in the context of the broader goals of the code, which are the prediction of temporal and spatial variations in recharge with very little direct calibration of model parameters using readily available remote-sensed and ground-based data sources. All parameters for this prediction were based on literature values (Tables 2 and A1–A3), except for the UEB model parameters (Table 1) as well as two hydraulic conductivity values and one unsaturated zone delay parameter that were calibrated using a very early version of the model (Table 2).

A plot of simulated versus observed heads (Figure 6) across this region shows a reasonable degree of agreement given the measurement uncertainty. Figure 6 shows no trending bias between simulated and observed heads, though a slight high-side bias is present at observed heads lower than approximately 200 meters. We would expect a higher degree of correlation between simulated and observed water levels in regions where pressure transducer data are available from wells, or if a parameter optimization were to be performed for this ILHM simulation.

Detailed evaluation of modeled flows is hampered by the lack of long duration stream gages with stable channels in the basin. All observed stream discharges were collected via established methods, however the rating curves for the two stream gages are currently inadequate to account for temporal adjustments of the channel geometry after flood events. As is commonly the case, we have few high flow measurements, which limits the accuracy of our flows calculated from the rating curve during large floods. In addition, flows in the Lower Cedar gage from January through March appear to suffer from ice-induced over-pressurization not observed at the Upper Cedar gage.

Despite almost no calibration of parameters in the near-surface components of the ILHM code, the model provided a reasonable prediction of observed flows (Figure 7) for the two available gage sites in this 100 square kilometer watershed (Figure 1) during the fall and winter months. The ILHM also provided reasonable predictions of baseflow for this watershed system during the entire year. Because this prediction is based almost entirely on a set of widely available meteorological inputs and GIS datasets combined with literature parameter values, the code appears suitable for directly simulating streamflows in ungaged basins.

The close agreement between observed and simulated baseflow levels also suggests that the model is providing reasonable predictions of recharge, which provides baseflow in these streams during low flow periods. During May and October 2003, and January–March 2004, the ILHM-simulated total flows typically agreed with gaged values within 10% (when the lower Cedar gage was not affected by ice cover). Baseflow levels in the smaller upper Cedar catchment proved highly sensitive to hydraulic conductivity in the outwash sand/gravel zone (Figure 2a). The conductivity values presented in Table 2 should not be viewed as a fully calibrated parameter set, as optimization of the total streamflow simulation is the subject of ongoing evaluation.

Despite the limitations imposed by some parts the stream gage data, the values from April through December of 2003 are reasonable for quantitative flow comparisons. During lower ET periods from April through December of 2003 are reasonable for quantitative flow comparisons. During lower ET periods from April through December, the ILHM-simulated total fluxes are approximately 20% higher than calculated from the flow gages installed at both the Upper and Lower Cedar gage sites. ILHM-simulated total fluxes during higher ET periods from June through October are much greater than observed, also likely due to an incomplete description of ET processes in this version of the ILHM code that is the subject of ongoing development.

Table 2. List of calibrated parameters for the unsaturated and saturated groundwater models.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Units</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Outwash conductivity</td>
<td>m d⁻¹</td>
<td>11.0</td>
</tr>
<tr>
<td>Till conductivity</td>
<td>m d⁻¹</td>
<td>4.4</td>
</tr>
<tr>
<td>Unsaturated zone delay</td>
<td>d m⁻¹</td>
<td>2.5</td>
</tr>
</tbody>
</table>
A scatter plot of simulated versus observed flows at the two gages on a log-log scale illustrates that most of the moderate to high flows in the system are reasonably described by the model (Figure 8). There is a larger degree of mismatch in the Upper Cedar Creek site, as illustrated by a significant amount of scatter about the 1:1 line. Simulated peak discharge values are similar to those that have been measured, however the simulated discharge peaks are narrower than observed. This temporal offset at near-peak discharge, seen in Figure 7b, appears in Figure 8 as a tendency toward low simulated flows relative to observed values. These narrow simulated peaks are largely related to the simple nature of our stream routing package, and the unidirectional linkage between groundwater and surface water processes in this version of the ILHM code. The assumption of a temporally constant $v_{\text{stream}}$ results in average flow velocities lower than those that would be expected, with the effect that simulated peaks would be even sharper if $v_{\text{stream}}$ were a function of discharge. However, there are several known flow damping mechanisms, including flow through wetlands and bank exchange, which are not yet represented in this version of the code.

A map of the simulated average annual recharge across the watershed implies that agricultural areas may have higher recharge than forested areas according to this simulation (Figure 9). Several inter-related factors combine to account for the simulated differences. Agricultural areas experience less canopy interception than forested areas due to lower LAI values. Although this increases the infiltration into agricultural soils, a resulting increase in transpiration tends to narrow the difference in recharge between these land use types. Our model may also under-represent soil evaporation in agricultural soils because it does not incorporate solar...
heating of the shallow soil layer. This simulated difference is the subject of future evaluation across the much larger MRW where more flow data are available.

The blocky nature of the simulated recharge in this map is mainly due to the large (1 km²) LAI cells. The effect of these coarse cells is to decrease forest LAI and increase agricultural LAI in regions with mixed land-uses. We plan to resolve this issue through downscaling the LAI information by assuming that the measured LAI value is a linear combination of the LAI of forest and agricultural land-uses represented by the much higher resolution IFMAP dataset. Thus unique “agricultural LAI” and “forested LAI” values can be approximated at the resolution of the IFMAP data constrained by the total measured LAI from MODIS.

Areas with low hydraulic conductivity soils experience reduced recharge (Figure 9), such as portions of the upper watershed to the south side of the stream where loams and silty loams are common (compare with the soils map in Figure 2c). In contrast, recharge can be greatly enhanced in internally drained regions. In this simulation we deactivated the runoff mechanism in internally drained areas, thus potentially increasing infiltration and shallow subsurface flow. This may be very important in areas with moderate to low-conductivity soils, but the internally drained areas in this watershed tended to also be sandy so the effect is only localized.

The Cedar Creek region experiences very little runoff from upland areas, which is consistent with the simulated map of precipitation excess (Figure 10). Nearly all the simulated precipitation excess in this watershed occurs in cells that are classified as “water” because there is no transpiration or percolation from those cells. The only cells with any significant precipitation excess that are not classified as water are in a region of lower conductivity sediments in the upper watershed. Despite a simple description of runoff processes, we do not expect significant runoff in most of the sediments across this watershed due to high infiltration capacities and saturated hydraulic conductivities. There is also very little simulated subsurface redistribution, or throughflow, throughout most of the watershed due to the highly conductive soils and relatively gentle topography. This process is most active in areas with steep slopes or water tables very near the surface.

The simulation provides evidence for a strong seasonality in recharge rates for the Cedar Creek watershed. The temporal variations in simulated deep percolation are shown in Figure 11. From September through March in the four illustrated years, the model predicts that approximately 70–80% of watershed available precipitation will percolate into the deep aquifer sediments where it eventually recharges groundwater. In contrast, the simulations show virtually no

Figure 9. ILHM-simulated average annual recharge for the Cedar Creek watershed. Annual total precipitation averaged approximately 83 cm during the period of simulation.

Figure 10. ILHM-simulated average annual precipitation excess runoff for the Cedar Creek watershed. Annual total precipitation averaged approximately 83 cm during the period of simulation.
deep percolation over the growing season from May through September for the same years, which is consistent with the statistical findings of Jayawickreme and Hyndman [2007]. This simulation indicates that agricultural areas have more recharge in the fall months than forested areas, while the opposite occurs with higher relative forest recharge in the spring months. This is reasonable as forests tend to have less extensive frozen soils during snowmelt periods [Schaetzl and Tomczak, 2001], and agricultural LAI often begins to decline earlier in the year than in forested areas. Although coniferous forests may transpire year round, they represent only a small percentage of the forested areas in this study region.

Temporal variation in evaporation and transpiration are clearly the causes of most of the simulated variations in deep percolation because these are the primary loss processes. As Figure 12 illustrates, evaporation is generally a much smaller component of water loss in this watershed than transpiration, and this component is larger in forested land relative to agricultural land due to much higher forest canopy interception. Transpiration shows a stronger seasonal trend than evaporation, as it depends more strongly on LAI. Total agricultural transpiration is greater than that of forested areas despite much greater potential transpiration in forested areas. Agricultural areas experience less canopy interception than forests, and thus greater infiltration and higher average soil moisture. As a result, agricultural areas tend to transpire closer to their potential rate than forested areas. Unexpectedly, agricultural transpiration also rises in the spring more quickly than that of forested areas according to these model results due to the similarity in LAI values during early spring and higher stomatal conductance values for agricultural areas relative to forests. As the LAI of forests increases in the late spring, the transpiration in these areas becomes larger than that of agricultural areas, until they reach approximate equality in late June that continues through the rest of the summer. Also during the summer, soil moisture levels reach their lowest point and often approach the permanent wilting point. As deep percolation cannot occur until the field capacity of the soils is reached, most of the water that does infiltrate the soil is transpired. Thus deep percolation is almost non-existent during summer months according to these simulations.

DISCUSSION AND CONCLUSIONS

We present the development and testing of a new suite of loosely coupled process-based codes that we call the Integrated Landscape Hydrology Model (ILHM). This modeling framework has several advantages over existing coupled hydrology codes. It can simulate much larger domains than fully coupled process-based codes, with fewer data requirements. In addition, the ILHM accounts for the processes and mass balance in a more rigorous manner than semi-distributed codes, which tend to lump or oversimplify important watershed processes and use parameters that cannot be independently measured. The ILHM also facilitates model development via direct input of readily available GIS data, in contrast to the impractical level of manual data input.
required for large domains from some existing process-based models such as Mike-SHE. Finally, ILHM is well-suited for forecasting purposes because it allows forcing data and component process models to be interchangeable; thus a model developed and calibrated with current data can be rapidly converted to a forecast simulation by adding the appropriate component process code.

This new modeling framework was designed to make development of models for large domains as simple as possible, while maintaining a rigorous fluid mass balance based on the primary processes that drive water movement over the landscape and through the subsurface. The approach is computationally efficient because it allows some processes to be simulated based on full numerical models while others can be described by simpler and thus faster water- and energy-balance approaches. Due to the loose-coupling framework, individual components can also be simulated at a variety of spatial and temporal scales appropriate to the individual processes. This framework also allows more rigorous simulation modules to be used in place of a simpler routine in cases where the additional computational burden provides necessary improvement in the model predictions. Alternatively, in cases where enough data exist to adequately describe a particular process, the data can be used in lieu of that process simulation module.

As currently configured, ILHM is designed to simulate flows through regions with connected surface water and groundwater regimes such as Cedar Creek. This test watershed has a sub-humid and temperate climate, with flow through a glacio-fluvial aquifer, largely covered with deciduous forests, agricultural land, and small percentage of urban cover. Thus the code is expected to provide reasonable predictions for similar environments in the sub-humid Midwest. The general processes are the same in arid and montane regions; however alternate modules would likely provide more accurate simulations in such cases. In particular, some high-relief environments may require a full two-dimensional representation of overland flow, including depth-dependent velocities for sheet or rill flow. Areas with large proportions of urban land uses will require additional modifications, especially when engineered storm water systems have a significant effect on the hydrograph shape after a storm event.

In the Cedar Creek watershed, precipitation excess runoff routing and subsurface moisture redistribution are both largely inactive over most of the modeled domain and time-frame. As a result, the simulation results are similar even if these modules are not active for upland areas. Therefore, further testing in domains where these processes are responsible for a significantly larger percentage of the flow in a river system will be needed for these modules. The current unsaturated zone module is a very simple representation of hydrologic processes, thus we will explore the use of direct solution methods ranging from the Green-Ampt model through the full Richards equations. Additionally, using MODFLOW or any finite difference scheme has the disadvantage of requiring somewhat cumbersome rectangular grids that limit cell refinement at regional scales. However, ILHM can easily be altered to interface with a finite element code capable of representing and accounting for groundwater discharges to streams.
The ILHM was tested in the Cedar Creek watershed because of the need for high-resolution flow simulations that provide the interface between land use change models and ecohydrology models in the near future. This first evaluation of the ILHM modeling framework demonstrated that these codes can reasonably predict groundwater recharge and streamflow through a 130 square kilometer watershed with very little calibration using readily available data. The simulation represented overall basin recharge accurately, but it appears to have slightly overestimated recharge in agricultural areas. This is likely due to an inadequate representation of soil evaporation that will be addressed in a future version of the ILHM. The simulated hydrograph peaks are too narrow and decline more rapidly than is observed because the current surface water/groundwater linkage cannot represent bank storage and release processes, nor can the unidirectional coupling between surface water and groundwater fully represent near-stream processes at our chosen spatial scales. Nevertheless, the recharge and streamflow predictions provide reasonable descriptions of system behavior and will be further refined in future versions of the ILHM applied to much larger domains.

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APPENDIX A: ILHM MODEL DEVELOPMENT

Evaporation and Transpiration

The basis for our potential evaporation and transpiration calculations is the modified Penman-Monteith equation [Monteith, 1965] presented by Chen et al. [2005]:

\[ PE,PT = \frac{\Delta \cdot F + \rho \cdot c_p \cdot \frac{e_s - e}{r_{si}}}{\lambda_v \left( \Delta + \gamma \left( 1 + \frac{r_{si}}{r_{st}} \right) \right)} \]  

(A1)

Variables appearing in Equation A1 and others that are not explicitly defined in the text are explained in Table A1. In Equation A1, \( F \) is the net radiation flux, which is the product of total solar radiation measured at the MAWN gage multiplied by the albedo of each cell. Here we assume that albedo varies seasonally from a leaf-off “brown albedo”, \( a_b \), condition to a peak growing season “green albedo”, \( a_g \) value. When \( LAI = 0 \), albedo equals \( a_b \) and increases linearly to \( a_g \) until canopy closure is complete, which we assume occurs at \( LAI = 3 \). Thus albedo

\[ a = \begin{cases} 
  a_g & \text{LAI} \geq 3 \\
  \frac{LAI}{3} (a_g - a_b) + a_b & \text{LAI} < 3 
\end{cases} \]  

(A2)

Values of \( a_b \) and \( a_g \) are provided in Table A2. \( \rho \) (kg m\(^{-3}\)) is the density of moist air given by the ideal gas law:

\[ \rho = \frac{P_{atm}}{R_g \cdot T} \]  

(A3)

The barometric pressure, \( P_{atm} \) (Pa) is calculated according to the barometric formula [Berberan-Santos et al., 1997]:

\[ P_{atm} = P_0 \cdot \exp \left( -\frac{M \cdot g_0 \cdot z}{R_g \cdot T} \right) \]  

(A4)

where \( z \) is the elevation relative to mean sea level (m) given by the DEM. \( e_s \) (Pa) is calculated from the Goff-Gratch equation [Goff and Gratch, 1946]:

\[ \log_{10} e_s = -7.90298 \left( \frac{T_{st}/T - 1}{T} \right) + 5.02808 \log_{10} \left( \frac{T_{st}}{T} \right) - 1.3816 \times 10^{-7} \left( 10^{1.134(1-T/T_{st})} - 1 \right) + 8.1328 \times 10^{-7} \left( 10^{-3.49149(T/T_{st} - 1)} - 1 \right) + \log_{10} e_{st} \]  

(A5)

\( \Delta \) (Pa K\(^{-1}\)) is the slope of the saturated vapor pressure-temperature curve, calculated as the numerical derivative of the Goff-Gratch equation; \( \varepsilon \), the product of \( e_s \) and measured fractional relative humidity, is the ambient water vapor pressure; \( \lambda_v \) is the latent heat of vaporization for water (J kg\(^{-1}\)) [Harrison, 1963]:

\[ \lambda_v = 10^3 (2500.5 - 2.359T) \]  

(A6)

and \( \gamma \) is the psychrometric coefficient (Pa K\(^{-1}\)) [Brunt, 1952]

\[ \gamma = 1.61 \cdot c_p \cdot P_{atm}/\lambda_v \]  

(A7)
The canopy resistance to vapor transport, $r_{ci}$, (m s$^{-1}$) is calculated as

$$r_{ci} = \frac{1}{g_{s} \cdot LAI} \tag{A8}$$

where $g_{s}$ is the stomatal conductance (s m$^{-1}$). Values for maximum stomatal conductance were taken from Schulze et al. [1994]. Unlike Chen et al. [2005], the aerodynamic resistance, $r_{ai}$ (m s$^{-1}$) is calculated based on canopy properties and height-adjusted gaged wind speed [Allen et al., 1998].

$$r_{ai} = \log_{e} \left( \frac{h_{mv} - h_{0}}{l_{m}} \right) \log_{e} \left( \frac{h_{mv} - h_{0}}{l_{v}} \right) \left( k^{2} \nu_{wa} \right) \tag{A9}$$

Because windspeed was only measured at one height, the effective measurement height is adjusted for canopy height. Here we assume that the height to which windspeed is adjusted, $h_{mv}$, is given by

$$h_{mv} = \max(h_{m0}, f_{o} \cdot h_{c}) \tag{A10}$$

where $h_{c}$ is the canopy height assumed constant for a given land cover (Table A2), and $f_{o}$ is a factor to move the adjusted wind height some distance above the canopy. The zero displacement height $h_{0}$ is assumed to be $2/3 \cdot h_{c}$ [Allen et al., 1998]. $l_{m}$ is the roughness length for momentum transport (m) taken as $0.123 \cdot h_{c}$ [Allen et al., 1998] (same as the zero displacement), and $l_{v}$ is the roughness length for vapor and heat transport (m) assumed to be $0.1 \cdot l_{m}$ [Allen et al., 1998]. $\nu_{wa}$ is the measured wind speed (m s$^{-1}$) adjusted for measurement height according to the wind profile power law assumption [Elliot et al., 1986].

$$\nu_{wa} = \nu_{w} \left( \frac{h_{mv}}{h_{m0}} \right)^{\alpha_{w}} \tag{A11}$$

where $\nu_{w}$ is the raw measured wind speed (m s$^{-1}$).

In order to calculate evaporation from leaf surfaces $E_{s}$, surface depressions ($E_{d}$), and the soil ($E_{s}$), Equation A1 is used but the two conductance terms are modified. $r_{ci}$ is set to 0 for evaporation from leaf surfaces and surface depressions, and $r_{ai}$ is calculated using a crop height of 2 cm for surface depression evaporation. To calculate surface soil layer evaporation, $r_{ci}$ is replaced by $r_{s}$ given by [Choudhury and Monteith, 1988]:

$$r_{s} = \frac{\tau \cdot l_{s}}{\Phi \cdot D_{v}} \tag{A12}$$

where $l_{s}$ is the depth from the surface to the top of the evaporative layer of water (m), here assumed to be half the depth of the top soil layer, and $\Phi$ is the total porosity.
The total transpiration in each cell $T_{(m)}$ is the sum of the root water uptake from each biologically active soil layer, $T_1$ and $T_2$. Following Manfreda et al. [2005], we assume in this version of the ILHM code that the actual transpiration is calculated from the potential value by linearly interpolating between 0 at the permanent wilting point and the potential rate at 75% of saturation according to:

$$T_i = \left( S_i > \Phi_{-33} \right) \cdot \min \left[ 1, 0.4 / 3 \cdot \frac{S_i}{(\Phi_i - l_i)} \right] \cdot P \cdot T \cdot \sum_{i} l_i$$  \hspace{1cm} \text{(A13)}$$

where $S_i$ is the soil moisture (m) of the $i^{th}$ soil layer, $\Phi_{-33}$ is the permanent wilting point of the soil (Table A3) and $l_i$ is the thickness of the $i^{th}$ soil layer (m). The term $S_i > \Phi_{-33}$ is a logical statement that returns a value of “1” if true and “0” if false. Porosity values, $\Phi_i$ are taken as a function of soil type as given by Table A3. Biologically active soil thickness is calculated as the depth above which 90% of the root mass lies using the asymptotic equation [Gale and Grigal, 1987]:

$$y = 1 - \beta^d$$  \hspace{1cm} \text{(A14)}$$

where $y$ is the cumulative root fraction at depth $d=0.9$ (cm) for this study; $\beta$ is a land cover-dependent parameter (Table A1). We use Equation A14 to solve for $d$ with a fixed cumulative root fraction $y=0.9$.

$$I = \sum_{i} l_i = \log_{\beta} (0.9)$$  \hspace{1cm} \text{(A15)}$$

Total evaporation $E_{(m)}$ is the sum of canopy evaporation, $E_c$, soil evaporation $E_s$, and surface depression evaporation $E_d$. Soil evaporation is calculated according to Chen et al. [2005] as

$$E_s = \min \left( P E_s d_s \cdot (1 - f_c) \right)$$  \hspace{1cm} \text{(A16)}$$

where $d_s$ is the soil-controlled exfiltration depth (m) calculated by

$$d_s = s_e \cdot \Delta t^{-1/2}$$  \hspace{1cm} \text{(A17)}$$

and $\Delta t$ is the model timestep length (s), while $s_e$ is the soil desorptivity (m s$^{-1/2}$), calculated as in Entekhabi and Eagleson [1989]:

$$s_e = \left[ \frac{8 \cdot \Phi_i \cdot k_{sat} \cdot \Phi_h}{3(1 + 3 \cdot m)(1 + 4 \cdot m)} \right]^{1/2} S_0^{m(2 + 2)}$$  \hspace{1cm} \text{(A18)}$$

where $k_{sat}$ is the saturated hydraulic conductivity of the first soil layer (m s$^{-1}$), $m$ is the pore size distribution index assumed to be a function of soil texture (Table A3), and $S_0 = \theta_i / (\Phi_i - l_i)$ is the fractional saturation. As in Manfreda et al. [2005], the closed canopy fraction ($f_c$) is defined by the empirical relationship [Eagleson, 1982]

$$f_c = 1 - e^{-\mu LAI}$$  \hspace{1cm} \text{(A19)}$$

where $\mu$ is a constant for a given land cover type given by Table A1.

Calculating $E_c$ requires a full canopy water balance model. The canopy water balance is calculated using:
\[ \Delta S_c = P - \text{Int} - E_c \]  

(A20)

where \( \Delta S_c \) is canopy water storage (m). Incoming rainfall is first subjected to interception up to the water holding capacity of the canopy given by Dickinson et al. [1991]:

\[ S_{c,\text{max}} = 1 \times 10^{-4} \cdot \text{LAI}(\text{m}), \]  

(A21)

where LAI is the leaf area index (m²/m²). The available interception capacity of the canopy is then given by

\[ \delta S = S_{c,\text{max}} - S_c^{t-1}. \]  

(A22)

Additionally, we modify the model of Chen et al. [2005] to allow some water to penetrate the canopy at all times based on the assumption that the canopy is not completely closed. Interception at time \( t \) is then

\[ \text{Int} = \min(\delta S, P \cdot f_c). \]  

(A23)

Canopy evaporation, \( E_c \), is then calculated as in Manfreda et al. [2005] with

\[ E_c = \min\left(\left(\frac{S_c}{S_{c,\text{max}}}\right)^{2/3} \cdot P E_c, S_c\right). \]  

(A25)

Surface depression evaporation, \( E_d \), occurs only when water is stored in surface depressions. At each time step, any water stored in surface depressions \( S_d \) from the previous timestep is added to throughfall from the canopy (or snowmelt from the UEB model) such that precipitation excess runoff, \( R_e \), is given by

\[ R_e = \min\left(0, P + S_d^{t-1} - \text{Inf} - \text{Int}\right). \]  

(A26)

where infiltration, \( \text{Inf} \), is calculated as discussed below. \( S_d \) is then calculated as

\[ S_d = \min\left(S_{d,\text{max}}, R_e + E_{x_1}\right) \]  

(A27)

where \( E_{x_1} \) is the exfiltration out of the first soil layer and the depression storage capacity, \( S_{d,\text{max}} \), is assumed to be constant for a given combination of slope, land cover, and soil type (see table in Manfreda et al. [2005] using values from Liu et al. [2003]). Depression evaporation, \( E_d \), is then given by

\[ E_d = \min\left[S_d, (1 - F_c) \cdot P E_d\right]. \]  

(A28)
INFILTRATION, PERCOLATION, THROUGHFLOW, AND EXFILTRATION

The next three terms of the water balance (Equation 1), percolation, $P_c$; throughflow, $T_r$; and exfiltration $E_x$ are calculated within the soil water balance model. First, the outputs of the canopy model, snowmelt model, and depression storage model are used to calculate infiltration into the surface soil layer.

\[ \text{Inf} = \min\{P - Int + S_d^{t-1}, i_{\text{max}}\} \]  
(A29)

The infiltration capacity, $i_{\text{max}}$, is a function of the moisture content of the surface soil layer and is calculated according to

\[ i_{\text{max}} = \max\left(\Phi_1 \cdot l_1 - S_1^{t-1}, i_{\text{sat}}\right) \]  
(A30)

where $l_1$ is the thickness of the first soil layer as defined previously, $S_1$ is the moisture stored within the first soil layer, and $i_{\text{sat}}$ is the saturated infiltration capacity, which can vary with time due to the influence of impermeable frozen soils.

Infiltration is applied to the first soil layer, which can then percolate into the second layer. First, the soil moisture storage at the end of each timestep in the first layer is calculated as

\[ S_1^{t} = S_1^{t-1} + \text{Inf} - E_x - T_1 + P_1 - E_x - S_1^{t} \]  
(A31)

where $P_1$ is the percolation of water from the first soil layer to the second. Note that $T_1$ requires $S_1$. To avoid having to solve the coupled equations, $T_1$ is calculated using an intermediate value of $S_1^{t} = S_1^{t-1} + l_1 - E_1$. Then $S_1 = S_1^{t-1} - T_1$ and $S_1^{t+} = S_1^{t} + T_1 - P_1 - E_x - E_x$. Given $S_1$, percolation into the second layer, $P_1$ is given by

\[ P_1 = \max\{S_1 - S_{\text{max}}, P_{\text{DREAM}}\} \]  
(A32)

where $S_{\text{max}} = \Phi_1 \cdot l_1$ and $P_{\text{DREAM}}$ is the percolation calculated according to Manfreda et al. [2005] given by:

\[ P_{\text{DREAM}} = \begin{cases} 0 \\ S_{\text{max}} - \left[ S_1 - \left( \frac{\Delta t \cdot k_{\text{sat}}(\gamma - 1)}{S_{\text{max}} - S_1^{t-1}} \right) \right] \end{cases} \]  
(A34)

where $\gamma = (2 + 3m)/m$. $P_{\text{DREAM}}$ effectively allows percolation only when soil moisture exceeds the field capacity given by $\Phi_1 \cdot l_1$.

The second-layer soil moisture at the end of the timestep, $S_2^{t}$, is calculated similarly to Equation A31:

\[ S_2^{t} = S_2^{t-1} + P_2 - T_2 + T_r - E_x, \]  
(A35)

where $T_2$ is calculated from the values of $S_2^{t-1}$ at the previous timestep.

\[ P_2 = P_{\text{DREAM}} \]  
(A36)

is calculated from an intermediate value of $S_1^{t-1} = S_1^{t-1} + P_1 - T_2$.

Throughflow out of a cell is calculated as

\[ T_{r_{\text{out}}} = l_1 \cdot \theta \cdot \frac{2}{\sqrt{k_{\theta_0} + k_{\theta \text{down}}}} \left( \frac{\Delta \theta}{\Delta x} + \frac{\Delta \theta}{\Delta x} \left( \frac{\theta_{\text{down}} - \theta_i}{\Delta x} \right) \right) \]  
(A37)

where the effective unsaturated hydraulic conductivity for subsurface flow in layer $i$ is taken as the harmonic average of the unsaturated hydraulic conductivity in the cell, $k_{\theta_0}$ (m s$^{-1}$) and the down slope value $k_{\theta \text{down}}$. $\Delta x$ is the model cell resolution (m); $\alpha$ is the vertical gradient in the down-slope direction; $\Delta \theta$ is the slope in meters of the moisture retention curve in layer $i$, and

\[ \theta = \frac{S_1(t)/S_{\text{max}} - \theta_0}{\Phi - \theta_0} \]  

where $\theta_0$ is the residual volumetric moisture content assumed to be a soil-texture dependent property (see Table A3). The assumption that flow only occurs parallel to the dip of the slope requires that $T_{r_{\text{out}}} \geq 0$. $T_r$ is then calculated as

\[ T_r = T_{r_{\text{out}}} - T_{r_{\text{out}}} \]  
(A38)

Finally, $E_x$ is calculated as the soil water in excess of saturation given by

\[ E_x = \max\left(0, S_2 + T_2 - S_{\text{max}}\right). \]  
(A39)

This is then applied to the first layer prior to calculating

\[ E_x = \max\left(0, S_1 + T_1 + E_x - S_{\text{max}}\right). \]  
(A40)

RUNOFF ROUTING

Water exfiltrated from layer one is then applied to the surface depression model, thus $R$ (m) is simply

\[ R = R_c - S_d \]  
(A41)
which is then routed to the streams using an approach modified from that presented by Manfreda et al. [2005]. Once generated, runoff cannot infiltrate and is instead routed overland and through streams according to the D8 flowdirection algorithm in ARC [ESRI, 2003] with runoff times given by the velocities in each cell along the flowpath. Runoff is assumed to travel overland at a velocity given by the Kerby time of concentration equation [Kerby, 1959]

\[
t_{cell} = \frac{86.735 \left( \frac{l_{cell} \cdot n}{\sqrt{s}} \right)^{0.467}}{(A42)}
\]

where \(t_{cell}\) is the time required to completely traverse a model cell (s), \(l_{cell}\) is the length of the model cell (m), \(n\) is the dimensionless Manning’s Roughness coefficient (values from [McCuen, 2004]) and \(s\) is the fractional slope of the cell in the downslope direction. The velocity (m s\(^{-1}\)) is then

\[
v_{land} = \frac{l}{t_{cell}} \quad (A43)
\]

Once the runoff enters the stream channel its velocity is calculated using Manning’s Equation [McCuen, 2004]

\[
v_{stream} = \frac{1}{n} r^{2/3} s^{1/2} \quad (A44)
\]

where \(r\) is the hydraulic radius (m) given by the ratio of the stream cross-sectional area to the wetted perimeter.

REFERENCES


Farrand, W. R., and D. L. Bell (1982), Quaternary Geology of Southern Michigan, The University of Michigan, Ann Arbor, MI.


