HYDROLOGICAL PROCESSES
Hydrol. Process. (2012)
Published online in Wiley Online Library
(wileyonlinelibrary.com) DOI: 10.1002/hyp.9364

Quantifying watershed surface depression storage: determination and application in a hydrologic model

Joseph K. O. Amoah,[1]* Devendra M. Amatya[2] and Soronnadi Nnaji[1]

[1] *Department of Civil and Environmental Engineering, Florida A&M University, Tallahassee, FL 32310, USA*
[2] *Center for Forested Wetlands Research, USDA Forest Service, Cordesville, SC 29434, USA*

Abstract:

Hydrologic models often require correct estimates of surface macro-depressional storage to accurately simulate rainfall–runoff processes. Traditionally, depression storage is determined through model calibration or lumped with soil storage components or on an ad hoc basis. This paper investigates a holistic approach for estimating surface depressional storage capacity (DSC) in watersheds using digital elevation models (DEMs). The methodology includes implementing a lumped DSC model to extract geometric properties of storage elements from DEMs of varying grid resolutions and employing a consistency zone criterion to quantify the representative DSC of an isolated watershed. DSC obtained using the consistency zone approach is compared to DSC estimated by "brute force" (BF) optimization method. The BF procedure estimates optimal DSC by calibrating DRAINMOD, a quasi-process based hydrologic model, with observed streamflow under different climatic conditions. Both methods are applied to determine the DSC for relatively low-gradient coastal plain watersheds on forested landscape with slopes less than 3%. Results show robustness of the consistency zone approach for estimating depression storage. To test the adequacy of the calculated DSC values obtained, both methods are applied in DRAINMOD to predict the daily watershed flow rates. Comparison between observed and simulated streamflow reveals a marginal difference in performance between BF optimization and consistency zone estimated DSCs during wet periods, but the latter performed relatively better in dry periods. DSC is found to be dependent on seasonal antecedent moisture conditions on surface topography. The new methodology is beneficial in situations where data on depressional storage is unavailable for calibrating models requiring this input parameter.

KEY WORDS depressional storage capacity; DRAINMOD; consistency zone; surface storage; brute force optimization; digital elevation model

Received 17 July 2011; Accepted 20 April 2012

INTRODUCTION

Quantitative description of surface depressional storage at the watershed scale is necessary to provide scientific understanding of the role they play in initiating runoff generation and creating or ameliorating water-related hazards, impact to groundwater table, and stream outflow (e.g. Hancock, 2005; Abedini *et al*., 2006; Harder *et al*., 2007; Wu *et al*., 2007; Dai *et al*., 2010). Surface depression storage is a sensitive input parameter in hydrologic models such as DRAINMOD, MIKESHE, SWMM, and SWAT (Skaggs *et al*., 1991; Tsihrintzis and Hamid, 1998; Dai *et al*., 2010; Muenich, 2011). In hydrology, depression storage is described in terms of depressional storage capacity (DSC), which is the maximum storage that has to be filled in depressions before runoff occurs. Elsewhere, it is referred to as maximum depression storage (e.g. Kamphorst *et al*., 2005; Carvajal *et al*., 2006). DSC is usually incorporated in models as part of initial abstractions or represented as an input parameter (Skaggs, 1978; SCS, 1986; Kim *et al*., 2012). DSC value is either assumed or obtained through model calibration (Ullah and Dickinson, 1979a; Chescheir *et al*., 1994; Richards *et al*., 2005; Harder *et al*., 2006; Dai *et al*., 2010). Spatial variability in the nature and size of depressional storage elements makes developing a generalized empirical relationship for DSC in terms of these characteristics very prohibitive. Linsley *et al*. (1949) suggested a popular analytical equation for estimating the volume of a depression element (V_d) as a function of precipitation excess as

$$V_d = D_{sc}\left(1 - e^{-kP_e}\right) \quad (1)$$

where D_{sc} is the maximum depression storage capacity (mm), k is the constant equivalent to $1/D_{sc}$ (mm^{-1}), and P_e is the rainfall excess (gross rainfall minus evaporation, interception, and infiltration) (mm). The ability to estimate surface depression storage at any given time using Equation (1) requires known values of D_{sc}.

A generalized and reliable procedure for estimating DSC has been the subject of scientific research for the past three decades (e.g. Mitchell and Jones, 1976; Sneddon and Chapman, 1989; Hayashi and van der Kamp, 2000; Abedini *et al*., 2006; Martin *et al*., 2008). Several methods have been suggested for quantifying DSC (e.g. Onstad, 1984; Hayashi and van der Kamp, 2000; Kamphorst *et al*., 2005; Carvajal *et al*., 2006).

*Correspondence to: Joseph K. O. Amoah, Now at St Johns River Water Management District, Palatka, FL 32177, USA.
E-mail: owarek@yahoo.com

The storage–depth model approach fits a relational function, mostly power–law model, between geometric properties of depression storage elements (Mitchell and Jones, 1976; Ullah and Dickinson, 1979b; Hayashi and van der Kamp, 2000). The models are set up as volume–depth, volume–area, and area–depth relationships. These models fail to account for interconnectivity and the three-dimensional nature of depression elements. Depression-filling methods for filling pits in digital terrain surfaces such as digital elevation models (DEMs) have been employed to estimate DSC by Martin *et al*. (2008), Kamphorst and Duval (2001), Planchon and Darboux (2001), Onstad (1984), and Moore and Larson (1979) at DEM square grid resolutions of 7.5–250 mm, 2–24 mm, 1 mm, 13 mm by150 mm (rectangular), and 50 mm, respectively. This approach uses various algorithms to identify and delineate depressions in DEMs, fill them with water until completely inundated or begin to overflow, and then subtract the empty storage from the filled DEM to obtain the DSC. Although the depression-filling approach yields much accurate estimates of depression storage, constructing DEMs in high resolution is cost intensive (Abedini *et al*., 2006), as it requires sophisticated equipment and remotely sensed satellite platforms, which are not affordable for most research studies. Consequently, the surrogate variable approach was developed by establishing an empirical relationship between DSC and site-specific surface roughness indices such as random roughness, limiting elevation difference, land slope, and tortuosity (Onstad, 1984; Hansen *et al*., 1999; Kamphorst *et al*., 2000). The Onstad (1984) equation, which has found widespread application, expresses maximum depressional storage, D_{sc} (cm), as

$$D_{sc} = 0.11R + 0.031R^2 - 0.012RS \qquad (2)$$

where R is the random roughness in centimetres and S is the slope steepness of land surface expressed in percent. The surrogate variable approach is not computationally efficient because a roughness index like random roughness depicts only transects of the DEM and does not account for spatial distribution of depressions (Kamphorst *et al*., 2005). Limited studies have attempted to measure DSC directly by imitating real soil micro-topography by rendering the soil impervious using polyester resin (Gayle and Skaggs, 1978; Kamphorst and Duval, 2001; Planchon and Darboux, 2001) or plastic film (Mwendera and Feyen, 1992).

Past research efforts focused predominantly on micro-relief storage (Carvajal *et al*., 2006; Martin *et al*., 2008), with few addressing isolated macro-relief storage (e.g. Gayle and Skaggs, 1978; Sneddon and Chapman, 1989; Chescheir *et al*., 1994). Most methods used in estimating depression storage were implemented on agricultural fields where tilled soil surfaces predominate (e.g. Darboux *et al*., 2002; Kamphorst *et al*., 2005; Carvajal *et al*., 2006). In addition, most of the studies were conducted on small plots of land (Darboux *et al*., 2002; Kamphorst *et al*., 2005) or on single isolated depressions (Abedini *et al*., 2006) and therefore fail to capture complex spatial distribution and interconnection of micro- and macro-relief storage on a watershed. The main objective of this study is to develop a new methodology for estimating DSC at the watershed scale and test the performance of estimated DSC values by applying them in DRAINMOD model to predict watershed flow rates on forested watersheds with large macro-topography.

MATERIALS AND METHODS

Study Site

The site comprises of six watersheds located in the lower Atlantic Coastal Plain in the south-eastern United States (Figure 1). Five of the watersheds are located at Santee Experimental Forest within USDA Forest Service Francis-Marion National Forest (FMNF), located 55 km north-west of Charleston, SC. Three of the five watersheds—WS80 (160 ha), WS77 (150 ha), and WS79b (166 ha)—are contained within Santee Experimental Forest (33°8′N, 80°49′W) near Huger, SC. The fourth watershed, WS79 (481 ha), is a combination of watersheds WS80, WS77, and WS79b. The fifth 5000-ha watershed, WS78 (Turkey Creek), is located adjacent to the Santee Experimental Forest. Bannockburn, the sixth watershed, is located within Bannockburn Plantation, an undeveloped 1377-ha parcel of land located in coastal Georgetown County, SC (33°22′48″N, 79°10′12″W).

The watersheds on low topographic relief ranging from 1% to 3 % are mostly drained by small streams, which are characterized by low gradient streambed and side slopes and relatively broad stream bottoms, which contribute to slow surface drainage. Soils in the region often have

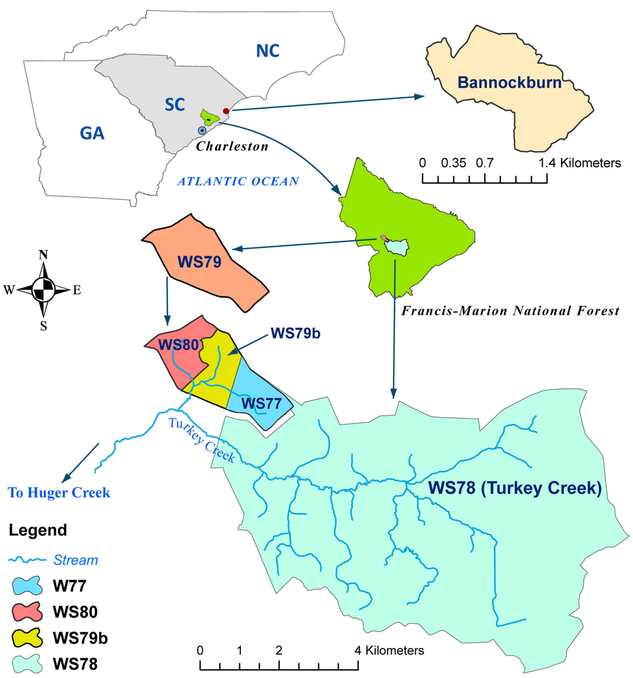

Figure 1. Location of study site

DOI: 10.1002/hyp

clayey subsurface layers, which restrict internal drainage. These headwater catchments in the south-eastern lower coastal plain often contain forested wetlands, which are classified into three general categories: riverine, depressional, and pine flatwood (Harder, 2004). Climate in the region is warm–temperate, with an average annual rainfall of 1350 mm and a temperature of 18.5 °C.

Lumped DSC Model

A tool called "lumped DSC (LDSC) model" is developed to automate the process of extraction and computation of geometric characteristics of storage elements in a watershed. The LDSC model, a time-saving utility tool, is built in the ModelBuilder environment within an ArcGIS platform (ESRI, 2009) and takes advantage of common and specialized geoprocessing system tools. Building the model in ModelBuilder has several advantages, including the ability to automate the model at once and a faster way of changing model parameters in a visual workflow (ESRI, 2009).

The LDSC model, which is a concatenation of geoprocessing tools, takes contour or spot elevation data as input to generate a DEM at a user-specified grid size. The model parameters include drainage enforcement, DEM extent, maximum number of iterations, and other optional parameters. Drainage enforcement provides options to remove artefacts from the derived DEMs. Next, the LDSC model extracts DEM based on the watershed demarcation. This step is necessary because DEM generated in the first process normally extends beyond the required watershed boundary. The model identifies depressions in the intermediate DEM and fills them with water using the eight-direction pour point algorithm presented by Jenson and Domingue (1988). To delineate a local depression, the model assumes that water flows downhill to adjacent cells with lowest elevation. The LDSC model identifies a low-elevation cell in a square-gridded DEM by comparing it with its eight neighbours, four diagonal and four orthogonal. Each low-elevation grid is added to the local depression in an iterative manner until an overflow or pour point grid is located (Ullah and Dickinson, 1979a). A pour point is a DEM grid cell whose surrounding neighbours are at higher elevation and serves as an outflow location. A connected chain of such pour points defines the boundary of an isolated surface depression in a DEM. In other words, on gridded digital elevation data, depressions occur when flow direction in a cell or set of spatially connected cells cannot be assigned to one of the eight valid flow directions (Mark, 1988).

In the next step, the LDSC model fills isolated or connected depressions with water until they are completely inundated or begin to overflow. This is accomplished by iteratively raising DEM grid cell elevations inside the depressions to the pour point elevation until it begins to overflow to surrounding cells. The difference between the filled and empty depressions is referred to as DSC. Finally, the LDSC model computes geometric properties of depressional storage elements. Surface area is calculated as the product of the number of DEM grid cells in a depression and grid area. Depressional volume is the summation of the product of each grid cell area and its depth to pour point. Further information about the LDSC model can be found elsewhere (Amoah, 2008).

The LDSC model was applied to compute geometric properties of surface depressions in the six study watersheds. Input data to the model consisted of 1-ft (0.3-m) contour data, except for the Bannockburn watershed, where 0.1-m vertical precision light detection and ranging (LiDAR) spot elevation data were acquired. For each watershed, the LDSC model was ran seven times to extract geometrics of surface storage elements based on intermediary DEM created at 1-m, 5-m, 10-m, 15-m, 20-m, 25-m, and 30-m square grid resolutions (Amoah, 2008).

Analysis

Execution of the LDSC model generates a map depicting spatial distribution of depressional storage elements and associated volume–area properties of each storage element within the watersheds. A typical display of spatial distribution of surface depressions in a watershed at different DEM grid resolutions is provided in Figure 2. Distribution of surface storage elements is based on the topography of the landscape. The number of depressional elements tends to decrease with increase in DEM grid size, as coarser DEMs fail to capture small storage elements. Beyond 10-m grid spacing, the number of depressional storage elements as well as spatial distribution reduces. The quantity, spatial distribution, and calculated DSC value are largely dependent on the quality of the source dataset (spot elevation or contour) used in generating DEMs.

The number of surface storage elements per area of watershed decreased with an increase in DEM grid cell size (Figure 3). This behaviour may be attributed to the effect of DEM resolution on the description of the topographical features. The effect of grid spacing on the total area flooded in each watershed is displayed in Figure 4. In general, the total area ponded decreased as DEM grid spacing increased, which is in contrast to pattern reported by Abedini *et al.* (2006), where they reported an increase in percentage area ponded with an increase in DEM resolution. The disparity in the trend of variation may be due to the scale effect. Abedini *et al.* (2006) utilized surface micro-topography at DEM grid size of 0.003 m to 0.03 m, as compared to the 5-m to 30-m DEM grid resolutions employed in this study. At relatively larger DEM grid resolutions, there is a smoothing of the surface topographic structure, and, as a result, depressional elements are lost, thereby decreasing the available total depressional surface area.

The geometric properties of depressional elements obtained from the LDSC model output are used to calculate the representative surface DSC of a watershed. The DSC for a watershed is calculated as follows:

$$D_{sc} = \sum_{i=1}^{n} V_i \Big/ \sum_{i=1}^{n} A_i \qquad (3)$$

DOI: 10.1002/hyp

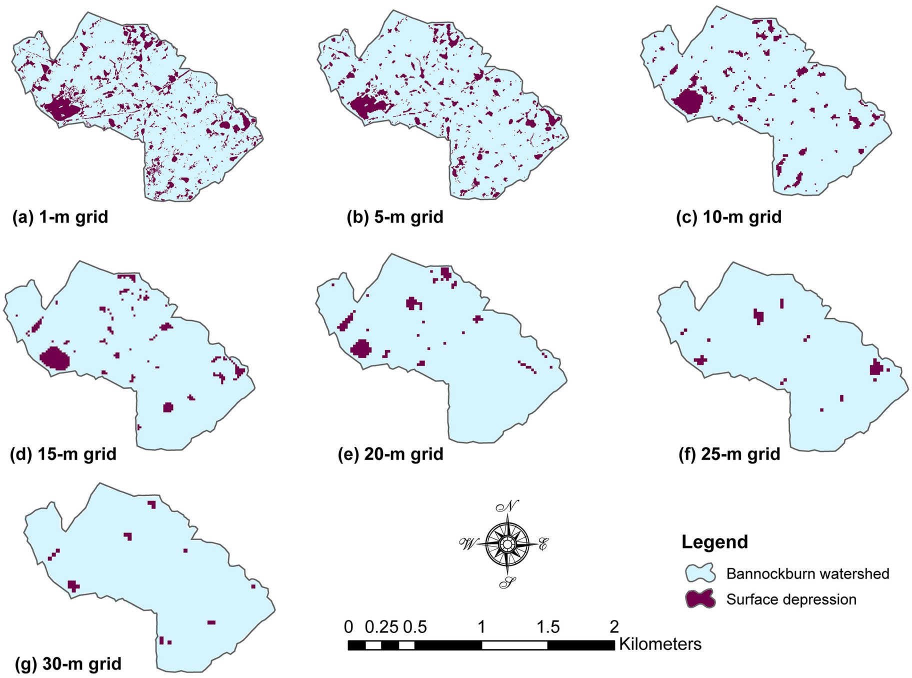

Figure 2. Spatial distribution of depression storage in Bannockburn watershed at different digital elevation model grid resolutions. The depressional elements are shown in plum

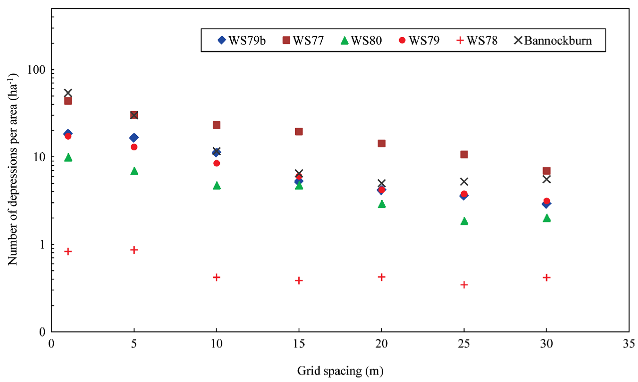

Figure 3. Effect of digital elevation model resolution on depression storage elements

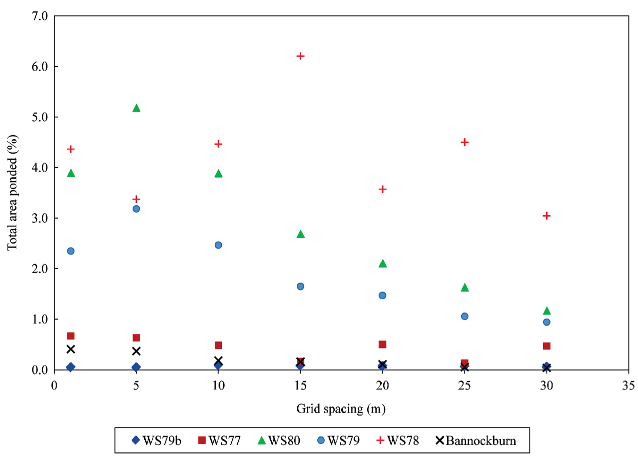

Figure 4. Influence of grid spacing on total area ponded

where D_{sc} is DSC expressed as depth, V_i is the volume of the ith depression element, A_i is the area of the ith depression element, and n is the total number of storage elements within a watershed. Equation (3) assumes that depressions within a watershed are disconnected and that each depression is filled to capacity.

Plots of calculated DSCs reveal a pattern in the distribution of DSC across different DEM grid spacing (Figure 5). In general, DSC does not differ substantially for the first four DEM grid resolutions and then begin to taper upwards or downwards as grid size increases. DSC remains reasonably consistent from 1-m to 15-m square grids. Over this range of DEM resolutions, DSC did not show significant variation for all watersheds considered in this study. However, beyond the 15-m square grid, DSC tends to decrease exponentially with an increase in DEM resolution for watersheds WS80, WS79, WS79b, and Bannockburn. This behaviour may be attributed to the nature of the topographic structure and the mechanism by which ANUDEM (Hutchinson, 1988) algorithm

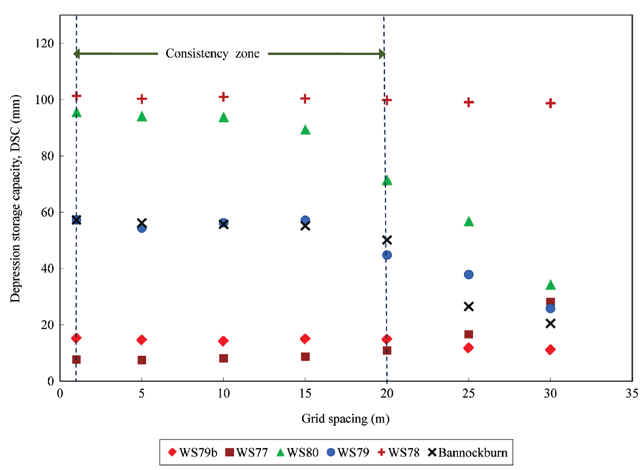

Figure 5. Variation of surface depressional storage capacity with digital elevation model grid resolution

DOI: 10.1002/hyp

constructs DEMs for given input contour data. Misrepresentation of actual surface topography at large grid spacings may potentially cause generated DEMs to be characterized by large depressions. The observed pattern of DSC beyond the 15-m square grid is similar to trends reported by Martin *et al.* (2008) and Carvajal *et al.* (2006). In contrast, DSC distribution for watershed WS77 followed an opposite pattern, in which large grid spacings produced high depression storage values. The observed trend in WS77 is in agreement with results obtained by Abedini *et al.* (2006). Watershed WS78 showed a different pattern of change compared to the other watersheds, as the DSCs obtained did not exhibit much variability across the various DEM grid resolutions tested herein. This invariability in WS78 may be attributed to the robust topographic structure of the terrain and the larger size and spread of surface storage elements, all of which could be verified with LiDAR data recently obtained for this site.

The DSC values for WS77 range between a minimum of 7.5 mm (21.3 m^3) at the 5-m square grid to a maximum of 28.2 mm (202.7 m^3) at the 30-m square grid, with an average of 12.5 mm (78.7 m^3). The DSC values exhibit moderate stability from the 1-m to the 15-m DEM grid and then begin to increase exponentially with an increase in grid spacing. For WS80, the DSC values range from 34.3 mm (650.6 m^3) at the 30-m grid to 95.6 mm (7905.5 m^3), with an average of 76.5 mm (4048.3 m^3). WS79b exhibits a trend similar to WS77, with a maximum DSC value of 15.2 mm (318.0 m^3) occurring at the 1-m grid and a minimum DSC of 11.1 mm (646.7 m^3) obtained at the 30-m grid, with a mean DSC of 13.8 mm (465.7 m^3). WS79 combines the surface depression characteristics of WS80, WS77, and WS79b. The minimum and maximum DSC values for WS79 are 25.8 mm (1171.3 m^3) and 57.3 mm (8324.0 m^3), occurring at 30-m and 1-m grid spacing, respectively, with a mean value of 47.7 mm (4610.7 m^3). The DSC values for the Bannockburn watershed ranges between 20.5 mm (405.9 m^3) and 57.3 mm (11,643 m^3), with a mean value of 45.9 mm (5016 m^3). The DSC value computed for WS78 remained fairly consistent, with minimum and maximum DSCs occurring at the 30-m and 1-m grids with values of 98.7 mm (150,369 m^3) and 101.3 mm (221,105 m^3), respectively, and a mean DSC of 100.1 mm (211,222 m^3).

CRITERION FOR ESTIMATING DEPRESSION STORAGE

From calculations and visual inspection of Figure 5, the relative difference between DSC values in the 1-m to 15-m grid spacing range is less than 7%, except for watershed WS77, which had a relative difference of 14%. Given the consistency of the DSC values over a region or zone of DEM grid resolutions, it is inferred that a representative surface DSC of a watershed is given by the region of grid spacings, where the calculated DSC does not change substantially (less than 15%) over given DEM resolutions. In this case, consistency zone refers to a region between 1-m and 15-m square grids. Hence, the DSC value between the upper and lower limits of the equilibrium region may represent the DSC of a watershed. This implies that the DEM of a relatively flat watershed (land slope less than 5%) with grid spacing between 1 m and 15 m should be suitable for estimating surface macro-relief storage. The consistency zone may vary based on the watershed characteristics and topographic relief. The suggested range of grid spacings for providing quantitative description of DSC is in agreement with the DEM resolution proposed by Zhang and Montgomery (1994) and Hancock (2005) for extracting hydrological and geomorphological features of the landscape. Exception occurs for conditions where DSC varies below 15% across all DEM resolutions under study. In such cases, scientific judgment may be employed to calculate the average representative value based on DSC distribution across the DEM resolutions.

Application of the consistency zone criterion yielded optimal DSC for each watershed (Table I). Standard deviation of the DSC values across DEM grid resolution is mostly marginal (0.5 mm), with the greatest deviation (2.7 mm) occurring in watershed WS80, which contained relatively larger wetland areas (Harder *et al.*, 2007; Dai *et al.*, 2010). DSC values obtained using the consistency zone methodology is compared to the traditional calibration method for estimating DSC. Harder *et al.* (2006) reported an average DSC of 80 mm in calibrating DRAINMOD (Skaggs, 1978) for WS80. Dai *et al.* (2010) applied a DSC value between 40 and 80 mm in watersheds WS80, and between 10 and 180 mm in WS79 in calibrating the MIKESHE model (DHI, 2005). Field reconnaissance survey on WS79b and the Bannockburn watershed suggests an average DSC value of less than 50 mm. Above reported DSC values reasonably match the values obtained using the consistency zone method. To validate the DSC obtained for WS77, the remainder of this paper focuses on employing the traditional calibration method to obtain optimum DSC.

APPLICATION OF DSC IN DRAINMOD

The depression storage value obtained for watershed WS77 using the preceding technique is compared to the estimate of the DSC value obtained by calibrating

Table I. Representative depression storage capacity and summary statistics based on consistency criterion for six watersheds

Statistic	WS77	WS78	WS80	WS79	WS79b	Bannockburn
Mean (mm)	10	100	93	56	15	56
Std. Dev. (mm)	0.5	0.5	2.7	1.3	0.5	0.9
Range (mm)	1.2	1.0	6.2	2.9	1.1	2.1

DOI: 10.1002/hyp

DRAINMOD (Skaggs, 1978) using the "brute force" (BF) optimization procedure. DRAINMOD is a quasi-process based, field-scale hydrologic model developed to simulate the hydrology of poorly drained, high water table soils on an hour-by-hour, day-by-day basis or for long periods of time (Skaggs, 1978). With the BF method, several DSC values are applied to calibrate DRAINMOD, assuming all other parameters are fixed, to find the optimum DSC value yielding minimal error relative to observed data. DRAINMOD accepts DSC as a direct input parameter and is represented as an average depth of depression storage that must be satisfied before runoff can begin. Surface depression storage is captured in DRAINMOD as STMAX and STORRO. STMAX represents the maximum surface storage (macro-relief storage) that must be filled before surface runoff occurs (Skaggs, 1978). STORRO (micro-relief storage) is the storage in local depressions due to soil structure and cover that control the movement of water on the surface. In DRAINMOD, DSC is represented as STMAX and serves as the basis for comparison with results obtained from the LDSC model approach. Details of the model and modelling procedure are described elsewhere (Skaggs, 1978; Amatya *et al.*, 1997; Amatya and Skaggs, 2001). Although surface storage in depressions varies with time (Viessman and Lewis, 2003) and also with soil moisture (Muenich, 2011), it is assumed to be constant in the current version of DRAINMOD (Skaggs, 1978).

Modelling Site and Input Data

The 150-ha headwater watershed WS77, located in Santee Experimental Forest (Figure 6), was established in 1963 as a treatment in the paired system with WS80 (control) with an objective of studying the hydrologic and water quality effects of prescribed burning on the poorly drained coastal plain soils (Harder *et al.*, 2007; Amatya *et al.*, 2006). A first-order stream drains watershed WS77 into the perennial Fox Gulley Creek further down to Turkey Creek, a tributary of Huger Creek, which drains further down to Cooper River, an estuarine river of the Atlantic Ocean.

Soils in WS77 are mainly of Wahee-Craven soil association, which are somewhat poorly to moderately drained sandy loam to clayey soils with seasonally high water tables (SCS, 1980). Land use is predominantly forest comprising of loblolly pine (*Pinus taeda L.*), longleaf pine (*Pinus palustris*), and some bottomland hardwoods along the stream riparian bank (Amatya *et al.*, 2006). This low-gradient watershed has surface elevations ranging from 10.5 m above mean sea level in upland areas to about 5.6 m at the watershed outlet, with topographic relief up to 2% slope. The climate in the region is warm–temperate, with an average daily temperature of 16 °C. The average annual rainfall is 1375 mm, with approximately 40% occurring during June–August (Amatya *et al.*, 2006). A gauging station with a compound concrete V-notch weir, installed at the outlet of WS77, measures stream outflows (Figure 6). DRAINMOD requires input parameters, including soil hydraulic properties (saturated hydraulic conductivity at each layer, soil water characteristics data, volume drained

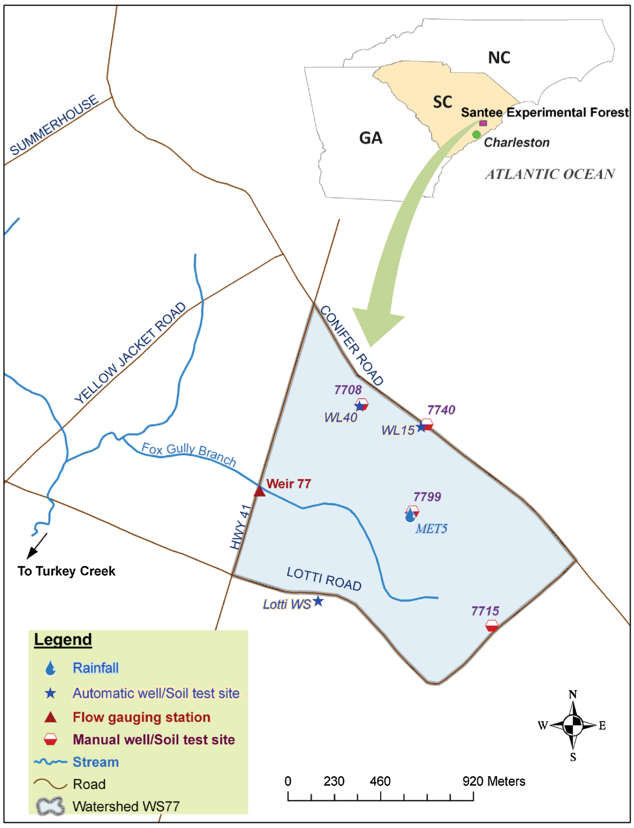

Figure 6. Location of watershed WS77 and Santee Experiment Forest, Huger, SC

and upward flux versus water table depth, wilting point, and Green-Ampt infiltration parameters), watershed characteristics, drainage design parameters (lateral ditch dimensions, spacings, surface depressional storage, slope, depth to impervious layer, initial water table depth), and climatological data (precipitation, air temperature, wind speed, relative humidity, and net radiation). Soil hydraulic properties data and drainage system characteristics applied to DRAINMOD are discussed in Amoah (2008) and summarized in Table II.

Daily potential evapotranspiration (PET) estimated by alternate methods can be input in the model for simulating

Table II. Characteristics of drainage system and soil hydraulic properties

Parameter	Value
Channel depth (cm)	40
Stream spacing (m)	800
Depth to impermeable layer (cm)	150
Drainage coefficient (cm/day)	10
Bottom width of channel (cm)	120
Channel side slope (H:V)	0.5
Initial water table depth (cm)	30
Hydraulic conductivity, cm/h (depth range, cm)	10 (0 to 30) 0.4 (30 to 80) 1 (80 to 150)
Drainable porosity	0.05
Water content at saturation (cm^3/cm^3)	0.40
Water content at wilting point (cm^3/cm^3)	0.13
Root depth (cm)	45

DOI: 10.1002/hyp

ET if the default method using the Thornthwaite method with daily maximum and minimum air temperature is not used. Accordingly, in this study, daily PET was estimated using the Penman–Monteith method for a grass reference instead of forest (Dai *et al*, 2010; Harder *et al.*, 2006). Generally, forest ET is simulated using PET with reference to micro-meteorology measured on forest canopy and physiological variables like leaf area index (LAI) and maximum stomatal conductance for a tall forest vegetation, but these data were unavailable for the site during this study period. Recent study by Sun *et al.* (2010) has shown that grass-referenced PET for a clear-cut stand with just understory vegetation may be 10%–20% lower than matured forest PET partly due to lower net radiation (albedo) of grass. Recently, Kim *et al.* (2012) showed the greatest sensitivity of DRAINMOD predicted event outflow to the method of estimating PET than three other parameters including surface storage. Similarly, Licciardello *et al.* (2011) demonstrated that the SWAT model was more sensitive to the PET parameter than six other parameters impacting surface runoff in a small Mediterranean watershed.

Model Calibration

DRAINMOD is calibrated with observed daily stream outflows for wet and dry seasons in a three-year period (2002–2004). 2003 represents a wet year (outflow is 36% of annual rainfall of 1782 mm), and 2004 represents a dry year (outflow is 9% of annual rainfall of 969 mm). The observed streamflow in 2002 is used as a "warm-up" period to stabilize the model runs. The BF optimization procedure is used to determine optimal DSC values for both wet and dry seasons. In the BF optimization process, a DSC value of 3.0 cm is initially applied in DRAINMOD to simulate the streamflow. Increments and decrements of DSC from an initial value are applied to the model and re-executed, while holding other parameters constant. Model performance evaluation is based on analysis of graphical plots and quantitative measures, including daily, monthly, and cumulative outflow volumes, average absolute daily deviation (AADD), Nash–Sutcliffe coefficient of efficiency (NSE), coefficient of determination (R^2), percent bias (PBIAS), and root mean square error (RMSE)–observed standard deviation ratio (RSR) (Moriasi *et al.*, 2007). Due to lack of established criteria for interpreting RMSE statistic in hydrologic modelling analysis (Legates and McCabe, 1999; Singh *et al.*, 2004), RSR is adopted as the error index for evaluating DSC values. RSR standardizes RMSE using the observations standard deviation, and it combines both an error index and scaling/normalization factor (Moriasi *et al.*, 2007). RSR is expressed as

$$\mathrm{RSR} = \frac{\mathrm{RMSE}}{\mathrm{STDEV_{obs}}} = \frac{\sqrt{\sum_{i=1}^{n}\left(Y_i^{\mathrm{obs}} - Y_i^{\mathrm{sim}}\right)^2}}{\sqrt{\sum_{i=1}^{n}\left(Y_i^{\mathrm{obs}} - Y^{\mathrm{mean}}\right)^2}} \qquad (4)$$

where Y_i^{obs} is the *i*th observation of the constituent, Y_i^{sim} is the *i*th simulated value of the constituent, $Y^{\mathrm{mean}}$is the mean of observed constituent, and n is the total number of observations.

The RSR varies from the optimal perfect model simulation value of 0 to a large positive value (Moriasi *et al.*, 2007). It implies that model simulation performance is better when RSR is low. Therefore, a DSC value that yields minimum RSR is deemed optimal for the watershed.

Calibration Results and Discussion

For each calibration scenario, all other input parameters are kept constant while DSC values between 0.5 cm and 10 cm are applied to DRAINMOD, and RSR calculated using simulated and observed streamflows. For the wet calibration period (2003), the lowest RSR occurs at an optimum DSC of 1.5 cm (Figure 7a). Daily and cumulative stream outflows simulated using an optimum DSC (DSC_{wet}) of 1.5 cm for the wet period is illustrated in Figure 8. DRAINMOD adequately predicted daily flow events, including the long dry period from Day 256 to the end of the year, but consistently overpredicted peak flow rates. Although, by itself, the lower DSC parameter (1.5 cm) that controls the surface runoff might have caused overpredictions, it may likely also be due to lack of flow routing process in DRAINMOD to account for travel time for movement of runoff. The field-scale model assumes that all simulated outflows are instantaneously discharged to the watershed outlet without the possibility of being in retention for evaporation or infiltration

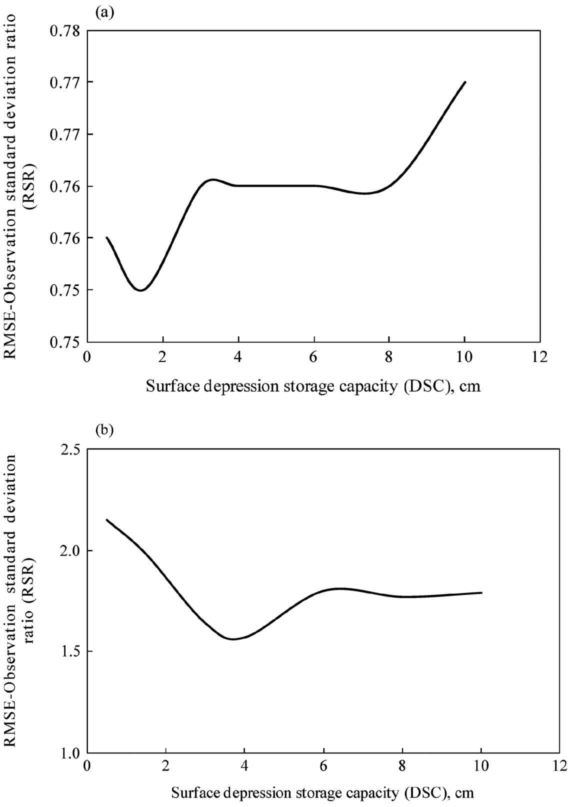

Figure 7. DRAINMOD flow calibration estimated optimum DSC for (a) wet period (2003) and (b) dry period (2004)

DOI: 10.1002/hyp

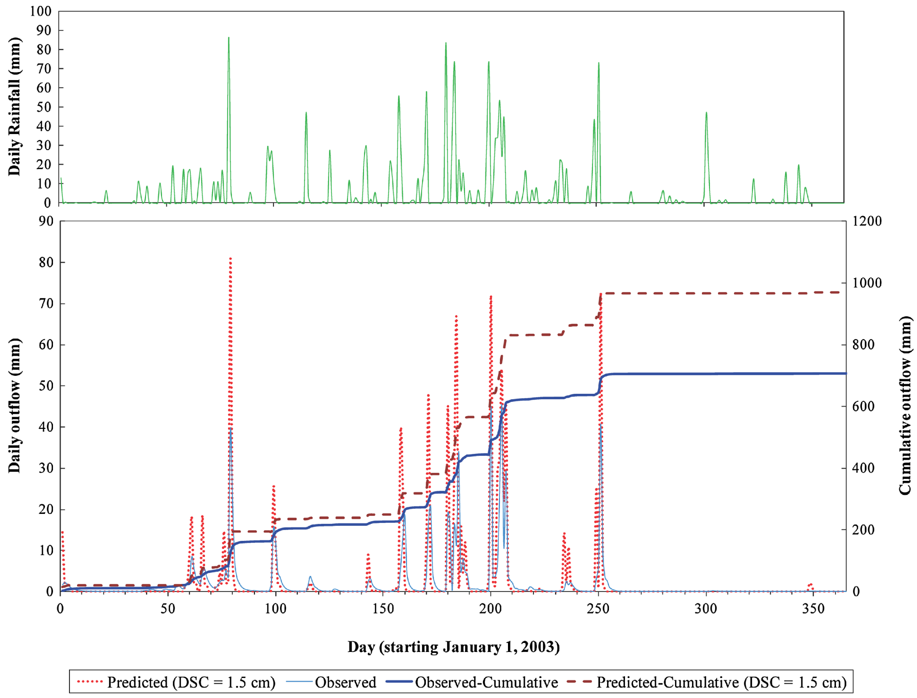

Figure 8. Observed and DRAINMOD predicted daily and cumulative outflows for wet period (2003) using optimum DSC of 1.5 cm in watershed WS77 during model calibration. Daily rainfall is also shown in the top plot

(Amatya *et al.*, 1997). Another possible reason for summer–fall events (Days 150–250) may be due to rainfall variability across the large watershed (153 ha). In 2003, MET5 had 1770 mm of rainfall recorded, which is 7.8% higher than the 1642 mm recorded at the Lotti Rain gauge (0.4 miles away). Overprediction may also be attributed to the use of grass-reference–based PET for the forest stands and the exclusion of forest canopy interception of rainfall, which is not modelled in the current version of DRAINMOD, possibly resulting in lower ET and higher drainage.

On an annual basis, the total cumulative stream outflow for the wet period (2003) was overpredicted by 263.2 mm, representing 37% of the total observed stream outflow of 706.9 mm (Table III). The largest difference occurred after Day 180 (July). The difference between the average monthly observed and predicted stream outflows is 13.7 mm (less than 15 mm) for the entire wet period. The error between the observed and simulated average daily streamflows for 2003 is 0.73 mm, representing an overprediction of 38%. AADD for DRAINMOD predicted streamflow is within 2.0 mm day^{-1} using a DSC_{wet} of 1.5 cm. In spite of consistent overprediction of flow peaks, DRAINMOD performed satisfactorily in flow predictions for the calibration period based on performance ratings (Moriasi *et al.*, 2007) with NSE (=0.67), R^2 (=0.96), high PBIAS (= −34%), and RSR (=0.55).

DRAINMOD is also calibrated with streamflow for the dry period (2004). Several DSC values ranging from 0.5 cm to 10 cm are applied to the model and executed for each adjustment of DSC parameter. Optimum DSC_{dry} occurred at a DSC of 3.5 cm (Figure 7b). The model overpredicted annual stream outflows, with the greatest overprediction occurring in February and September 2004 in response to intense rainfall events (Figure 9). DRAINMOD failed to capture small events in April and May (Days 110 and 127).

Table III. Comparison of simulation results for wet periods (2003 and 2005) during DRAINMOD calibration and validation using optimum DSC of 1.5 cm and direct application of DSC of 1 cm

		Optimization DSC_{wet} = 1.5 cm		Consistency DSC_L = 1 cm
Statistics	Period	Calibration (2003)	Validation (2005)	Application (2005)
Predicted mean daily flow (observed), mm	Daily	2.66 (1.93)	1.14 (0.85)	1.11 (0.85)
Predicted annual flow (observed), mm	Annual	970.1 (706.9)	408.2 (304.0)	398.8 (304.0)
Rainfall, mm	Annual	1770	2476	2476
AADD, mm day^{-1}	Daily	2.0	1.1	1.3
R^2	Monthly	0.96	0.88	0.88
NSE	Monthly	0.67	0.75	0.75
RSR	Monthly	0.55	0.48	0.48
PBIAS, %	Monthly	−34	−34	−31

DOI: 10.1002/hyp

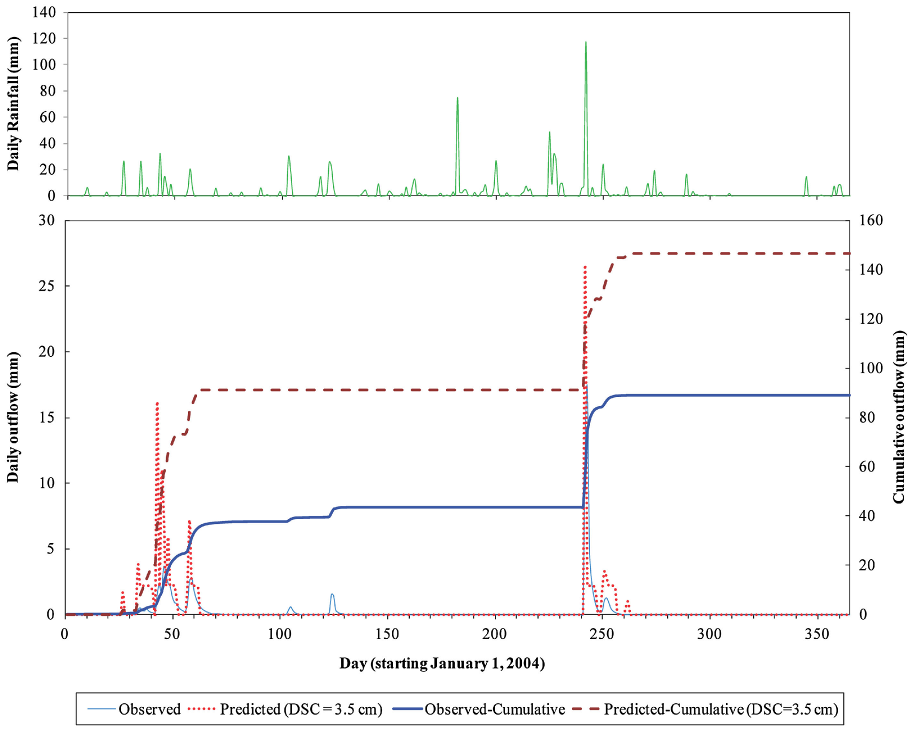

Figure 9. Observed and DRAINMOD predicted daily and cumulative outflows for dry period (2004) using optimum DSC of 3.5 cm in watershed WS77 during model calibration. Daily rainfall is also shown in the top plot

Part of flow overprediction in February 2004 may be due to high rainfall of 169 mm measured at MET5 in January and February compared to only 146 mm at the nearby Lotti Rain gauge. In February alone, MET5 recorded 18 mm more rain, which may result in higher flow rates when soils are near saturation during winter months with low ET demands. A similar oversimulation reason may hold true in September 2004, where the Lotti Rain gauge recorded only 246 mm of rain compared to 281 mm at MET5. When soils are already saturated with depressions still filled up, any excess rainfall directly becomes runoff yielding overpredictions. This observation is consistent with the measured water table elevations (not shown). Again, the model assumption of instantaneous flow arrival at the outlet without consideration of time in retention due to routing may cause the overprediction of peak flow rates. Likewise, the dry forest canopy absorbing part of the precipitation prior to reaching the land surface during the dry summer months and the possible underprediction of ET using grass-reference PET employed in simulations, especially during summer months when ET demands are high, may have contributed to overprediction. As discussed above, daily flows predicted by DRAINMOD (Kim *et al.*, 2012) and SWAT (Licciardello *et al.*, 2011) are very sensitive to the estimated PET values input into the model, especially during the growing season. Furthermore, computed Penman-Monteith reference evapotranspiration (REF-ET) for a clear-cut stand has been shown to be about 15%–20% lower than for forest reference depending upon rainfall/moisture conditions (Sun *et al.*, 2010) and may result in overprediction of stream outflows (Amatya *et al.*, 2003; Harder *et al.*, 2006; Dai *et al.*, 2010).

The predicted annual streamflow exceeded the observed data by 30.5 mm, representing an oversimulation of 37%. An R^2 of 79% suggests a somewhat acceptable linear relationship between the observed and predicted monthly streamflows. However, the low NSE, high RSR, and high PBAIS indicate unsatisfactory model performance during dry periods (Table IV). Harder *et al.* (2006) and Dai *et al.* (2010) also found DRAINMOD to perform unsatisfactorily during dry periods for adjacent control watershed WS80 mainly due to errors in modelling ET.

Model Validation and Application

Optimal DSC values obtained from DRAINMOD calibration are applied during validation stage. The consistency zone approach–calculated DSC values are concurrently applied to DRAINMOD (without further calibration), and the results are compared with the validation simulations.

In the first scenario, DRAINMOD is validated for the wet period (2005) using DSC_{wet} (1.5 cm). DSC_L (1 cm) estimated using the consistency technique is also applied to DRAINMOD to simulate the streamflow for 2005 (Figure 10). In general, both the DSC_{wet} and DSC_L input parameters correctly predicted flow events, although they again consistently overpredicted flow peaks primarily due to the scaling effects of flow routing in stream, which is not considered by DRAINMOD. Such effects may be important when the size of a field exceeds some threshold value such as 75 ha, as shown by Amatya *et al.* (2000). Both DSC values yielded closely matched daily hydrographs. DSC_L overpredicted peak events more

DOI: 10.1002/hyp

Table IV. Comparison of simulation results for dry periods (2004 and 2006) during DRAINMOD calibration and validation using optimum DSC of 3.5 cm and direct application of DSC of 1 cm

Statistics	Period	Optimization $DSC_{dry} = 3.5$ cm		Consistency $DSC_L = 1$ cm
		Calibration (2004)	Validation (2006)	Application (2006)
Predicted mean daily outflow (observed), mm	Daily	0.33 (0.24)	0.65 (0.49)	0.56 (0.49)
Predicted annual outflow (observed), mm	Annual	119.6 (89.1)	235.2 (172.3)	203.9 (172.3)
AADD, mm day^{-1}	Daily	0.26	0.45	0.69
R^2	Monthly	0.79	0.81	0.77
NSE	Monthly	0.34	0.64	0.70
RSR	Monthly	0.78	0.58	0.53
PBIAS, %	Monthly	−37	−36	−18

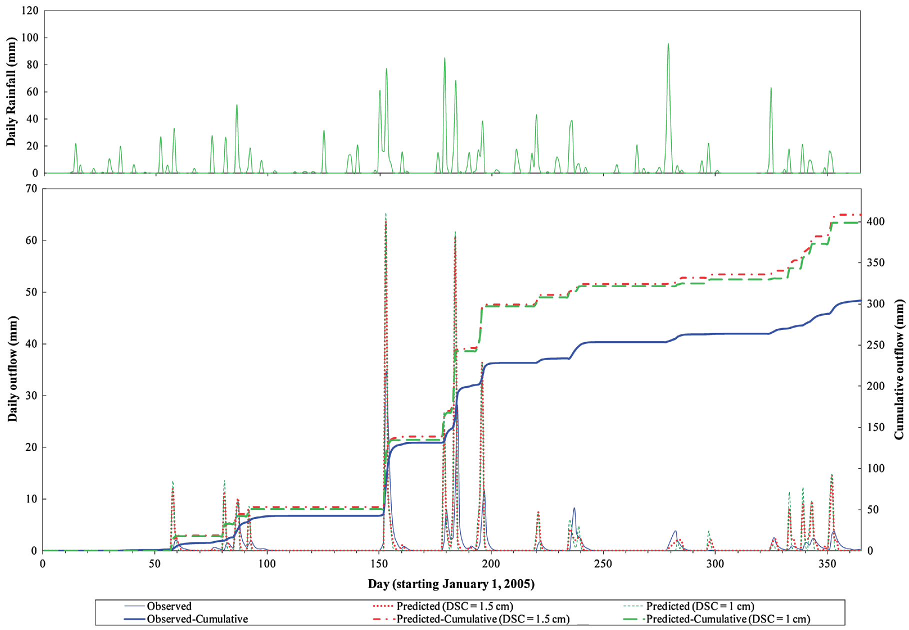

Figure 10. Observed and DRAINMOD predicted daily and cumulative stream outflows in watershed WS77 using optimization- and consistency zone–estimated DSC values of 1.5 and 1 cm, respectively, during the 2005 wet period validation. Daily rainfall is also shown in the top plot

than DSC_{wet}. This may be due to quick runoff generation as a result of the small depression storage capacities DSC_L within the watershed being filled relatively faster than large depressions DSC_{wet}. These observations are somewhat similar to other studies on hydrological connectivity, a key factor in generating runoff from forested watersheds (James and Roulet, 2007). The smaller the DSC, the faster will be the water movement and, consequently, the connectivity and vice versa. The slow rising and falling limbs of individual event hydrographs for DSC_{wet} indicate extra capacity of the depressions (5 mm) to store more excess rainfall for infiltration and evaporation, therefore yielding low flow peaks than DSC_L. The model overestimated annual stream outflows by 34% using DSC_{wet} (1.5 cm) and by 31% using DSC_L (1 cm). The difference between the observed and predicted average monthly means of outflows is less than 15 mm for both DSC values, suggesting that model predictions are reasonable for the period. Model performance using consistency-estimated DSC_L (1 cm) is similar to BF optimization–estimated DSC_{wet} (1.5 cm) during the wet period (Table III). This is consistent with the results of Kim *et al*. (2012), who found no effect of DSC values between 0.25 and 1.5 cm on event outflows for a coastal agricultural watershed in North Carolina.

For the dry period (2006), DRAINMOD is validated with streamflow using optimum DSC_{dry} (3.5 cm). Likewise, DSC_L (1 cm) is applied directly to DRAINMOD for the same period. Both DSC inputs overpredicted peak events, with DSC_L (1 cm) overshooting peaks relatively more than DSC_{dry} (3.5 cm) (Figure 11). Extreme oversimulation of peak flow rates using DSC_L may be caused by the rapid filling of its shallow surface depressions much faster as compared to DSC_{dry}, thereby causing rapid discharge to the watershed outlet. DRAINMOD overestimated annual stream outflows by 36% using DSC_{dry} and by 18% using DSC_L. The larger

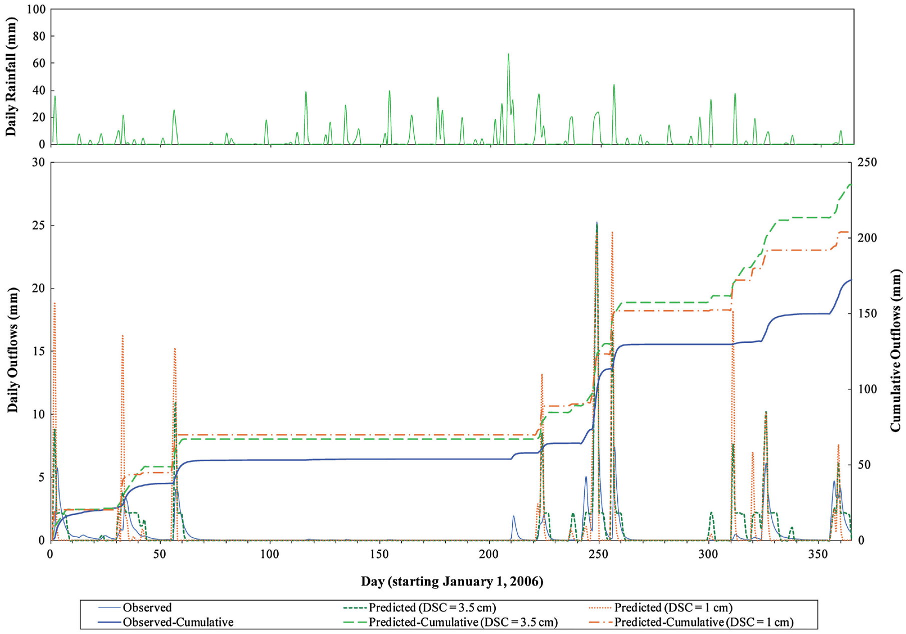

Figure 11. Observed and DRAINMOD predicted daily and cumulative stream outflows in watershed WS77 using optimization- and consistency zone–estimated DSC values of 3.5 and 1 cm, respectively, during the 2006 validation period. Daily rainfall is also shown in the top plot

annual overprediction for DSC_{dry} scenario despite the lower peak rate is due to sustained prolonged flows caused by higher saturation than the DSC_L in which the flow rate drops faster soon after the peak discharge due to lower saturation. This was evident from the water table predictions shown by Amoah (2008). The over-prediction in both cases, especially during the growing season with high ET demands, is likely due to underestimation of ET.

SUMMARY AND CONCLUSIONS

A consistency zone approach is established for estimating watershed-scale DSC for application in hydrologic and geomorphic models. The procedure was implemented to calculate the DSC of six forested watersheds varying in sizes between 150 and 5000 ha and located in low-gradient coastal plains of South Carolina. The methodology developed herein is tested for watersheds with landscape slopes less than 3% and DEM resolutions up to 30 m square grid. The process involved extracting geometric properties of the watersheds at various DEM grid resolutions and formulating an equation to calculate the DSC. Variability of the DSC values across different DEM grid resolutions led to the establishment of a consistency zone criterion for estimating the overall DSC for a given watershed. The calculated DSC values obtained in this study compared well with the values determined through a traditional model calibration approach.

Watershed WS77 was selected to validate the calculated DSC using techniques developed in this study. The BF optimization procedure is used to estimate the DSC by calibrating DRAINMOD, a quasi-process–based hydrologic model, with observed daily stream outflows for wet and dry periods. An RSR parameter is used for the optimization of prediction error in streamflow. Both the BF-optimized DSC and the consistency zone approach DSC values are applied to validate DRAINMOD with measured streamflow data. Although the results show a good agreement in predicting daily runoff events, only a satisfactory agreement was found between both DSC values in predicting daily streamflows. The results demonstrated robustness of DRAINMOD to changes in DSC values. Most of the discrepancies in overprediction of peak flows were attributed to the large area of the watershed to which the field-scale model was applied, without consideration of flow routing and spatial variability in rainfall. Furthermore, possible error in modelling ET, especially during the summer months with a high ET demand, including the use of grass-reference–based PET, may have also been factors in cumulative annual flow overprediction. The study revealed that the surface DSC is dependent on seasonal antecedent moisture conditions on surface topography, which is somewhat consistent with the findings of James and Roulet (2007). The authors reported that *priori* spatial patterns in shallow soil moisture in forested terrains may not always be a good predictor of critical hydrologic connectivity that leads to threshold change in runoff generation. Techniques developed in this study were applied only to forested watersheds in low-gradient coastal plains, which are known for gentle slopes and small surface depressional depths (less than 15 cm). Although the approach should be reproducible in watersheds with similar characteristics, further verification/refinement of the method using DEMs based on very high-resolution LiDAR data is suggested. The new methodology is particularly beneficial for estimating DSCs in situations where no data is available for model calibration, thus reducing the number of calibration parameters.

DOI: 10.1002/hyp

ACKNOWLEDGEMENTS

We would like to acknowledge the US Forest Service (USFS)—Southern Research Station (SRS) for the travel grant awarded through the Civil Rights Committee's Summer Student Hire Program. We thank USFS-SRS's Center for Forested Wetland Research and Tom Williams, PhD at the Baruch Institute of Coastal Ecology and Forest Science—Clemson University for the use of their data and facilities. We also would like to recognize Carl Trettin, PhD (Project Leader); Charles Harrison (Hydrologic Technician) at the Center for Forested Wetland Research; and Tim Callahan, PhD (College of Charleston), for their technical support.

REFERENCES

Abedini MJ, Dickinson WT, Rudra RP. 2006. On depressional storages: The effect of DEM spatial resolution. *Journal of Hydrology* **318**(1–4): 138–150.

Amatya DM, Chescheir GM, Fernandez GP, Skaggs RW. 2000. Effects of field sizes in watershed scale drainage modeling. In *2000 ASAE Annual International Meeting*, *ASAE Paper 00–2095*. ASAE: Milwaukee, WI.

Amatya DM, Miwa M, Harrison CA, Trettin CC, Sun G. 2006. Hydrology and water quality of two first order forested watersheds in coastal South Carolina. In *2006 ASABE Annual International Meeting*, *ASABE Paper No. 062182*. ASABE: St. Joseph, MI.

Amatya DM, Skaggs RW. 2001. Hydrologic modeling of pine plantations on poorly drained soils. *Forest Science* **47**(1): 103–114.

Amatya DM, Skaggs RW, Gregory JD. 1997. Evaluation of a watershed scale forest hydrologic model. *Agricultural Water Management* **32**(3): 239–258.

Amatya DM, Sun G, Skaggs RW, Trettin CC. 2003. Testing of DRAINMOD for forested watersheds with non-pattern drainage. In *2003 ASAE Annual International Meeting*, *ASABE Paper Number: 032046*. ASABE: St. Joseph, MI.

Amoah JKO. 2008. A new methodology for estimating watershed-scale depression storage. Ph.D. Dissertation, Florida A&M University, Tallahassee, FL.

Carvajal F, Aguilar MA, Aguera F, Aguilar FJ, Giraldez JV. 2006. Maximum Depression Storage and Surface Drainage Network in Uneven Agricultural Landforms. *Biosystems Engineering* **95**(2): 281–293.

Chescheir GM, Amatya DM, Skaggs RW. 1994. Modeling the hydrology of a natural forested wetland. In *1994 ASAE Annual International Meeting*, *ASAE Paper no. 942597*. ASABE: St. Joseph, Michigan.

Dai Z, Li C, Trettin CC, Sun G, Amatya DM, Li H. 2010. Bi-criteria evaluation of the MIKE SHE Model for forested watershed on the South Carolina coastal plain. *Hydrology and Earth System Sciences Discussions* **7**: 179–219.

Darboux F, Davy P, Gascuel-Odoux C. 2002. Effect of depression storage capacity on overland-flow generation for rough horizontal surfaces: water transfer distance and scaling. *Earth Surface Processes and Landforms* **27**(2): 177–191.

DHI. 2005. *MIKE SHE Technical Reference: Version 2005*. Danish Hydraulic Institute: Denmark

ESRI. 2009. *ArcGIS Desktop Help*. Environmental Systems Research Institute (ESRI): Redlands, Ca.

Gayle G, Skaggs RW. 1978. Surface storage on bedded cultivated lands. *Transactions of ASAE* **21**(1): 101–104, 109.

Hancock GR. 2005. The use of digital elevation models in the identification and characterization of catchments over different grid scales. *Hydrological Processes* **19**(9): 1727–1749.

Hansen B, Schjonning P, Sibbesen E. 1999. Roughness indices for estimation of depression storage capacity of tilled soil surfaces. *Soil Tillage Research* **52**(1/2): 103–111.

Harder SV. 2004. Hydrology and water budget of a first order forested coastal plain watershed, South Carolina. MS Thesis, College of Charleston, Charleston, SC.

Harder SV, Amatya DM, Callahan TJ, Trettin CC. 2006. Modeling the monthly water balance of a first order coastal forested watershed. In *International Conference on Hydrology and Management of Forested Wetlands*, Williams TM, Nettles JE (eds). ASABE Paper No. 701P0406, ASABE: St. Joseph, Michigan.

Harder SV, Amatya DM, Callahan TJ, Trettin CC, Hakkila J. 2007. Hydrology and water budget for a forested atlantic coastal plain watershed, South Carolina. *Journal of the American Water Resources Association* **43**(3): 563–575.

Hayashi M, van der Kamp G. 2000. Simple equations to represent the volume-area-depth relations of shallow wetlands in small topographic depressions. *Journal of Hydrology* **237**: 74–85.

Hutchinson MF. 1988. Calculation of hydrologically sound digital elevation models. In *Proceedings of the Third International Symposium on Spatial Data Handling, Sydney*. International Geography Union: Columbus, OH; 117–133.

James AL, Roulet NT. 2007. Investigating hydrologic connectivity and its association with threshold change in runoff response in a temperate forested watershed. *Hydrological Processes* **21**(25): 3391–3408.

Jenson SK, Domingue JO. 1988. Extracting topographic structure from digital elevation data for geographic information system analysis. *Photogrammetry Engineering and Remote Sensing* **54**: 1593–1600.

Kamphorst EC, Chadoeuf J, Jetten V, Guerif J. 2005. Generating 3D soil surfaces from 2D height measurements to determine depression storage. *Catena* **62**(2–3): 189–205.

Kamphorst EC, Duval Y. 2001. Validation of a numerical method to quantify depression storage by direct measurements on moulded surfaces. *Catena* **43**(1): 1–14.

Kamphorst EC, Jetten V, Guérif J, Pitkänen J, Iversen BV, Douglas JT, Paz A. 2000. Predicting depressional storage from soil surface roughness. *Soil Science Society of America Journal* **64**(5): 1749–1758.

Kim H, Amatya DM, Broome SW, Hesterberg DL, Choi M. 2012. Sensitivity analysis of the DRAINWAT model applied to an agricultural watershed in the lower coastal plain, North Carolina, U.S.A. *Water and Environment Journal* **26**(1): 130–145.

Legates DR, McCabe GJ. 1999. Evaluating the use of "goodness-of-fit" measures in hydrologic and hydroclimatic model validation. *Water Resource Research* **35**(1): 233–241.

Licciardello F, Rossi CG, Srinivasan R, Zimbone SM, Barbagllo S. 2011. Hydrologic evaluation of a Mediterranean watershed using the SWAT Model with multiple PET estimation methods. *Transactions of the ASABE* **54**(5): 1615–1625.

Linsley RK, Kholer MA, Paulhus JLH. 1949. *Applied Hydrology*. McGraw Hill: New York; 689.

Mark DM. 1988. Network Models in Geomorphology. In *Modelling in Geomorphological Systems*. John Wiley and Sons: Chichester, UK; 73–96.

Martin Y, Valeo C, Tait M. 2008. Centimetre-scale digital representations of terrain and impacts on depression storage and runoff. *Catena* **75**: 223–233.

Mitchell JK, Jones BA. 1976. Micro-relief surface depression storage: analysis of models to describe the depth-storage function. *Journal of the American Water Resources Association* **12**(6): 1205–1222.

Moore ID, Larson CL. 1979. Estimating micro-relief surface storage from point data. *Transactions of ASAE* **22**(5): 1073–1077.

Moriasi DN, Arnold JG, Liew MWV, Bingner RL, Harmel RD, Veith TL. 2007. Model Evaluation Guidelines for Systematic Quantification of Accuracy in Watershed Simulations. *Transactions of the ASABE* **50**(3): 885–900.

Muenich MR. 2011. The impacts of soil properties, subsurface drainage, and surface depressions on streamflow in a recently glaciated landscape. PhD Dissertation, Purdue University, Lafayette, IN.

Mwendera EJ, Feyen F. 1992. Estimation of depression storage and Manning's resistance coefficient from random roughness measurements. *Geoderma* **52**: 235–250.

Onstad CA. 1984. Depressional storage on tilled surfaces. *Transactions of ASAE* **27**(3): 729–732.

Planchon O, Darboux F. 2001. A fast, simple and versatile algorithm to fill the depressions of digital elevation models. *Catena* **46**: 159–176.

Richards P, Ryan G, Cannon D. 2005. Depression Storage in Land Uses Common to the Finger Lakes Region. In *First Annual Conference of the Finger Lakes Institute*. Finger Lakes Institute: Geneva, NY.

SCS. 1980. Soil Survey Report of Berkeley County, South Carolina. United States Department of Agriculture.

SCS. 1986. Urban hydrology for small watersheds. Technical Release 55. Chester, PA: U.S. Department of Agriculture

Singh J, Knapp HV, Demissie M. 2004. *Hydrologic modeling of the Iroquois River watershed using HSPF and SWAT*. ISWSCR2004-08. Illinois State Water Survey: Champaign, IL.

Skaggs RW. 1978. A water management model for shallow water table soils. Technical Report No. 134. Raleigh, NC: Water Resources Research Institute of the University of North Carolina, NC State University

Skaggs RW, Gilliam JW, Evans RO. 1991. A computer simulation study of pocosin hydrology. *Wetlands* **11**(Special Issue): 339–416.

Sneddon J, Chapman TG. 1989. Measurement and analysis of depression storage on a hillslope. *Hydrological Processes* **3**(1): 1–13.

Sun G, Noormets A, Gavazzi MJ, McNulty SG, Chen J, Domec JC, King JS, Amatya DM, Skaggs RW. 2010. Energy and water balance of two contrasting loblolly pine plantations on the lower coastal plain of North Carolina, USA. *Forest Ecology and Management* **259**(7): 1299–1310.

Tsihrintzis VA, Hamid R. 1998. Runoff quality prediction from small urban catchments using SWMM. *Hydrological Processes* **12**(2): 311–329.

Ullah W, Dickinson WT. 1979a. Quantitative description of depression storage using a digital surface model: I. Determination of depression storage. *Journal of Hydrology* **42**(1–2): 63–75.

Ullah W, Dickinson WT. 1979b. Quantitative description of depression storage using a digital surface model: II. Characteristics of surface depressions. *Journal of Hydrology* **42**(1–2): 77–90.

Viessman W, Jr., Lewis GL. 2003. *Introduction to Hydrology*. Prentice Hall: Upper Saddle River, NJ; 136–142.

Wu S, Li J, Huang GH. 2007. Modeling the effects of elevation data resolution on the performance of topography-based watershed runoff simulation. *Environmental Modelling and Software* **22**(9): 1250–1260.

Zhang W, Montgomery DR. 1994. Digital elevation model grid size, landscape representation, and hydrologic simulation. *Water Resources Research* **30**(4): 1019–1028.