

MODELING STREAMFLOW IN A SNOW-DOMINATED FOREST WATERSHED USING THE WATER EROSION PREDICTION PROJECT (WEPP) MODEL

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ABSTRACT. *The Water Erosion Prediction Project (WEPP) model was originally developed for hillslope and small watershed applications. Recent improvements to WEPP have led to enhanced computations for deep percolation, subsurface lateral flow, and frozen soil. In addition, the incorporation of channel routing has made the WEPP model well suited for large watersheds with perennial flows. However, WEPP is still limited in modeling forested watersheds where groundwater baseflow is substantial. The objectives of this study were to (1) incorporate nonlinear algorithms into WEPP (v2012.8) for estimating groundwater baseflow, (2) auto-calibrate the current and modified WEPP model using a model-independent parameter estimation tool, and (3) evaluate and compare the performance of the current version of WEPP without baseflow (WEPP-Cur) and the modified WEPP model with baseflow (WEPP-Mod) in simulating the hydrology of a snow-dominated watershed in the U.S. Pacific Northwest. A subwatershed of the Upper Cedar River Watershed in western Washington State was chosen for WEPP application and assessment. Simulations were conducted for two periods: 1997-2003 to calibrate the model and 2004-2011 to assess the model performance. The WEPP-Cur simulations resulted in Nash-Sutcliffe efficiency (NSE) and deviation of runoff volume (D_v) values of 0.55 and 24%, respectively, for the calibration period, and 0.60 and 21%, respectively, for the assessment period. The WEPP-Mod simulated streamflow showed improved agreement with observed streamflow, with NSE and D_v values of 0.76 and 6%, respectively, for the calibration period, and 0.74 and 2%, respectively, for the assessment period. The WEPP-Mod model reproduced hydrograph recessions during the low-flow periods and the general trend of the hydrographs, demonstrating its applicability to a watershed where groundwater baseflow was significant. The incorporation of a baseflow component into WEPP will help forest managers to assess the alterations in hydrological processes and water yield for their forest management practices.*

Keywords. *Baseflow, Forest watershed, Hydrological modeling, Streamflow, U.S. Pacific Northwest, WEPP.*

Snow-dominated mountainous watersheds are a principal source of freshwater for rivers and lakes (Christensen et al., 2008). Unlike small watersheds, large watersheds are often highly heterogeneous in climatic and physiographic settings (Singh, 1997). The complex interactions of climate with varying soils, vegetation, and geology in snow-dominated mountainous watersheds

can have a substantial effect on streamflow generation (Kuraš et al., 2008; Safeeq et al., 2013). Assessment and understanding of hydrology of such snow-dominated watersheds are important for sound management of aquatic ecosystems, and water supply and demands.

In undisturbed forests, surface runoff is rare due to the presence of groundcover, organic litter, and highly permeable soils that lead to enhanced infiltration (Elliot, 2013). However, surface runoff can occur as a return flow from subsurface lateral flow or groundwater baseflow at the bottom of hillslopes (Bachmair and Weiler, 2011). In mountainous forests, rapid subsurface lateral flow from the soil profile due to the presence of soil macropores and the preferential paths of tree root systems (especially decayed roots) contributes to the peak flows of a hydrograph, whereas baseflow from the underlying bedrock is indicated by a slow release of water (Fiori et al., 2007; Bachmair and Weiler, 2011) that contributes to gradual hydrograph recessions.

Baseflow is generated from water stored in shallow, and often unconfined, aquifers during dry seasons, represented by the recession part of the hydrograph. As the size of a watershed increases, baseflow plays an increasingly important role in surface water and groundwater interactions and in regulating streamflow, particularly during low-flow periods (Winter et al., 1998; Tague and Grant, 2009). The quantification of baseflow in areas where it substantially contributes

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to streamflow is crucial for monitoring and managing water resources.

Baseflow recession is reflected in streamflow when aquifer recharge stops, and surface storage, groundwater abstraction, and evapotranspiration have a negligible influence (Wittenberg, 2003). The relationship between groundwater in the aquifer and baseflow from the aquifer can be described by storage-outflow conceptual models (Moore, 1997; Wittenberg, 1999; Dewandel et al., 2003), which can be linear or non-linear, depending on the type of aquifer, watershed characteristics, soil properties, climate, and season (Dewandel et al., 2003).

Some earlier studies have supported the linear-reservoir model as a good approximation for baseflow recession (Zecharias and Brutsaert, 1988; Vogel and Kroll, 1992; Hornberger et al., 1998; Chapman, 1999). More recent studies also show the linear-reservoir approach to be adequate in representing baseflow recession (Fenicia et al., 2006; Brutsaert, 2008; van Dijk, 2010; Krakauer and Temimi, 2011). The linear reservoir model ($S = aQ_b$, where S is groundwater storage, Q_b is baseflow, and a is the recession constant) is acceptable as a simplification to describe the baseflow from aquifers. However, non-linear models sometime simulate aquifers better than linear models. Dewandel et al. (2003) and Chapman (1999) reported that non-linear relationships are appropriate for shallow unconfined aquifers, while linear relationships are appropriate for deep confined aquifers. The non-linear storage-outflow relationships reported by Moore (1997) and Wittenberg (1999, 2003) may be due to the presence of multi-reservoirs, floodplain storage, variations in rainfall, evapotranspiration, thickness of the aquifer (Nathan and McMahon, 1990), and a decrease in the saturated hydraulic conductivity of soil with depth (Wittenberg and Sivapalan, 1999). To simulate such complex surface water-groundwater processes that contribute to non-linearity in baseflow, a sound understanding of the study area's geology is required, but is often lacking. To overcome this complexity in simulating groundwater baseflow, Wittenberg (1999) suggested a non-linear storage-outflow relationship for unconfined aquifers.

The Water Erosion Prediction Project (WEPP) model is a physically based, continuous-simulation, distributed-parameter erosion prediction system (Flanagan and Nearing, 1995; Flanagan et al., 2007). The WEPP model is based on the fundamentals of hydrology, hydraulics, plant science, and erosion mechanics (Nearing et al., 1989). WEPP was originally intended for use in simulating hillslope profiles and small field- or farm-sized watersheds in cropland, rangeland, and disturbed forest areas. Most watershed applications have been on small or medium-scale catchments where streamflow is mainly from surface runoff (Baffaut et al., 1997; Liu et al., 1997; Amore et al., 2004; Pandey et al., 2008; Abaci and Papanicolaou, 2009).

The addition to WEPP of improved routines for subsurface lateral flow enhanced the model's applicability to simulate small forest watersheds where subsurface lateral flow is generated through preferential flow paths and perched water tables (Covert et al., 2005; Dun et al., 2009; Boll et al., 2015). However, WEPP applications to these studied forest watersheds have resulted in underprediction of streamflow,

primarily due to the neglect of groundwater baseflow (Dun et al., 2009; Zhang et al., 2009). Srivastava et al. (2013) and Brooks et al. (2016) showed good agreements between simulated and observed streamflow from watersheds in the Priest River Experimental Station and Lake Tahoe Basin, respectively, by bypassing stream channel algorithms and simulating daily streamflow using hillslope output from the WEPP model. Baseflow was simulated using a linear reservoir approach as a post-processing step from cumulative percolation losses from all hillslopes in the watershed. The most recent addition of channel routing algorithms to WEPP allows simulation of streamflow from watersheds with perennial flows (Wang et al., 2010).

The current WEPP model lacks a groundwater baseflow component. Therefore, incorporating a groundwater baseflow component into WEPP is necessary for applying the model to larger watersheds where there is a significant contribution of groundwater baseflow to streamflow. The objectives of this study were to (1) incorporate non-linear algorithms into WEPP (v2012.8) for estimating groundwater baseflow, (2) auto-calibrate the modified WEPP model using the model-independent nonlinear parameter estimation (PEST) technique (Doherty, 2005), and (3) evaluate and compare the performance of the modified WEPP model (with baseflow) and the current version of WEPP (without baseflow) using observed streamflow data from a snow-dominated watershed in the U.S. Pacific Northwest.

METHODS

WATERSHED DESCRIPTION

Our model assessment focused on the Upper Cedar River Watershed, located in the Cascade Mountains in King County of western Washington State (fig. 1). Known also as the Cedar River Municipal Watershed, it is a protected, mountainous, forest catchment that supplies drinking water to the city of Seattle. The Cedar River generally flows northwest into Cedar Lake (Wiley and Palmer, 2008). The 105 km² study watershed is upstream of Cedar Lake and ranges from 477 to 1655 m in elevation (USDA, 2014). The watershed receives 70% of its annual precipitation during October to March and the remaining 30% during April to September. Based on location, elevation, and aspect, the watershed can be characterized into three zones dominated by rain, mixed rain and snow, and snow (Wiley and Palmer, 2008; Elsner et al., 2010). Precipitation from these zones periodically feeds the river and results in two-peak streamflow hydrographs (fig. 2). The rain-dominant zone is located at lower elevations, creating the peak of streamflow during winter. The rain-snow transition zone contributes to both winter and spring-melt peaks depending on the type of precipitation and temperature. The snow-dominant zone primarily causes the spring-melt peak.

Based on the STATSGO map (USDA, 2012), there are two dominant soil series on the uplands above the USGS stream gauge (table 1, fig. 3a): Nimue (Andic Haplocryods, loamy-skeletal, mixed), a loamy sand that covers 56% of the watershed, and Altapeak (Andic Haplocryods, sandy-skeletal, mixed), a gravelly sandy loam that covers 15% of the

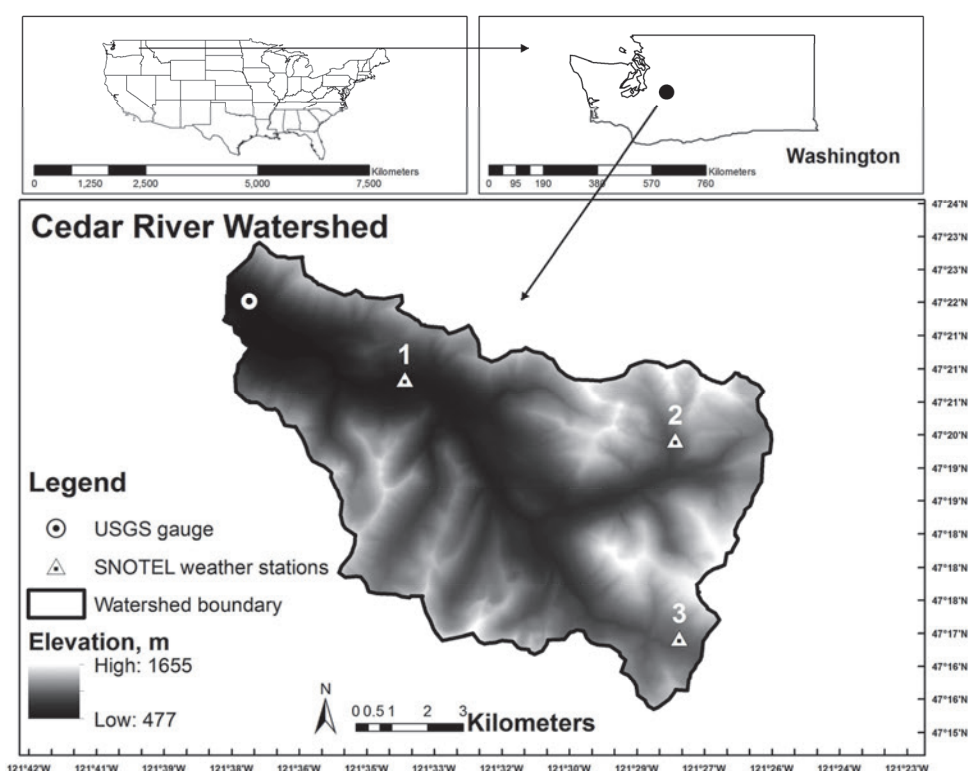


Figure 1. Location of the Cedar River Watershed, Washington State. Triangles represent SNOTEL weather stations (1 = Mount Gardner, 2 = Tinkham Creek, and 3 = Meadows Pass). The USGS gauge is the watershed outlet of the study watershed.

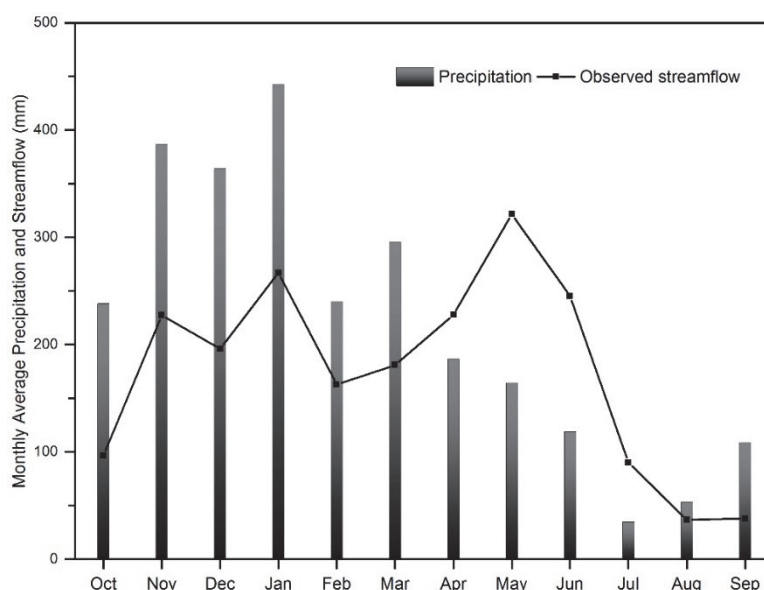


Figure 2. Monthly average precipitation and observed streamflow for 1997 to 2011 showing two-peak hydrograph resulting from rain, mixed rain and snow, and snow-dominant zones in the watershed.

watershed. Kaleetan (Typic Haplohumods, loamy-skeletal, mixed, frigid) sandy loam is typically found in the lowlands and covers 29% of the watershed (USDA, 2012) (fig. 3a). The upper soil layers are formed in a thin mantle of volcanic ash with paralthic materials (partially weathered bedrock or weakly consolidated bedrock such that roots cannot enter, except in cracks) at a 350 to 650 mm depth (USDA, 2012). The lower soil layers below the rooting zone, forming un-

consolidated weathered rocks in the shallow unconfined aquifer, are derived from pumice over colluvium: residuum originated from tuff breccia or igneous rock in Nimue soils, granitic and metamorphic rock in Altapeak soils, and tuff breccia and till or andesite and till in Kaleetan soils (USDA, 2012). The bedrock underlying the unconsolidated weathered rocks in the study area consists of intrusive igneous and metamorphic rocks of low porosity and permeability on the

Table 1. Soil characteristics in the study watershed.

Soil Series	Soil Type	Sand (%)	Clay (%)	Organic Matter (%)	CEC (meq per 100 g)	Rock (%)	Hydraulic Conductivity (mm h ⁻¹)
Nimue	Loamy sand	79.2	5.0	12.5	25.0	40	330
Kaleetan	Sandy loam	66.6	10.0	7.0	20.0	35	102
Altapeak	Gravelly sandy loam	66.6	10.0	7.5	7.5	30	102

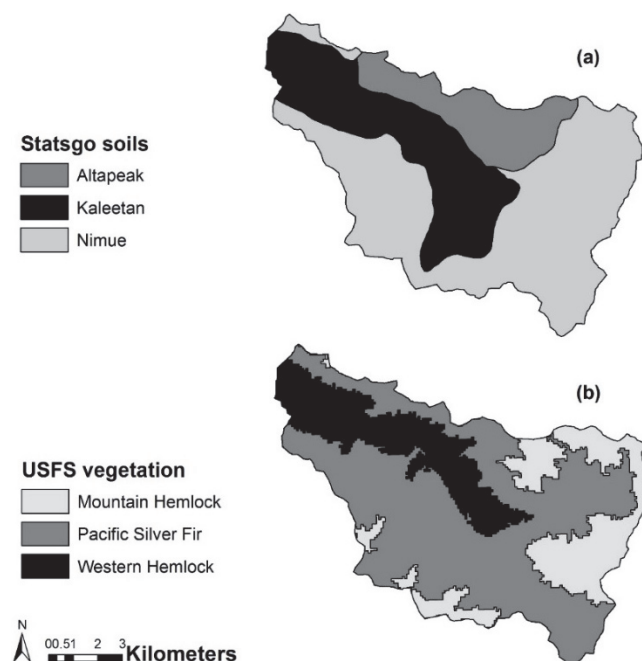


Figure 3. (a) Distribution of the STATSGO soil series and (b) USFS vegetation zones in the study watershed.

uplands (USGS, 2014) and the unconsolidated deposits (volcaniclastic deposits, andesite flows, and quaternary alluvium) in the lowland stream valley (Shellberg et al., 2010). The unconsolidated weathered materials above the bedrock on the uplands are adequately permeable for water to move laterally to recharge the unconsolidated-deposit aquifer that discharges water to nearby streams or lakes (McWilliams, 1955; USGS, 2014).

The watershed is vegetated by a combination of old-growth and second-growth conifer forests. The old-growth forest is 250 to 680 years old and is mostly found at higher elevations (Seattle, 2015). The second-growth forests at lower elevations are younger due to timber harvest in the 20th century (Seattle, 2015).

Vegetation cover in the watershed generally falls into three forest vegetation zones (fig. 3b). Franklin and Dyrness (1988) and Henderson et al. (1992) classified different vegetation zones by elevation and climate for western Washington and northwestern Oregon. In our study watershed, forests below 800 m a.m.s.l. are in the Western Hemlock Zone (19%) and consist mainly of western hemlock (*Tsuga heterophylla*) and Douglas fir (*Pseudotsuga menziesii*). This part of the watershed is rain-dominated (Rolf Gersonde, Seattle Public Utilities, personal communication, 2014). Forests at elevations between 800 and 1200 m a.m.s.l. lie in the Pacific Silver Zone (61%) and contain primarily Pacific silver fir (*Abies amabilis*) and western hemlock. This part of the watershed is snow-dominated with vegetation growing with greater density (>70% canopy closure) and intercepting ap-

proximately 60% of snow (Martin et al., 2013). Forests at elevations above 1200 m a.m.s.l. are in the Mountain Hemlock Zone (20%) and are dominated by Pacific silver fir and mountain hemlock (*Tsuga mertensiana*), which have less dense stands (<60% canopy closure) with substantially lower snow interception compared to those at low- and mid-elevations.

OBSERVED CLIMATE AND STREAMFLOW DATA

There are three NRCS SNOw TELelemetry (SNOTEL) weather stations located at higher elevations of the watershed that have available weather data dating back to 1996 (fig. 1). Weather data recorded at these stations include daily precipitation, daily maximum and minimum temperatures, and daily snow water equivalent (SWE). The annual precipitation ranges from 1727 to 3929 mm, averaging 2593 mm over 1996 to 2011. Based on the monthly observed data from the three SNOTEL stations, we found that precipitation generally increases with elevation (at about 0.13 mm m⁻¹) in winter and decreases with elevation (at about 0.01 mm m⁻¹) in summer (table 2). Mean air temperature decreases with elevation by 0.004°C m⁻¹ during both winter and summer. The observed monthly precipitation was lowest in July to August when the temperature was the highest, and it was greatest in December to January when the temperature was the lowest. Snowfall in the region is highly dependent on elevation, typically starting at mid-November, peaking around mid-April, and melting by the end of May or June. Comparison of measured snow at the three weather stations indicated an increase in snowfall with elevation. On average, 32% of precipitation in the region falls as snow.

The USGS gauging station (12115000, 47° 22' 12" N, 121° 37' 25" W) at the watershed outlet is immediately upstream of Cedar Lake and has been recording daily streamflow since 1945. The observed monthly streamflow is lowest in August to September and highest in May to June. The long-term (1945 to 2011) average annual runoff coefficient (ratio of streamflow to precipitation) is 0.85 (varies from 0.61 to 1.18 annually). The average seasonal runoff coefficient varies from 0.65 (range of 0.43 to 1.06) in winter (October to March) to 1.3 (range of 0.65 to 3.30) in spring and summer (April to September), indicating large spring snow-melt runoff.

WEPP MODEL

WEPP conceptualizes watersheds as a network of rectangular hillslopes and channels (Baffaut et al., 1997), the dimensions of which are defined by the stream channel network and average flow length along a specific stream reach. Winter processes, such as snow accumulation and melt as well as frost and thaw, are performed on an hourly basis internally in the model (Savabi et al., 1995a). Rainfall excess in a minute time step is computed as a difference between rainfall rate and infiltration capacity determined by the

Table 2. Monthly average precipitation and maximum and minimum temperatures for the three SNOTEL stations in the study watershed for 1997 to 2011 (P = precipitation, $T_{\max}$ = maximum temperature, and $T_{\min}$ = minimum temperature.).

Month	Mount Gardener			Tinkham Creek			Meadows Pass		
	P (mm)	$T_{\max}$ (°C)	$T_{\min}$ (°C)	P (mm)	$T_{\max}$ (°C)	$T_{\min}$ (°C)	P (mm)	$T_{\max}$ (°C)	$T_{\min}$ (°C)
1	412	2.2	-1.8	445	1.6	-3.0	417	2.9	-2.7
2	223	3.8	-1.4	237	3.6	-2.7	219	4.6	-2.9
3	293	5.9	-0.7	298	5.9	-1.8	271	6.3	-2.1
4	197	9.3	0.7	191	9.3	-0.6	165	9.2	-0.9
5	177	12.8	3.6	183	12.5	1.5	150	12.4	1.3
6	135	16.1	6.6	134	15.9	4.3	109	15.5	4.5
7	44	20.9	9.7	41	21.5	7.6	34	20.4	7.0
8	62	20.9	10.0	52	21.6	8.3	46	20.5	7.2
9	140	17.5	7.9	107	17.7	6.1	109	17.3	5.2
10	242	10.7	4.2	246	10.8	2.9	225	11.3	2.4
11	383	5.0	0.6	383	4.6	-0.4	364	5.6	-0.5
12	352	1.6	-2.0	364	1.1	-3.0	330	2.2	-2.9
Mean	2659	10.6	3.1	2683	10.5	1.6	2437	10.7	1.3

Green-Ampt Mein-Larson equation (Stone et al., 1995). Surface runoff is routed along the hillslope using kinematic wave equations or an approximate method, accounting for depression storage (Stone et al., 1995). Infiltrated water is redistributed in the soil layers via percolation, evapotranspiration (ET), and subsurface lateral flow on an hourly or daily basis. Percolation of water from upper layers to lower layers within the soil profile is estimated using storage-routing techniques (Savabi et al., 1995b). ET is determined using the Penman (1963), Priestly-Taylor (1972), or the FAO Penman-Monteith method (Savabi et al., 1995b; Allen et al., 1998). Subsurface lateral flow is computed following Darcy's law (Savabi et al., 1995c).

Surface runoff and subsurface lateral flow from each hillslope are routed through the downstream channel network to the watershed outlet. A water balance output file is generated by the model that includes simulated daily surface runoff, deep percolation, ET, subsurface lateral flow, and total soil water content for each hillslope and channel segment. Deep percolation in the current WEPP model (v2012.8) is the amount of the water that drains vertically out of the soil profile or the model domain. The absence of a groundwater baseflow component in the current WEPP model renders it difficult to assess water yield for watersheds where a significant amount of baseflow contributes to streamflow.

Incorporation of Baseflow Component in WEPP

Figure 4 shows the WEPP hillslope hydrologic processes with a baseflow component added. In WEPP, percolation through the soil layers is simulated on a daily basis when the soil water content exceeds the field capacity of the soil. Deep percolation in the current WEPP model (v2012.8) is the amount of water that drains out of the bottom of the soil profile or the model domain. Following the approach used by Srivastava et al. (2013) and Brooks et al. (2016), groundwater baseflow in this study was determined by assuming that this deep percolation recharges directly to an underlying shallow unconfined aquifer, adding to the groundwater storage. Groundwater discharge to the adjacent stream as baseflow is then estimated following Wittenberg (1999). To simulate baseflow for this study, we assumed: (1) the aquifer system underneath the watershed has uniform hydrogeolog-

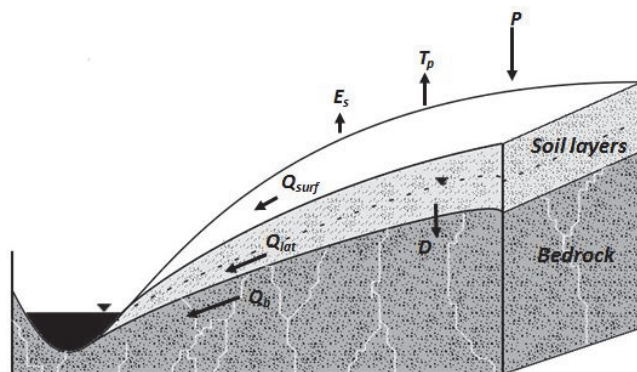


Figure 4. Schematic of hillslope hydrologic processes: P = precipitation, E_s = soil evaporation, T_p = plant transpiration, Q_{surf} = surface runoff, Q_{lat} = subsurface lateral flow, D = deep percolation, and Q_b = baseflow. The dashed line represents the water level. Deep percolation (D) in the current WEPP (v2012.8, indicated as WEPP-Cur in this article) is considered out of the model domain.

ical characteristics and is "watertight" (i.e., no leakage is allowed through the boundaries), (2) the deep percolation into an unconfined aquifer is independent of its storage and baseflow, and (3) evapotranspiration from groundwater is negligible due to the presence of gravely to extremely gravely textured soils below the rooting depth, which offer no upward capillary flow from groundwater.

A continuity equation for an unconfined aquifer (eq. 1), in combination with a storage-outflow relationship (eq. 2), was used to compute baseflow. The storage-outflow relationship (Wittenberg, 1999) states that the outflow from the groundwater reservoir is a non-linear function of the storage.

Successive groundwater storage values can be estimated using a finite forward differencing method as follows:

$$S_i = S_{i-1} + (D_i - Q_{b,i})\Delta t \quad (1)$$

where D is deep percolation [$L\ T^{-1}$], Q_b is baseflow [$L\ T^{-1}$], S is groundwater storage [L], Δt is the daily time interval [T], and i is time (days).

The nonlinear storage-outflow relationship as described by Wittenberg (1999) is:

$$S = aQ_b^b \quad (2)$$

where S is the hillslope groundwater storage as previously

defined (mm), Q_b is baseflow (mm d⁻¹), parameter a has dimensions of mm^{1-b} d^b, and exponent b is dimensionless. Parameter a represents the watershed storage characteristics, and parameter b describes the shape of baseflow recessions, with b equal to 1 for the special case of a linear reservoir.

By inverting equation 2, baseflow from the aquifer can then be computed as:

$$Q_b = \left(\frac{S}{a} \right)^{1/b} \quad (3)$$

The baseflow algorithms were coded into the water balance subroutine of the WEPP model following the deep percolation computations. The calculated baseflow values from each hillslope were output into the master watershed pass file, the file required by the WEPP watershed version before it performs runs for channel segments. The contributions of baseflow from adjacent hillslopes were added to the inflows of each channel and routed to the outlet of the same channel.

MODEL APPLICATION

WEPP Model Setup and Inputs

WEPP requires four main input files for both hillslope and channel elements: topography, climate, management, and soil. For WEPP simulations, we delineated the study watershed into hillslopes and channels (fig. 5) using GeoWEPP, a geospatial interface of WEPP (Renschler, 2003), from a 30 m digital elevation model (DEM) (USDA, 2014). GeoWEPP uses TOPAZ (Garbrecht and Martz, 1999) to generate drainage networks for the watershed and topographic variables, such as slope length and width, slope gradient, and aspect for hillslopes and channels, from a DEM (Garbrecht and Martz, 1999). The minimum-source channel length and critical source area were set to 500 m and 25 ha, respectively, in delineating the drainage network to match the National Hydrography Dataset streams (USDA, 2014). In all, the TOPAZ-discretized watershed consisted of 253 hillslopes and 103 channels (fig. 5), including 52 first-order, 21 second-order, 7 third-order, and 23 fourth-order channels, based on the Strahler channel link order scheme.

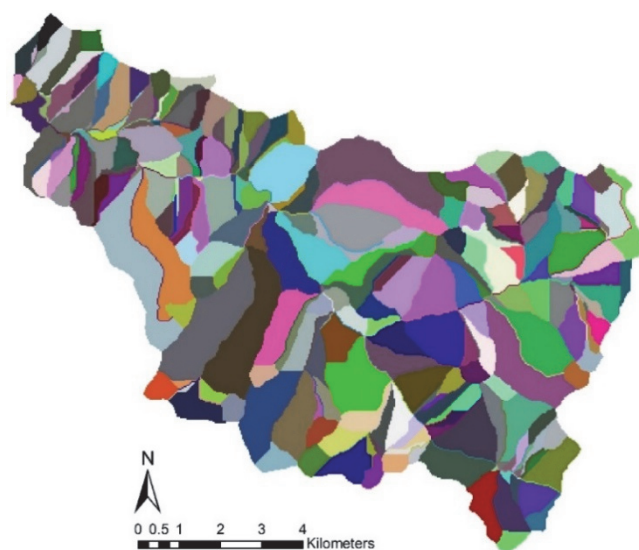


Figure 5. GeoWEPP delineated hillslopes and channel networks for the study watershed.

A WEPP climate input file requires daily precipitation, maximum and minimum temperatures, solar radiation, dew-point temperature, wind speed, and wind direction. To incorporate the topological effect of the mountainous watersheds for WEPP simulations, unique weather input files were developed for the centroid of each hillslope from gridded datasets. Daily precipitation amounts, solar radiation, and dew-point temperatures (computed from actual vapor pressure) were obtained from the 1000 m resolution Daymet gridded datasets (Thornton et al., 2014). Daily maximum and minimum temperatures were taken from the 800 m resolution TopoWx dataset (<http://www.ntsg.umt.edu/project/TopoWx>) with SNOTEL corrected temperature bias (Oyler et al., 2015). Daily wind speed and direction were acquired from the University of Idaho's Gridded-Surface meteorological 4000 m resolution dataset (Abatzoglou, 2013).

WEPP requires additional climate inputs that describe precipitation characteristics, including event total depth and duration, normalized peak intensity, and time to peak intensity. These inputs, excluding event depths, were stochastically generated using CLIGEN v5.3 (Nicks et al., 1995), an auxiliary program included in the WEPP program package, based on the monthly statistics of the long-term historical weather data from the NCDC weather station near Cedar Lake, Washington. For channels, the lowest-elevation climate data were used because only one set of climate inputs is allowed for channels in the current version of WEPP.

Soil and vegetative characteristics of each hillslope or channel element were assigned in GeoWEPP using ESRI ArcGIS 10.1. For this study, soil textural and hydraulic properties were extracted from the lower-resolution STATSGO map (USDA, 2012; fig. 3a, table 3) instead of the higher-resolution SSURGO map that had incomplete information on soil textural properties. USFS vegetation maps (Rolf Gersonde, Seattle Public Utilities, personal communication, 2014; fig. 3b) were used for determining vegetative characteristics.

The management inputs for hillslopes were based on the default values for a perennial forest in the WEPP database (table 3). The maximum canopy height was taken from Dale et al. (1986). The percentage canopy cover and the leaf area index (LAI) were computed in ESRI ArcGIS 10.1 from available raster data provided by Seattle Public Utilities (Rolf Gersonde, personal communication, 2014).

WEPP v2012.8 has three channel-routing methods (i.e., linear kinematic-wave, constant-parameter Muskingum-Cunge, and modified three-point variable-parameter Muskingum-Cunge) that are appropriate for large watersheds (Wang et al., 2010). For this study, we chose the linear kinematic-wave method, which is suitable for steep and long channels (Wang et al., 2010).

WEPP Model Parameterization and Calibration

All WEPP simulations were performed at a daily time step in a continuous mode. Daily streamflow records were separated into two periods: the first (January 1996 to September 2003) for model calibration that included nine months of a warm-up period (January 1996 to September 1996) and the second (October 2003 to September 2011) for

Table 3. Model parameters values used in WEPP.

Parameters	Initial Values	Parameter Range for PEST Calibration	PEST Calibrated Values	
			WEPP-Cur (without baseflow)	WEPP-Mod (with baseflow)
Rain-snow threshold (°C)	0	-1.0 to 1.5	0.5	0.5
Soil albedo	0.23	-	-	-
Initial saturation level (%)	80	-	-	-
Rill erodibility (s m^{-1})	5.0×10^{-4}	-	-	-
Interrill erodibility (kg s m^{-4})	4.0×10^5	-	-	-
Critical shear (Pa)	1.5	-	-	-
Anisotropy ratio	25	1 to 30	26	26
Saturated hydraulic conductivity of restrictive layer (mm h^{-1})	0.4	0.001 to 0.5	0.03	0.1
Leaf area index ($\text{m}^2 \text{m}^{-2}$)	7, 10, 9 ^[a]	-	-	-
Mid-season crop coefficient (K_{cb})	0.95	0.70 to 0.95	0.95	0.95
Readily available water coefficient (p)	0.70	-	-	-
Initial groundcover (%)	100	-	-	-
Initial canopy cover (%)	0.79, 0.63, 0.58 ^[a]	-	-	-
Canopy height (m)	69, 61, 46 ^[a]	-	-	-
Maximum root depth (m)	0.5	-	-	-
a ($\text{mm}^{1-b} \text{d}^b$) ^[b]	60.6	0 to 300	-	142
b (dimensionless) ^[c]	0.5	-	-	-

^[a] Three values are for western hemlock, Pacific silver fir, and mountain hemlock, respectively.

^[b] Parameter a represents the watershed storage characteristics.

^[c] Parameter b describes the shape of baseflow recessions.

model assessment. The calibrations of major parameters in the current version of WEPP without baseflow (WEPP-Cur) and the modified WEPP with baseflow (WEPP-Mod) were conducted separately. For model parameter calibration, the model-independent nonlinear parameter estimation package PEST (Doherty, 2005) was used in conjunction with WEPP to minimize the least square error between observed and predicted SWE, daily streamflow, and the daily logarithm of streamflow. These observation groups were equally weighted using the PEST weight adjustment tool. Following WEPP calibration, we assessed model performance using the same inputs for the second period of streamflow records.

The major parameters in WEPP controlling snow accumulation and melt, evapotranspiration (ET), soil water content, and groundwater baseflow were identified for calibration, and their initial values were either obtained from the literature or set to the default values for forest conditions (table 3). All these parameters were assumed uniform for the entire study watershed. The optimization of a particular parameter using PEST was limited to a predefined, expected range.

In the WEPP v2012.8 model, snow accumulation, snowpack density, and snowmelt are calculated on an hourly basis in the winter routines. For snow accumulation and melt computations, the daily climate inputs of precipitation, temperature, and solar radiation are downscaled to hourly values within WEPP (Savabi et al., 1995a). The equation for estimating hourly values from given daily values were taken from de Wit et al. (1978) for temperatures and from Swift and Luxmoore (1973) and Jensen et al. (1990) for solar radiation. Daily precipitation is internally disaggregated into hourly values within WEPP using a random number generator. The density of freshly fallen snow is assumed to be 100 kg m^{-3} . Precipitation is partitioned into rain and snow using a single threshold value ($T_{\text{rain/snow}} = 0^\circ\text{C}$ as the initial value). Hourly precipitation is determined as rain when the hourly air temperature is greater than the threshold and as snow otherwise. The depth and density of the snowpack are adjusted for new falling and drifting snow, snow settling, and snowmelt. The settling of the snowpack is computed

when the density of the snowpack is below 250 kg m^{-3} . Snowmelt is computed from the generalized basin snowmelt equation for open areas developed by the U.S. Army Corps of Engineers (USACE, 1956) and modified by Hendrick et al. (1971) for mountainous forest conditions. The hourly snowmelt is a sum of snowmelt caused by solar radiation, longwave radiation, convection-condensation, and warm rain. The albedo of melting snow is assumed to be 0.5. Snowmelt occurs when the daily average temperature is greater than 0°C and snow density exceeds 350 kg m^{-3} . For this study, $T_{\text{rain/snow}}$ was the only snow parameter we considered to calibrate to represent snow accumulation. A uniform value of $T_{\text{rain/snow}}$ across the watershed was optimized against the three observed SWE at the SNOTEL sites.

For this study, we used the FAO Penman-Monteith method (Allen et al., 1998) for ET computations because the “bulk” surface resistance concept describes the resistance of vapor flow both through the transpiring crop and from the evaporating soil surface (Allen et al., 1998). In this method, actual ET is computed from the potential ET of a grass reference crop based on vegetation growth and environmental conditions. The mid-season vegetation coefficient (K_{cb}), the ratio of the actual crop ET over the reference ET, and the fraction (p) of the total available water in the root zone that can be depleted before a plant experiences moisture stress are critical parameters to determine actual ET. Recommended K_{cb} and p values for conifers are 0.95 and 0.70, respectively (Allen et al., 1998). However, Allen et al. (1998) reported that for forests in well-watered conditions, the K_{cb} value could easily be below 0.95. Therefore, we calibrated K_{cb} and kept the value of p at 0.70.

The soil anisotropy ratio, which represents the dominance of lateral over vertical hydraulic conductivity of soil, was initially set to the default value of 25 for steep forest slopes and was further calibrated for the study watershed (table 3). The saturated hydraulic conductivity of the restrictive layer (K_{sat}) for basaltic-andesite volcanic bedrock was reported by Belcher et al. (2002) as ranging from 0.0017 to 250 mm h^{-1} with a geometric and arithmetic mean of 2.5 and 62.5 mm h^{-1}

for volcanoclastic sedimentary rocks and younger volcanic rocks, respectively. Todd and Mays (2005) reported a value of 0.42 mm h^{-1} for basalt rocks. We set the initial K_{sat} value to 0.4 mm h^{-1} (table 3) and used the PEST model to calibrate this parameter.

The aquifer storage-discharge during the low-flow periods of observed streamflow exhibited non-linearity, as indicated by the recession curve of $-dQ/dt$ versus the average Q on a log-log plot (fig. 6). Brutsaert and Nieber (1977) proposed a power law function that describes the rate of change in discharge as a function of discharge:

$$-\frac{dQ}{dt} = cQ^d \quad (4)$$

In equation 4 and with reference to the recession curve relationship in figure 6, the value of exponent d is 1.5, and the value of c is $0.033 \text{ mm}^{-0.5} \text{ d}^{-0.5}$ (estimated as $10^{-1.48}$). Substituting $S = aQ^b$ into $Q = -dS/dt$ (the equation of continuity during recession, neglecting evapotranspiration), we obtain:

$$-\frac{dQ}{dt} = \frac{1}{ab} Q^{2-b} \quad (5)$$

Comparing equation 5 with the values of coefficients found for equation 4, we have the estimated values of a and b as $60.6 \text{ mm}^{0.5} \text{ d}^{0.5}$ and 0.5, respectively.

Sánchez-Murillo et al. (2014) found non-linear baseflow responses for a series of streamflow gauging stations in 26 inland Pacific Northwest watersheds. The non-linear behavior of streamflow results from the complex interactions of climate, vegetation, topography, and soil and geologic characteristics (Wittenberg, 1999; Sánchez-Murillo et al., 2014). For groundwater baseflow simulations in WEPP, we assumed no ET loss from groundwater for the study watershed. We set the estimated value of 0.5 for parameter b in equation 3 and calibrated the initial estimated value of parameter a to ensure adequate outflow volume. The initial groundwater storage of 230 mm was estimated by averaging the groundwater storage at the beginning of each year from the preliminary WEPP run.

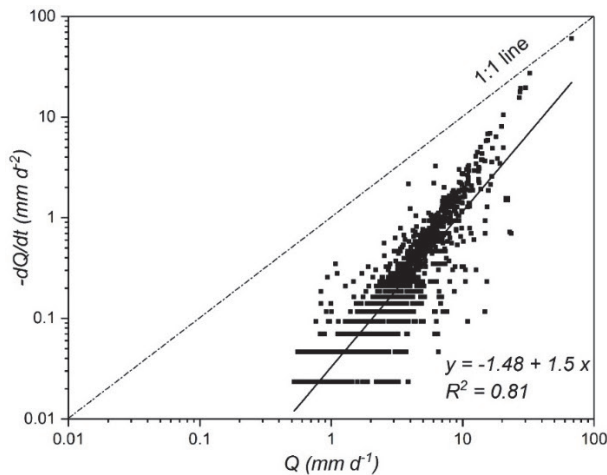


Figure 6. Recession plot on a log-log scale of $-dQ/dt$ versus the average Q indicating non-linearity in observed streamflow recessions. A linear streamflow recession response would be indicated by a best fit line with a slope of 1.

The WEPP model maintains a continuous water balance on a daily basis following the equation:

$$RM - ET - Q - (\Delta TSW + \Delta GW) = 0 \quad (6)$$

where RM is rain plus snowmelt, ET is evapotranspiration, ΔTSW and ΔGW are the changes in soil water and groundwater, respectively (calculated as the current day's value minus the previous day's value), and Q is total simulated streamflow, which is the sum of surface runoff (Q_{surf}), subsurface lateral flow (Q_{lat}), and baseflow (Q_b).

PERFORMANCE EVALUATION

To assess the performance of the modified WEPP model, we calculated two commonly used model performance indices for simulated and observed daily SWE and streamflow discharge for both the model calibration and assessment periods. These indices are the Nash-Sutcliffe efficiency (NSE; Nash and Sutcliffe, 1970) and the deviation of runoff volume (D_v ; Gupta et al., 1999). NSE, a goodness-of-fit criterion, is given by:

$$NSE = 1 - \left[\frac{\sum_{i=1}^n (Y_{obs,i} - Y_{sim,i})^2}{\sum_{i=1}^n (Y_{obs,i} - Y_{mean})^2} \right] \quad (7)$$

where $Y_{obs,i}$ and $Y_{sim,i}$ are the i th observed and simulated values, respectively, Y_{mean} is the mean of the observed data, and n is the total number of observations. NSE ranges from $-\infty$ to 1. A value of 1 indicates a perfect fit between simulated and observed data, whereas a value of 0 suggests that the model results are no better than the mean observed value. A negative value indicates that the observed mean value is better than the predicted values.

The percent deviation of runoff volume (D_v) is computed as:

$$D_v = \left[\frac{\sum_{i=1}^n (Y_{obs,i} - Y_{sim,i}) \times 100}{\sum_{i=1}^n Y_{obs,i}} \right] \quad (8)$$

The D_v value reflects model accuracy in terms of over- or underprediction of the observed values (Gupta et al., 1999). A value of zero indicates that the total volume of overpredicted runoff on some days is the same as the total volume of underpredicted runoff on other days. Positive values signify model underprediction, and negative values signify overprediction.

The lack of hydrogeological information (e.g., well construction, groundwater levels) in our study watershed prevented us from independently corroborating the storage of the groundwater reservoir and the baseflow coefficients. To evaluate the low-flow conditions, we compared simulated and observed streamflow recessions during the low-flow periods (August to September). In addition, we used the flow duration curve technique to examine the simulated low flows and compared them with the observed flow duration curves. The simulated and observed daily streamflows ($\text{m}^3 \text{ s}^{-1}$) were ranked in descending order. We then used the Weibull formula to compute the probability of exceedance of flows:

$$P = \frac{m}{(N+1)} \times 100 \quad (9)$$

where N is the total number of simulation days, and m is the rank of the daily streamflow series.

RESULTS AND DISCUSSION

SNOW WATER EQUIVALENT

Daily comparisons of WEPP-simulated and SNOTEL-observed SWE for 1996 to 2011 are shown in figure 7. For the periods of model calibration (1996 to 2003) and model performance assessment (2004 to 2011), WEPP generally underpredicted SWE. For the period of model calibration, NSE for SWE ranged from 0.89 to 0.95 for the three SNOTEL sites, averaging 0.92, and D_v ranged from -1% to 23%, averaging 8%, indicating underprediction. For the period of model performance assessment, NSE varied from 0.77 to 0.88, averaging 0.83, and D_v ranged from 2% to 28%, with an average of 17%, suggesting underprediction (fig. 7). The trend of simulated SWE, in general, matched that of observed SWE during the calibration period. However, during the model assessment period, simulated SWE underpredicted the observed SWE, particularly in the years 2006, 2007, and 2008. During these three years, the model underpredicted the observed peaks by 43%, 25%, and 25%, respectively (fig. 7).

A close examination of the WEPP-simulated and observed daily SWE for the Tinkham Creek SNOTEL site as impacted by the daily average air temperature, snowpack density, and cumulative snowmelt for the water year 2006 is shown in figures 8a and 8b. The WEPP-simulated SWE generally follows the observed trend except for the periods of mid-November and early January to early February, when the daily average air temperature fluctuates around 0.5°C, the threshold temperature at which WEPP partitions precipitation into rain and snow. During these periods, the WEPP-simulated snowpack density was consistently below 350 kg m⁻³, which prevented snowmelt from the snowpack (fig. 8b). During the same peri-

ods, the daily average air temperature (fig. 8a) generally was above the threshold temperature. This caused the WEPP model to partition precipitation more into rain than snow (fig. 8b), causing WEPP to underestimate snow accumulation and resulting in underprediction of simulated SWE.

The large discrepancies in simulated and observed SWE in figures 7 and 8 following 2003 were due to replacement of the standard temperature sensors with extended-range temperature sensors. The temperature sensors were replaced at the Meadows Pass SNOTEL site in July 2003 and at Tinkham Creek and Mount Gardener in May 2005. When analyzing temperature data at these three SNOTEL sites, we found the average air temperature of 1.9°C during winter (November to April) for the model assessment period to be 0.7°C higher than that for the model calibration period (1.2°C), indicating substantial temporal changes in air temperature. The temperature sensors were replaced by the NRCS at Pacific Northwest SNOTEL sites to capture temperature readings to -40°C (Julander, 2011). Julander (2011) showed that the new extended-range temperature sensors recorded 1°C to 2°C warmer than the standard sensors at the Trinity Mountain SNOTEL site in Idaho. Such inconsistency in temperature recording was observed at all the Idaho SNOTEL sites (Julander, 2011).

In this study, we used daily temperatures derived from the TopoWx gridded-temperature dataset that was corrected for the systematic inconsistencies in high-elevation SNOTEL temperatures when homogenized with the low-elevation NCDC stations. However, Oyler et al. (2015) reported that such a homogenization technique would not correct the strong seasonal dependencies within the SNOTEL temperature bias. Moreover, the authors stated that homogenization could also have the potential to remove unique microclimate phenomena occurring at the SNOTEL sites. This untimely change in temperature measurement methods between the model calibration and assessment periods of this study likely contributed to the underprediction of snow accumulation and SWE during the model assessment period.

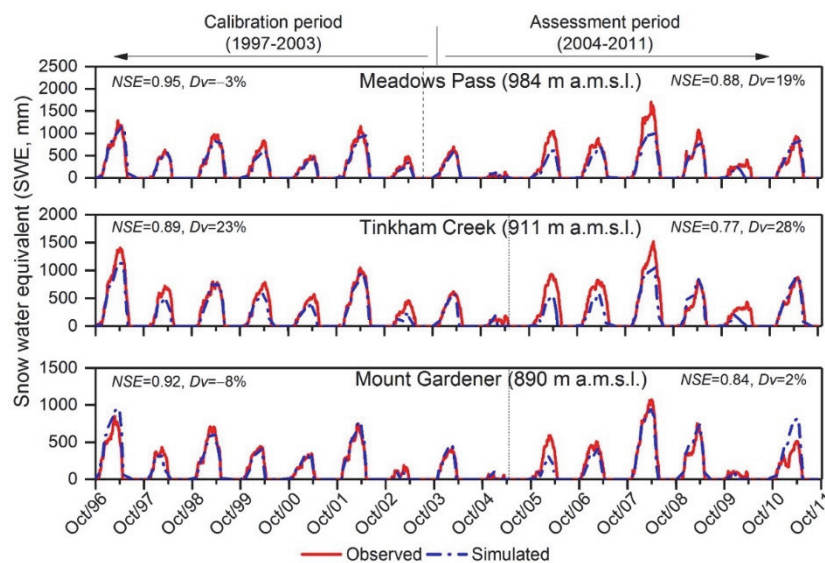


Figure 7. Comparison of observed and WEPP-simulated daily SWE at three SNOTEL sites for the model calibration (1997 to 2003) and assessment (2004 to 2011) periods. Vertical lines indicate the period of replacement of standard temperature sensors with extended-range temperature sensors at the SNOTEL sites (July 2003 at Meadows Pass; May 2005 at Tinkham Creek and Mount Gardener).

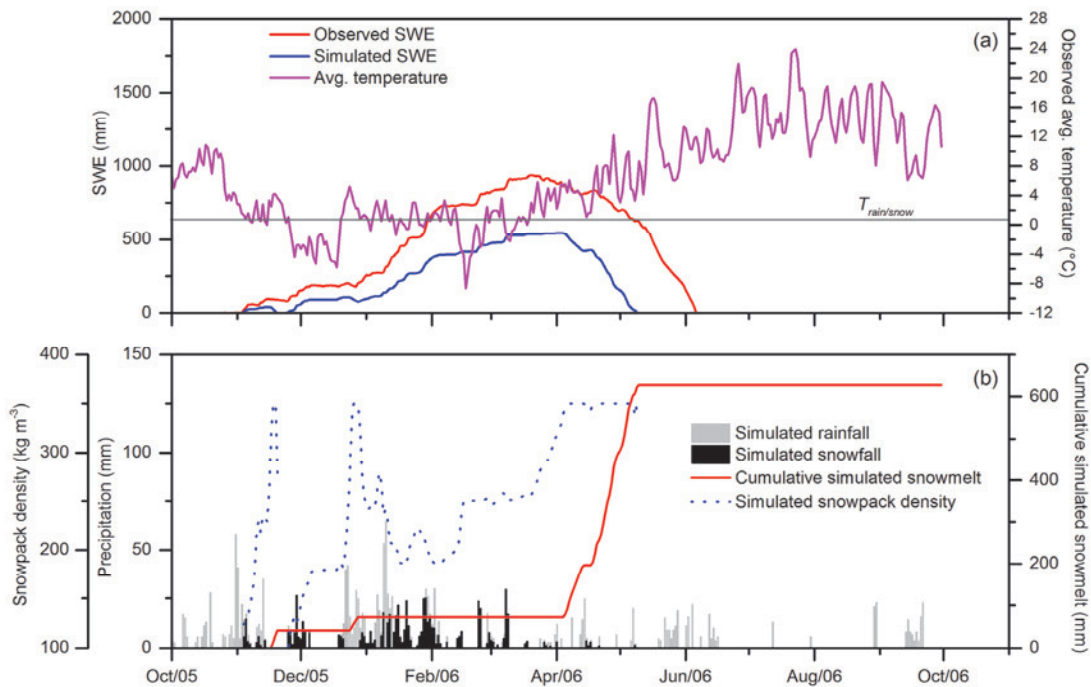


Figure 8. Simulated snow accumulation and melt for Tinkham Creek SNOTEL site for water year 2006: (a) comparison of observed and simulated daily SWE along with daily average temperature, and (b) daily precipitation, daily simulated snowpack density, and cumulative snowmelt.

STREAMFLOW

Figure 9 shows the WEPP-simulated hydrographs on a logarithmic scale without baseflow (WEPP-Cur) and with baseflow (WEPP-Mod) in comparison with the observed streamflow for the model calibration and assessment periods. With WEPP-Mod, the simulated streamflow included surface runoff, subsurface lateral flow, and baseflow,

whereas with WEPP-Cur, the simulated streamflow included surface runoff and subsurface lateral flow. The WEPP-Cur simulations show low streamflow discharge compared to observed streamflow because of the absence of baseflow (fig. 9a). However, WEPP-Mod generated improved streamflow discharge during low-flow periods, comparable with that of the observed hydrograph (fig. 9b).

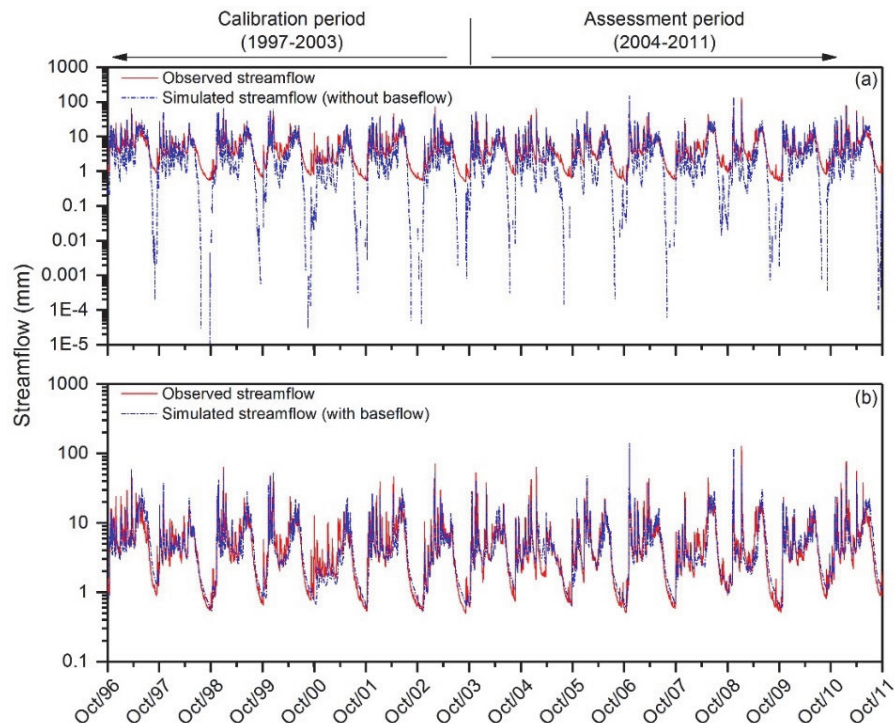


Figure 9. Comparison of observed and simulated hydrographs on a logarithmic scale (a) without baseflow (WEPP-Cur) and (b) with baseflow (WEPP-Mod) for the model calibration (1997 to 2003) and assessment (2004 to 2011) periods.

The NSE and D_v for streamflow with WEPP-Mod and WEPP-Cur for both the calibration and assessment periods are presented in table 4. Overall, the NSE and D_v from the WEPP-Mod simulations were much improved over those from the WEPP-Cur simulations. On average, the NSE was 0.58 with WEPP-Cur and 0.75 with WEPP-Mod, indicating improved simulation of streamflow discharge by the latter model over the entire period. The higher D_v of 22% with WEPP-Cur represents a substantial underestimation of streamflow compared to the D_v of 4% with WEPP-Mod. The low values of NSE and high values of D_v from WEPP-Cur are indicative of the need to incorporate a baseflow component into the WEPP model for these types of larger catchment simulations.

The NSE with WEPP-Mod for the calibration period ranged from 0.55 to 0.85, averaging 0.76, and D_v varied from -8% to 13%, averaging 6% (table 4). For the model assessment period, NSE ranged from 0.66 to 0.85 with an average of 0.74, and D_v ranged from -2% to 8% with an average of 2%. The positive values of D_v for both the model calibration

and assessment period suggest that WEPP-Mod slightly underpredicted the streamflow for the study watershed.

GROUNDWATER BASEFLOW

Overall, the simulated baseflow from WEPP-Mod matched observed streamflow reasonably well during the low-flow conditions (fig. 9b). Daily groundwater storage volume and the simulated proportion of streamflow composed of baseflow, surface runoff, and subsurface lateral flow for water year 2005 are shown in figure 10. Daily simulated baseflow fluctuated between 0.6 and 4 mm, following the trend of simulated groundwater storage that varied from 111 to 275 mm. The delayed simulated baseflow peaked after subsurface lateral flow declined; after that, baseflow receded until the initiation of the next storm or snowmelt event. The model predicted that the streamflow hydrograph was composed primarily of baseflow. During the low-flow periods of August and September, the streamflow was composed nearly entirely of baseflow.

Table 4. Observed and simulated annual streamflow, Nash-Sutcliffe efficiency (NSE), and deviation of runoff volume (D_v) with baseflow (WEPP-Mod) and without baseflow (WEPP-Cur).

Water Year	Observed Streamflow (mm)	WEPP-Mod			WEPP-Cur		
		Simulated Streamflow (mm)	NSE	D_v (%)	Simulated Streamflow (mm)	NSE	D_v (%)
1997	2798	2872	0.55	-8	2427	0.41	13
1998	1699	1699	0.82	3	1278	0.27	25
1999	2399	2344	0.76	5	1973	0.52	18
2000	2244	2012	0.69	13	1638	0.59	27
2001	1331	1125	0.60	12	826	0.34	38
2002	2498	2240	0.65	8	1860	0.44	26
2003	1591	1435	0.85	12	1109	0.77	30
2004	2049	1869	0.75	7	1557	0.49	24
2005	1508	1496	0.69	-2	1040	0.63	31
2006	1998	1908	0.81	0	1550	0.59	22
2007	2258	2237	0.71	7	1865	0.62	17
2008	2230	2092	0.66	3	1711	0.61	23
2009	2157	2012	0.70	8	1666	0.61	23
2010	1784	1809	0.85	-1	1443	0.31	19
2011	2774	2824	0.73	-2	2346	0.62	15
Calibration (1997-2003)	2030	1961	0.76	6	1587	0.55	24
Assessment (2004-2011)	2095	2031	0.74	2	1647	0.60	21
Overall (1997-2011)	2062	1998	0.75	4	1619	0.58	22

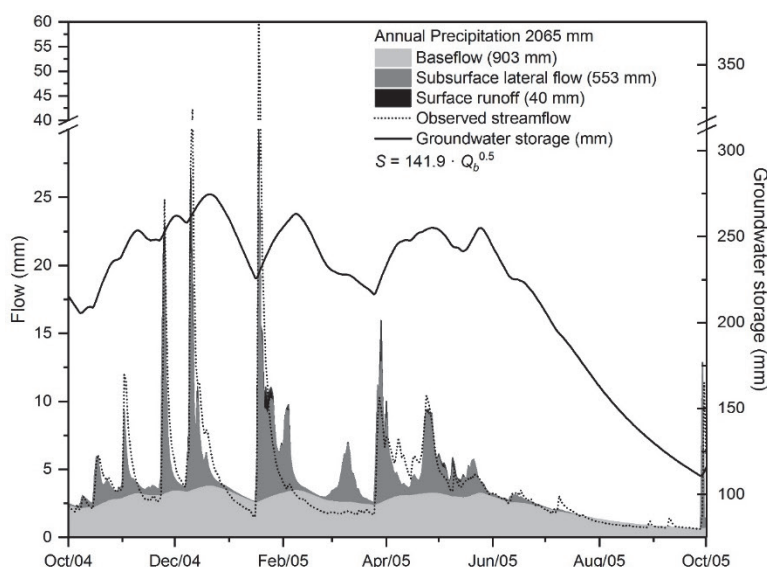


Figure 10. Cumulative watershed hydrological response for the water year 2005 from WEPP-Mod.

The simulated streamflows with WEPP-Mod and WEPP-Cur during the low-flow recession periods (August to September) were compared with the observed streamflow (fig. 11). The overall NSE of 0.79 for WEPP-Mod shows improved stream discharge compared to the NSE of -0.74 for WEPP-Cur. Streamflow volume simulated with WEPP-Mod matched the observed streamflow more closely than streamflow simulated with WEPP-Cur, as indicated by the D_v values. On average, WEPP-Mod overestimated streamflow by 14%, whereas WEPP-Cur underestimated streamflow by 74%.

Figure 12 suggests that the WEPP-Mod daily streamflows are comparable to the observed streamflow at all probabilities of exceedance. In contrast, the WEPP-Cur daily streamflows are underpredicted due to the absence of baseflow at lower probabilities of exceedance.

For this study, a value of 0.5 was determined for the exponential parameter b in equation 3, based on the recession analysis (fig. 6), to describe the rate and dynamics of the baseflow recession as affected by aquifer properties. Wittenberg (1999) found that the distribution of b ranges between 0 and 1.1 for 80 gauging stations in northern Germany and suggested that a value of 0.5 for b could be most appropriate for shallow unconfined aquifers. In a recent study, Sánchez-Murillo et al. (2014) examined 26 watersheds that varied in size from 6 to 6,500 km² in the western U.S. and determined that b ranged from 0.7 to 1.2. The authors found lower b values in steeper forested watersheds and higher b values in flatter, semi-arid forested watersheds.

The PEST-estimated value of parameter a for this study was 141.9 mm^{0.5} d^{0.5}. The value of a indicates the groundwater

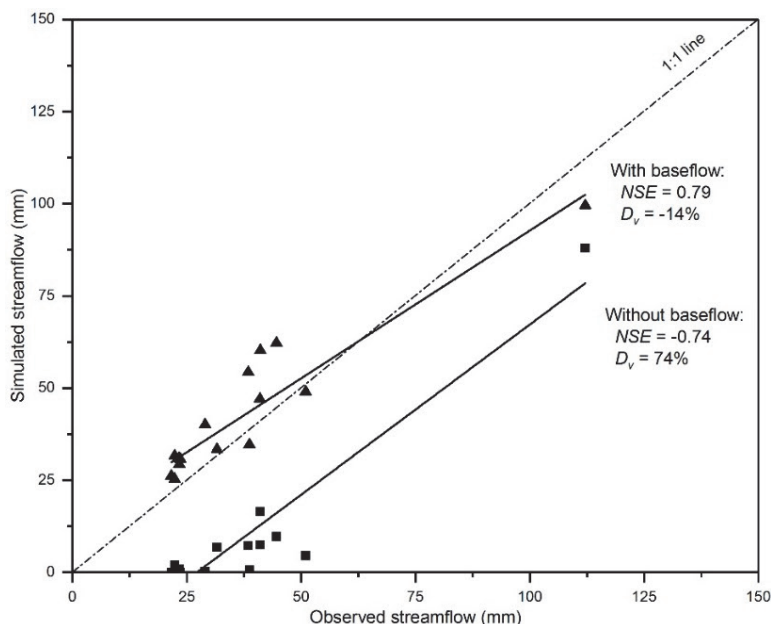


Figure 11. Comparison of observed and simulated streamflow with baseflow (WEPP-Mod) and without baseflow (WEPP-Cur) for low flow periods of August and September months. Each point denotes the total sum of yearly streamflow during August and September for 1997 to 2011.

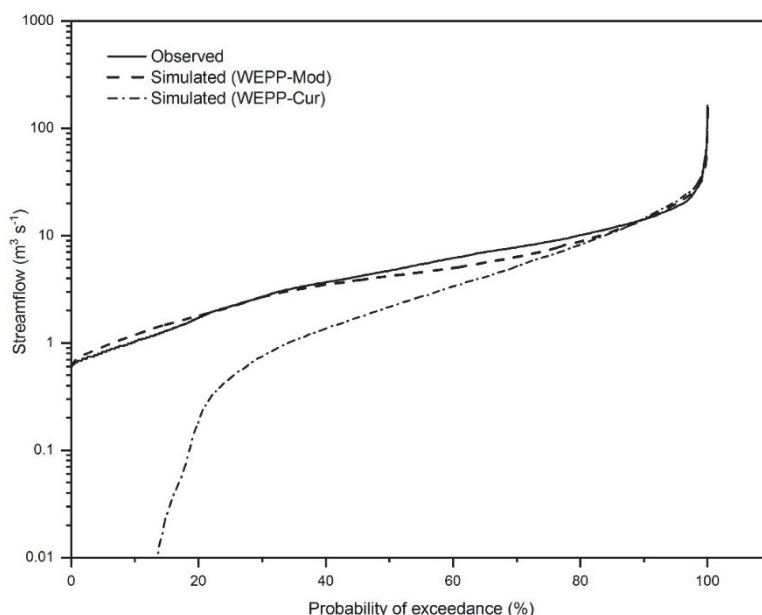


Figure 12. Probability of exceedance of observed and simulated daily streamflow for 1997 to 2011.

storage capacity, which depends on the watershed size and aquifer characteristics, and therefore exhibits a wide range of variation (Wittenberg, 1999; Sánchez-Murillo et al., 2014). The baseflow coefficients obtained in this study for the Cedar River Watershed fall into the ranges reported in the literature.

WATER BALANCE

Major water balance components for the whole watershed for the WEPP-Mod model calibration and assessment periods are presented in table 5. The mean annual precipitation and rain plus snowmelt during 1997 to 2011 were 2620 and 2625 mm, respectively. The simulated streamflow was 76% of the precipitation on an average annual basis. The simulated surface runoff from individual hillslopes was 11% of the total simulated streamflow. Simulated subsurface lateral flow ranged from 533 to 1411 mm, averaging 927 mm or 46% of annual simulated streamflow over the entire simulation period. The remainder of the streamflow (43%) was from baseflow.

Annual ET for the Cedar River Watershed varied from 541 to 686 mm, averaging 625 mm or 24% of mean annual rain plus snowmelt. Link et al. (2004) reported simulated ET (528 mm) as 21% of annual precipitation (2467 mm) for old-growth coniferous trees, exceeding 450 years in age and 60 m in height, located 405 km south of the Cedar River Watershed and with a mean annual temperature of 8.7°C. In a study conducted on the eastern Cascades in British Columbia, 415 km northeast of the Cedar River Watershed, Whitaker et al. (2003) reported 314 mm of simulated ET for coniferous forests (tree heights ranging from 12 to 36 m) receiving 1498 mm of average annual precipitation with a mean annual temperature of 5°C. In another study conducted on the east coast of Vancouver Island, British Columbia, 225 km northwest of the Cedar River Watershed, Jassal et al. (2009) measured ET for

58, 19, and 7 year old coniferous forests (mean tree heights of 33, 7.5, and 2.4 m, respectively) using the eddy covariance technique. The corresponding measured yearly ET values were 27% (404 mm), 25% (409 mm), and 18% (274 mm) of annual precipitation, ranging from 1450 to 1608 mm, at a mean annual temperature of 8.3°C. The simulated ET as a percent of annual precipitation in our study is comparable with literature values considering the similarities in climate.

The WEPP-simulated soil water normally started increasing in September following the dry summer months typical of a Mediterranean climate. This wetting of the soil profile occurred from winter rains and spring snowmelt and typically filled the soil profile around April. In mid-June, soil water declined and remained low throughout the dry months of August and September (fig. 13). The simulated soil water varied from year to year, with an accumulation of 80 mm in 2010, a deficit of 116 mm in 1998, and a net decrease of 1 mm over the entire simulation period (table 5). That the soil may continually sustain vegetation growth while carrying a soil water deficit from one year to the next shows the importance of understanding long-term weather patterns. Drought conditions may need several years to develop in subalpine forests and, once developed, can lead to reduced forest health and increased fire risk (Schoennagel et al., 2004).

The simulated deep percolation into groundwater storage followed the trend of the simulated soil water (fig. 13), i.e., the wetter the soil, the greater the deep percolation to the aquifer. Much of the simulated deep percolation to the aquifer occurred during the wet periods of the winter-spring season. No deep percolation was simulated during the summer season. The simulated annual percolation from the soil profile varied from 569 to 1127 mm, with an average of 853 mm (32% of mean annual rain plus snowmelt, table 5). In forested areas with shallow soils having high hydraulic conductivity, soil

Table 5. Annual water balance from WEPP-Mod for 1997 to 2011 water years.^[a]

Water Year	<i>P</i> (mm)	RM (mm)	REF-ET (mm)	ET (mm)	<i>D</i> (mm)	ΔTSW (mm)	ΔGW (mm)	<i>Q_{surf}</i> (mm)	<i>Q_{lat}</i> (mm)	<i>Q_b</i> (mm)	BFI (%)
1997	3629	3631	796	686	1127	34	38	371	1411	1089	38
1998	2084	2088	832	559	844	-116	-54	90	710	897	53
1999	3039	3044	787	677	888	-3	27	319	1162	862	37
2000	2712	2716	806	642	799	74	-13	218	982	811	40
2001	1727	1729	803	673	569	-62	-7	68	481	576	51
2002	2885	2891	821	646	900	-8	12	267	1084	888	40
2003	2011	2017	901	541	684	56	-15	113	622	699	49
2004	2670	2675	812	671	946	32	105	113	912	843	45
2005	2065	2070	780	644	801	31	-104	40	553	903	60
2006	2436	2444	800	593	790	-58	2	204	916	787	41
2007	2835	2842	755	610	846	-6	0	370	1020	847	38
2008	2774	2776	745	670	835	-10	26	292	989	810	39
2009	2590	2593	806	626	685	-20	-25	394	908	709	35
2010	2494	2501	709	558	993	80	54	66	803	940	52
2011	3348	3353	708	587	1085	-32	-27	355	1356	1111	39
Calibration (1997-2003)											
Mean value	2584	2588	821	632	830	-3	-1	207	922	832	42
Percentage of RM	-	-	32	24	32	0.1	0.1	8	36	32	-
Assessment (2004-2011)											
Mean value	2652	2657	764	620	873	2	4	229	932	869	43
Percentage of RM	-	-	29	23	33	0.1	0.1	9	35	33	-
Overall (1997-2011)											
Mean value	2620	2625	791	625	853	-1	1	219	927	852	43
Percentage of RM	-	-	30	24	32	0.0	0.1	8	35	32	-

^[a] *P* = precipitation, RM = rain plus snowmelt, REF-ET = reference evapotranspiration for grass crop, ET = evapotranspiration, *D* = deep percolation from the soil profile, ΔTSW and ΔGW = yearly change in soil water and groundwater, respectively (calculated as the last day's value minus the first day's value), *Q_{surf}* = surface runoff, *Q_{lat}* = subsurface lateral flow, *Q_b* = baseflow from groundwater storage, and BFI = baseflow index (baseflow in percent streamflow). The sum of *Q_{surf}*, *Q_{lat}*, and *Q_b* forms the total streamflow.

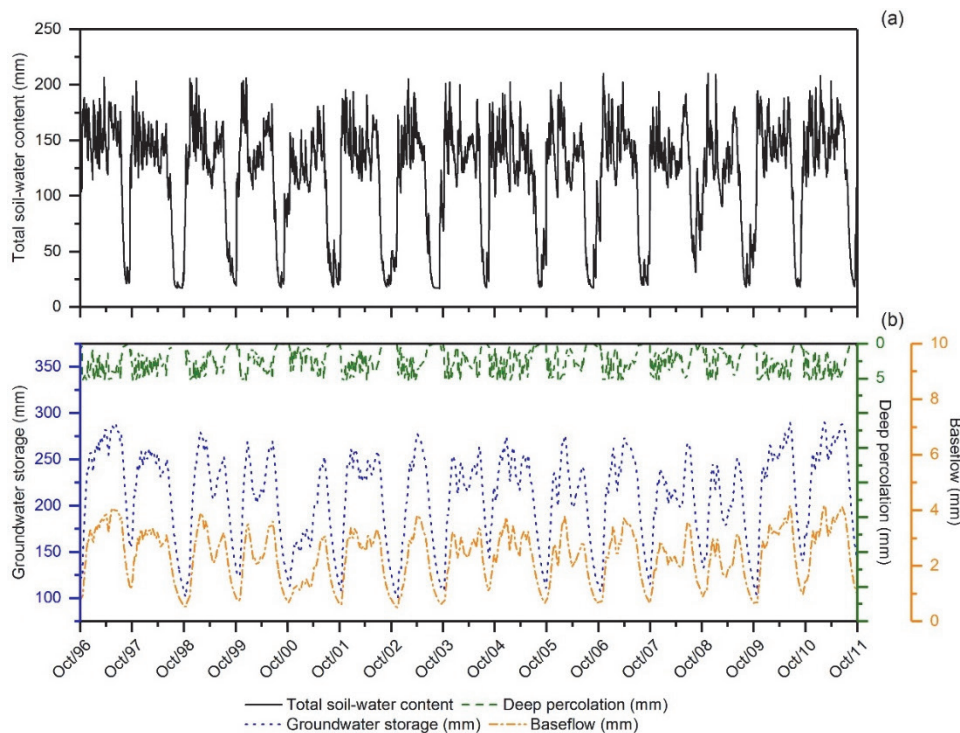


Figure 13. WEPP-Mod simulated hillslope-averaged (a) total soil-water content and (b) deep percolation, groundwater storage, and baseflow (1997 to 2011).

water may readily make its way to the underlying bedrock. Anderson et al. (1997) and Asano et al. (2002) studied forest watersheds in the state of Oregon and in central Japan and determined that water movement was predominantly vertical from the soil profile to the bedrock and then lateral through the bedrock to the streams. Tromp-van Meerveld et al. (2007), in a rainfall simulation study, found that more than 90% of the applied water was lost as deep percolation into steep forested hillslopes in the state of Georgia. Terajima et al. (1993) observed that 30% and 18% of annual precipitation percolated into the bedrock at two forested watersheds in Japan.

The WEPP-simulated annual baseflow varied from 576 to 1111 mm and formed the second largest water balance component, accounting for 32% of the mean annual rainfall plus snowmelt. The annual BFI (baseflow in percent streamflow) ranged from 35% to 60%, averaging 43% for the simulation period.

The assumption that there was no net evapotranspiration from the baseflow is appropriate for the study watershed where the steep slopes and coarse-textured soils are not conducive to transporting deep water into the rooting zone. For watersheds with ET loss from groundwater, users can model an unconfined aquifer by defining an effective soil depth based on an estimate of the depth to the consolidated bedrock layer. Boll et al. (2015) and Brooks et al. (2016) showed a better representation of the spatial distribution of ET, subsurface lateral flow, and return flow processes with multiple OFEs in a hillslope profile.

SUMMARY AND CONCLUSIONS

The WEPP model was applied to a subwatershed of the

Upper Cedar River Watershed (105 km²) in Washington State to simulate the hydrologic processes of a mountainous forest watershed in the U.S. Pacific Northwest. A baseflow component was implemented directly into the WEPP model to estimate groundwater contributions to streamflow. WEPP was calibrated using observed snow-water equivalent and streamflow data for 1997 to 2003. The PEST model was used to auto-calibrate the major parameters in WEPP with and without baseflow, and the model performance was evaluated for 2004 to 2011 using the same PEST-optimized parameter values. Simulated streamflows without baseflow (WEPP-Cur) and with baseflow (WEPP-Mod) were compared against observed streamflow.

For the entire simulation period (1997 to 2011), the observed mean annual streamflow averaged 2062 mm, while the simulated streamflow was 1998 mm from WEPP-Mod and 1619 mm from WEPP-Cur. The hydrograph peaks and baseflow recessions were better simulated by WEPP-Mod than by WEPP-Cur, indicating the need for a baseflow component in WEPP for watersheds of this size and type. Simulated annual hydrograph flow components from WEPP-Mod showed that, on average, contributions of surface runoff, subsurface lateral flow, and baseflow to streamflow were 11%, 46%, and 43%, respectively. An overall NSE of 0.75 and D_v of 4% from the WEPP-Mod simulations demonstrated the model's applicability to a snow-dominated forested watershed with substantial groundwater baseflow. With the incorporation of a baseflow component into WEPP, future research may include evaluation of the dominant hydrological pathways in watersheds to improve forest management.

Future modeling efforts may include a sensitivity analysis of the major WEPP parameters, an analysis of uncertainties

in ET and the range of the vegetative coefficients for different tree species, and application of the WEPP model to watersheds with hydrogeological information. Researchers have found that the precipitation phase is closely linked to humidity and that the dew-point and wet-bulb temperatures are better indicators of the precipitation phase during mixed rain-snow events. As such, future studies could be pursued in implementing this method in WEPP and evaluating the model performance for snow accumulation and melt.

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