Predicting Discharge and Erosion for the Abay (Blue Nile) with a Simple Model

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ABSTRACT

Models accurately representing the underlying hydrological processes and sediment dynamics in the Nile Basin are necessary for optimum use of water resources. Previous research in the Abay (Blue Nile) has indicated that direct runoff is generated either from saturated areas at the lower portions of the hill slopes or from areas of exposed bedrock. Thus, models that are based on infiltration excess processes are not appropriate. Furthermore, many of these same models are developed for temperate climates and might not be suitable for monsoonal climates with distinct dry periods in the Nile basin. The objective of this study is to develop a simple hydrology and erosion models using saturation excess runoff principles and interflow processes appropriate for a monsoonal climate and a mountainous landscape. We developed a hydrology model using a water balance approach by dividing the landscape into variable saturated areas, exposed rock and hillslopes. Water balance models have been shown to simulate river flows well at five day or longer intervals when the main runoff mechanism is saturation excess. The hydrology model was developed and coupled with an erosion model using available precipitation and potential evaporation data and a minimum of calibration parameters. This model was applied to the Blue Nile. The model predicts direct runoff from saturated areas and impermeable areas (such as bedrock outcrops) and subsurface flow from the remainder of the hillslopes. The ratio of direct runoff to total flow is used to predict the sediment concentration by assuming that only the direct runoff is responsible for the sediment load in the stream. There is reasonable agreement between the model predictions and the ten-day observed discharge and sediment concentration at the gauging station on Blue Nile upstream of Rosaries Dam at the Ethiopia-Sudan border

Key words: Model, Erosion, Sedimentation, Rainfall-runoff, monsoonal climate

INTRODUCTION

The Abay (Blue Nile) River in Ethiopia contributes significant flow and sediment to the Nile River. Thus, a better understanding of the hydrological processes, erosive losses, and sedimentation mechanisms in the various watersheds in the headwaters of the Nile River is of considerable importance. There is a need to improve and augment current resource management and development activities in areas with heavy degradation and low productivity, particularly in Ethiopia, where it is generally believed that only five percent of surface water is utilized (Weiß and Schaldach,2008). There is a particular need to develop further existing hydropower and irrigation potential of the Abay (Blue Nile) for socio-economic development in Ethiopia, while maintaining sustainable operation of water infrastructure systems downstream in Sudan and Egypt. Sustainable operation is dependent in large part on preventing silting up of reservoirs. This paper focuses on characterizing the rainfall-runoff-sediment relationships for the Ethiopian portion of the Blue Nile River. The majority of the sedimentation of rivers in the basin occurs during the early period of the rainy season and peaks of sediment are consistently measured before peaks of discharge for a given rainy season. Typical erosion models based on stream power would predict the greatest concentration to occur when the velocity and discharges are at their maximum (e.g., SWAT, AnnAGNPS, GWLF). Thus, innovative models are called for to predict erosion and sedimentation that are consistent with the hydrology of the region. Once developed, these models can be used for managing and or mitigating the sedimentation of newly constructed reservoirs.

A review by Awulachew et al. (2009) shows that the number of models simulating the discharge from watersheds in the Blue Nile and other river basins in Ethiopia and Africa has increased exponentially in recent years. Most of these models were originally developed for applications in temperate regions. They range from relatively simple engineering approaches such as the Rational Method (Desta 2003), to more complex models such as SWAT (Setegn et al. 2008) the Precipitation Runoff Modeling System (PRMS) (Legesse et al. 2003), Water Erosion Prediction Project (WEPP) (Zeleke 2000), the Agricultural Non-Point Source model (AGNPS) (Haregeweyn and Yohannes 2003; Mohammed et al. 2004), and water balance approaches (Ayenew and Gebreegziabher 2006; Kim and Kaluarachchi 2008). Implementation of these models yielded mixed results. For example, AGNPS was tested in the highlands of Ethiopia on the Augucho Catchment but could not reproduce observed runoff patterns. PRMS was similarly tested by Legesse et al. (2003) for South Central Ethiopia, and needed extensive calibration to predict the monthly runoff. It should not have been surprising that the above mentioned models are not performing well because they are based on the SCS curve number approach, of which the parameter values are obtained statistically from plot data in the USA with a template climate. The watershed behavior in a temperate climate than in a monsoonal climate where during the dry period the soil dries out completely, something that does not happen in the USA. Statistical methods are only valid for conditions that they are tested for.

Many simple water balance type approaches have been attempted for the Nile Basin. Both Mishra et al (2006) and Conway (1997) developed useful results with grid-based water balance models for the Blue Nile Basin using monthly discharge data from the El Deim Station in Sudan, located close to the Ethiopian border. They were studying the spatial variability of flow parameters and the sensitivity of runoff to changes in climate. Using a water balance model Kebede et al. (2006) concentrated on Lake Tana and developed a water balance utilizing relatively long durations (>30 years) of data for precipitation, evaporation, inflows of major tributaries and outflows to the Blue Nile. The simple water balance models often perform better especially over monthly time steps than their more complicated counterparts that have many more calibration parameters, but are not without problems either because different parameter sets are required for different basin sizes in the Blue Nile Basin as shown by Kim and Kaluarachchi (2008). One of the weaknesses of Kebebe et al. (2006) was that they did not differentiate between the hills and valleys in their simplified model.

To model the hydrology realistically the conceptual framework for the model should be correct. According to Liu et al. (2008), saturation excess runoff from saturated areas dominates the runoff process in several watersheds in the Ethiopian highlands. Subsequent field visits showed that runoff was produced from exposed hardpan and bedrock as well. Runoff from these almost impermeable areas can be modeled with either saturation excess models with a very small amount of retention before runoff occurs or infiltration excess models with a minimal infiltration capacity. Water balance models are consistent with the above-mentioned type of runoff processes, since the runoff can be related to the available watershed storage capacity and the amount of precipitation but not generally to the precipitation intensity. Moreover, as described above models developed and intended for use in temperate regions where rainfall is generally well distributed throughout the year do not perform well in regions with monsoonal rainfall distributions (Liu et al., 2008). Therefore, water balance models, that track soil moisture levels and the degree of saturation, often perform better than more complicated models in Ethiopian type landscapes (Johnson and Curtis, 1994; Conway, 1997; Kebede et al., 2006; Liu et al., 2008).

Despite the copious literature on runoff and hydrology in the Nile Basin, there are very few erosion models published in the refereed literature for Ethiopia. Haregeweyn and Yohannes (2003) applied AGNPS model Augucho catchment and predicted the sediment loads for this small watershed with some success. The Universal Soil Loss Equation (USLE) was calibrated for Ethiopian conditions by Haile et al. (2006). Tamene and Vlek employed the USLE together with sediment deposition routine. Other approaches use expert judgment in the erosion predictions (Feoli et al. 2002; Sonneveld, 2003; Nyssen et al. 2007). Some publications use erosion assessments as part of an economic evaluation of soil and water conservation practices (Hengsdijk et al, 2005; Okumu et al 2004; Shiferaw and Holden 2000) however this practice is not without controversy, as the erosion estimates are, at best, a subject to dispute (Nyssen et al. 2006). Models predicting soil loss for large watersheds do not exist in the refereed literature (Hurni et al. 2005), thus there is a need to develop and test erosion models for larger scales. These erosion models necessarily need to be based on the proper hydrology. Only then, can the drastic land use changes that have occurred during the last 30 years in Ethiopia (as documented by Zeleke, 2000 and Zeleke and Hurni, 2001) be modeled and analyzed successfully.

Since saturation excess runoff is the dominate runoff production mechanism from the low laying areas and rock outcrops in the Ethiopian highlands and most models are based on infiltration excess runoff mechanisms, these models do not always perform well. Thus, a more realistic model needs to be developed. Consequently, the objective of this study is to develop a physically based runoff and sediment loss model (using mainly existing data sources as input data with a minimum of calibration parameters) that is based on the saturation excess runoff process and is valid for monsoonal climates. We expect by using the correct conceptual hydrological model, scaling issues will be minimized.

MODEL DEVELOPMENT

In this section, we develop simple water balance type hydrology and erosion model. The hydrology model assumes that overland flow is generated from saturated areas in the relatively flatter areas in the landscape and areas where bedrock is exposed. The remainder of the landscape mainly is assumed to have sufficiently high conductivity so that rainfall infiltrates and is lost subsequently as evaporation, interflow or base flow. The erosion model predicts sediment concentrations based on the assumption that interflow and base flow are sediment free and that the sediment is carried by overland flow. The model, therefore, directly uses the input from the hydrology model for the calculations of the sediment concentrations.

A similar water balance type rainfall-runoff model was developed and tested by Collick et al. (2008) to predict the stream flow for four relatively small watersheds (< 500 ha) in the Blue Nile Basin. The authors reported reasonable predictions on a daily or weekly time step using nearly identical parameters for watersheds hundreds of kilometers apart. In this paper, some minor modifications were made with respect to interflow generation for predicting the discharge of the entire Blue Nile. For clarity, we will present the complete watershed water balance model and add a simple erosion model. Model parameters to predict the discharge were initially set to that of Collick et al. (2008). Model parameters for discharge were then refined and sediment parameters calibrated with the discharge and sediment concentration measured at the Ethiopia-Sudan border gauge station for 1993-1994 water year and then validated with data available for 2003 and 2004 rainy season.

Predicting direct runoff, interflow and base flow

The watershed is divided into two sections, the hillslopes, and the relatively flatter areas that become saturated during the rainfall season (Figure 1). The hillslopes are divided in two regions that have either restricted infiltration and storage or have high percolation rates (McHugh, 2006) and most of the water is transported subsurface as interflow (e.g. over a restrictive layer) or base flow (percolated from the soil profile to deeper soil and rock layers). The flatter areas that drain the surrounding hillslopes become runoff source areas when saturated (Figure 1). These areas can usually be recognized during the rainfall season as wet areas under permanent grass cover located near a stream. Evapotranspiration is extracted from a root zone. On the high infiltration hill slope areas excess water in the root zone is percolated through the subsoil. On the exposed hardpan or bedrock and in the saturated contributing areas, all excess water becomes surface runoff.

The amount of water stored in the topmost layer (root zone) of the soil, *S* (mm), for hillslopes and the runoff source areas were estimated separately with a water balance equation of the form:

$$S = S_{t-\Delta t} + \left(P - AET - R - Perc\right)\Delta t \qquad (1)$$

where *P is* precipitation, (mm d^{-1}); *AET* is the actual evapotranspiration, (mm d^{-1}), $S_{t-\Delta t}$, previous time step storage, (mm), *R* saturation excess runoff (mm d^{-1}), *Perc* is percolation to the subsoil (mm d^{-1}) and Δt is the time step.

During wet periods when the rainfall exceeds potential evapotranspiration, *PET* (i.e., $P>PET$), the actual evaporation, *AET*, is equal to the potential evaporation, *PET*. Conversely, when evaporation exceeds rainfall (i.e., $P<PET$), the Thornthwaite and Mather (1955) procedure is used to calculate actual evapotranspiration, *AET* (Steenhuis and van der Molen, 1986). In this method *AET* decreases linearly with moisture content, e.g.:

$$\mathrm{AET} = \mathrm{PET}\left(\frac{\mathrm{S_t}}{\mathrm{S_{max}}}\right) \qquad (2)$$

Where S_t (mm) is the available water storage in the root zone per unit area and S_{max} (mm) is the maximum available soil storage capacity and is defined as the difference between the amount of water stored in the top soil layer at wilting point and the maximum moisture content, equal to either the field capacity for the hill slope soils or saturation (e.g., soil porosity) in runoff contributing areas. S_{max} varies according to soil characteristics (e.g., porosity, bulk density) and soil layer depth. Based Eq. 2 the surface soil layer moisture storage can be written as:

$$S_t = S_{t-\Delta t}\left[\exp\left(\frac{(P-PET)\Delta t}{S_{\max}}\right)\right] \qquad when\, P < PET \qquad (3)$$

In this simplified model, direct runoff occurs only from the runoff contributing area when the soil moisture balance indicates that the soil is saturated. Recharge and interflow originate from the remaining hill slopes. It is assumed that the surface runoff from these areas is minimal. This will underestimate the runoff during major rainfall events but since our interest in weekly to monthly intervals was not considered a major limitation.

In the overland flow contributing areas when rainfall exceeds evapotranspiration and fully saturates the soil, any moisture above saturation becomes runoff, and the runoff, R, can be determined by adding the change in soil moisture from the previous time step to the difference between precipitation and actual evapotranspiration, e.g.,

$$R = S_{t-\Delta t} + (P - AET)\Delta t \qquad (4a)$$
$$S_t = S_{\max} \qquad (4b)$$

For high infiltration areas on hillslopes the water flows either as interflow or baseflow to the stream. Rainfall in excess of field capacity becomes recharge and is routed to two reservoirs that produce

baseflow or interflow. We assumed that the baseflow reservoir is filled first and when full, the interflow reservoir starts filling. The baseflow reservoir acts as a linear reservoir and its outflow, *BF,* and storage, $BS_{t,}$ are calculated when the storage is less than the maximum storage, BS_{max} as:

$$BS_t = BS_{t\text{-}\Delta\text{-}} + \left(Perc - BF_{t-\Delta t}\right)\Delta t \qquad (5a)$$

$$BF_t = \frac{BS_t\left[1-\exp(-\alpha\Delta t)\right]}{\Delta t} \qquad (5b)$$

where α is the half life of the aquifer or the time it takes for half of the volume of the aquifer to flow out without the aquifer being recharged.

When the maximum storage, BS_{max}, is reached then:

$$BS_t = BS_{max} \qquad (6a)$$

$$BF_t = \frac{BS_{\max}\left[1-\exp(-\alpha\Delta t)\right]}{\Delta t} \qquad (6b)$$

Interflow originates from the hillslopes and with the slope of the landscape as the major driving force of the water. Under these circumstances, the flow decreases linearly (i.e., a zero order reservoir) after a recharge event. The total interflow, IF_t at time *t* can be obtained by superimposing the fluxes for the individual events (details are given in the Appendix):

$$IF_t = \sum_{\tau=0,1,2}^{\tau^*} 2Perc^*_{t-\tau}\left(\frac{1}{\tau^*} - \frac{\tau}{\tau^{*2}}\right), \qquad \tau \le \tau^* \qquad (7)$$

where τ^* is the duration of the period after the rainstorm until the interflow ceases, IF_t is the interflow at a time *t,* $Perc^*_{t-\tau}$ is the percolation on t-τ days.

Predicting sediment concentration

The Blue Nile runs through a deep gorge partly over bedrock before it reaches the Sudan border. This means that the sediment concentration depends on the amount of suspended sediment delivered by contributing reaches to the main stem of the Nile. Assuming that subsurface flow does not cause erosion then all sediment is contributed by the direct surface runoff (Mul et al., 2008). Therefore, it is reasonable to assume that the sediment concentration in the Nile is determined by direct runoff from the contributing areas. Initially, at the beginning of the rainy season, the contributing areas expand and once the watershed is sufficiently saturated the contributing areas do not expand further and the hillslopes begin contributing interflow. Thus, once the watershed is saturated (i.e., the hillslopes are contributing water to the stream); the sediment concentration in the water is a function of the surface runoff and interflow components. In other words, the subsurface flow dilutes the concentration of sediment delivered by the direct runoff delivered to the stream. The sediment concentration in the river, C^*, occurs just before the hillslopes begin contributing interflow. The discharge is R^* at that time.

Based on the conceptual model above for the period that the hillslopes are contributing interflow the sediment concentration, C, in the river water is the ratio of the direct runoff and total runoff multiplied by C^*, viz:

$$C = C^* \frac{R}{R + IF + BS} \qquad (8)$$

where R, runoff, IF, interflow, and BS, baseflow are predicted by the water balance model, above.

For the period when the subsurface flow is negligible at the onset of the rainy season, the soil erodibility is the greatest because the soil is dry and loose. At the same time from the beginning of the rainfall season the contributing area increases and initially the discharge is less for any given amount of rainfall than it would be later in the season. Although we do not know the exact mechanisms, it is reasonable to assume that the concentration is equal to the ratio of predicted runoff to direct runoff, $R*$ viz:

$$C = C*\frac{R}{R*} \tag{9}$$

Thus, the concentration $C*$ and R* are calibration parameters, and are set equal to the ten day averaged sediment concentration and the discharge during the period just before interflow starts as simulated by the model.

APPLICATION: THE ABAY (BLUE NILE)

The Blue Nile Basin at the border with Sudan covers an area of approximately 180,000 km^2. The river and its tributaries drain a large proportion of the central, western and southwestern highlands of Ethiopia. The basin is characterized by a highly rugged topography and considerable variation of altitude ranging from about 500 m at Sudan border to over 4,250 m above mean sea level (msl) in the Ethiopian highlands. Together with the Dinder and Rahad that join the Blue Nile in Sudan, Ethiopia provides 62% of the flow reaching Aswan (World Bank 2006).

Rainfall varies significantly with altitude and is considerably greater in the Ethiopian highlands than on the Plains of Sudan. Rainfall ranges from less than 1,000 mm/yr near the border of Sudan to between 1,400 and 1,800 mm over parts of the upper basin, in particular, some areas south of Lake Tana. Rainfall exceeds 2,000 mm in parts of the Didessa and Beles catchments.

Both the temporal and spatial distribution of rainfall is governed, , by the movement of air masses associated with the Inter-Tropical Convergence Zone (ITCZ). During the winter dry season (known in Ethiopia as Bega) the region is affected by a dry northeast continental air-mass. From March to May (Belg) the ITCZ brings rain particularly to the southern and southwestern parts of the Basin. In May, there is a short intermission before the main wet season (known locally as Kremt). Around June, the southwest airstream extends over the entire Ethiopian highlands and produces the main rainy season. The summer months account for a large proportion of mean annual rainfall, roughly 70% occurs between June and September and this proportion generally increases with latitude, ranging from 60 to 80%

Available discharge and sediment data: There is relatively little sediment concentration data available for the Blue Nile. One data set of continuous sediment concentrations is given by Ahmed (2003) and consists of ten day averaged sediment concentrations at the gauge station upstream of Rosaries Dam north of the Ethiopia-Sudan border for the period June-October 1993. The 10-day discharge values at this station and the averaged precipitation over the entire Blue Nile basin in Ethiopia are also available for the period of May 1st 1993 to April 30th 1994. In addition, discharge and sediment concentration were obtained for July, August, September, and October for 2003 and 2004. A long record of rainfall in Ethiopia was available from 1995 to 2006 for 15 stations in the Nile basin. We use the 1993 data for calibration and the 2003 and 2004 data for validation.

Calibration: To use the water balance in 1993-1994 water year for calibration we need to start the simulations before the rainfall period begins (and the sediment data were available), thus, we choose to start in January 1993 (Figure 2a).

Parameters needed to simulate discharge include *PET*, which varies little between years and it was set at 5 mm d^{-1} during the dry season and 3.3 mm d^{-1} during the rainy season. The maximum storages, S_{max}, for the contributing area and hillslopes were based initially on the values from Collick et al. (2008) for three SRCP watersheds. Note that for the relatively flat contributing areas and bedrock areas this maximum storage term represents the amount of water that is required to fill up a dry soils before it is saturated and overland flow will occur. For the hillslopes, S_{max} is the moisture required to bring a dry soil up to field capacity after which any extra water will percolate downward. Although the Collick et al. (2008) values gave a reasonable fit, we decided to vary them slightly to improve the agreement between observed and predicted values as the correct distribution between subsurface flow and overland flow directly determines the predicted sediment concentrations. Collick et al. (2008) assumed that 40% of the landscape had a S_{max} value of 100 mm. This represents the contributing area in their model. For the Blue Nile basin, we found a slightly better fit by reducing the contributing area to 30%. We divided the contributing area in two parts (Table 1a): 20% of the area (consisting of the exposed hardpan or bed rock areas) needed little rain to generate direct runoff (i.e., S_{max} = 10mm) and 10% (the saturated bottom lands) needed 250 mm of effective precipitation after the dry season before generating runoff (i.e., S_{max} = 250mm). Note that the weighted average S_{max} for the runoff contributing area in the Blue Nile Basin in Ethiopia compares well with the S_{max} value of 100 mm storage for two of the three SRCP watersheds (Collick et al., 2008).

Scale is important when simulating the hydrological dynamics of the hillslopes in the Blue Nile as compared to the SRCP watersheds located in the upper reaches of the basin (Collick et al., 2008; Hurni et al., 2004). We used a S_{max} value of 500 mm for the hillslopes where the water infiltrates (Table 1a). In two of three SRCP watersheds, approximately 20% of the moisture was lost to deep percolation. To simulate deep percolation, Collick et al. (2008) assumed that the S_{max} was essentially infinite (4000 mm). If we ignore this reservoir, because the deep percolation over the whole Blue Nile in Ethiopia is negligible, we find that the S_{max} = 500 mm for the high infiltration area of the Nile basin (Table 1a) compares well with the values used in Collick et al. (2008).

Scale impacts the interflow and baseflow predictions in the conveyance zone more than the storage values in the uppermost soil layer. A more complicated approach was needed to represent adequately the complex landscape by using both a linear ground water reservoir and a zero order hillslope reservoir. Fitted parameters are given in Table 1b. The τ^* value of 140 days indicates that the hillslopes contribute interflow up to 140 days after the storm occurs. To model sediment concentration (Eqs. 8 and 9), the only calibration parameters is the observed concentration, C*, before interflow occurs and the flux, R*, at that time. We have set this concentration at 5000 mg/l (Table 1b). The discharge at that time is equivalent to 1.4 mm/day over the whole basin. The remaining parameter values are all obtained from the water balance model presented in Figures 2a and 3. Observed and predicted sediment concentrations are shown in Figure 4a. The simulated discharge and sediment concentration fit the observed values well as indicated by the regression coefficient R^2 and Nash-Sutcliffe model efficiency coefficient close to one in Table 2.

Simulation results

The calibrated parameters in Tables 1a and 1b were used to predict the discharge and sediment concentrations of the Blue Nile for the years 2003 and 2004 at the station upstream of Rosaries dam. The predicted and observed stream discharges are depicted in Figures 2b and 2c. The observed sediment concentrations are compared with the predicted concentrations in Figures 4b and 4c. The fit between observed and predicted discharge and sediment concentrations are shown in Table 2. Despite both the different rainfall pattern (total annual precipitation in 1993, 1425 mm; in 2003, 1215 mm; in 2004, 1275 mm) and the simplicity of the model, the discharge and sediment concentrations are reasonably well simulated. In all cases the *F-test* was significant indicating that the observed and predicted values were not significantly different (Table 2). In 2003 the R^2 of the discharge regression was somewhat low although the visual inspection of the regression in Figure 2a shows the fit to be acceptable. The low R^2 was caused by missing the trend in discharge predictions at the end of June and beginning of July. The range in data for 2003 was smaller than for 1993 since we had only the data for the rainy season and not for the dry season. Although the goodness of fit in Table 2 was reasonable for 2004, the discharge was under predicted in August and the first half of September. One of the problems in accurate modeling of the discharge is that the precipitation measurements do not exist in the large area of the northwestern Blue Nile. Figures 1 and 3 show that the flow at the end of the dry season is extremely small indicating that there is little carryover of water from one year to the next.

It is surprising that the sediment concentrations are predicted with reasonable accuracy (Table 2) and particularly that the sediment concentration from the calibration period (1993) performs well despite the significantly higher sediment concentration observed in 2003 and 2004.. It is reassuring that the model captures the high sediment concentration on the rising limb and the lower concentration on the falling limb, verifying that the majority of the flow in the river is base flow that is sediment free (Mul et al., 2008). It is interesting that this simple sediment model can predict the sediment concentrations well using fluxes predicted by the water balance model. We cannot predict the sediment concentration at the end of July when the concentration suddenly drops.

DISCUSSION AND CONCLUSIONS

The hydrological model presented here is based on generating direct runoff on saturated areas, and is reasonably robust. In the beginning of the rainy season almost all flow in the river is direct runoff generated from the 20% of the area that has the smallest storage and likely originates from the bare rock or low storage areas where there is little infiltration or storage of rainfall. As the rainy season progresses (cumulative rainfall increases), the rest of the landscape wets up and direct runoff is generated from the remaining 10% of the contributing area followed by base and interflow from the hill slopes around early July (Figure 3). Note that this corresponds to the time that the sediment concentration in the river is decreasing from the maximum (Figure 4).

The annual volumes of discharge can be found as areas under the discharge curve in Figure 2. The volumes of predicted and observed discharge in 1993-1994 water year (Figure 2a) and 2003 (Figure 2b) are equal. Since there is no carryover storage of flows from year to year the predicted discharge is equal to the annual precipitation minus the annual evapotranspiration. Thus, the water balance of the Blue Nile balances within a hydrologic year. In 2004 (Figure 2c) although we obtained only a partial record (just like in 2003), it seems that we could not close the water balance as well but that might be partially caused by the uncertainty in precipitation.

Figure 2 shows that the discharge is only 20-30% of the precipitation in June, July, and August, during the period when the majority of rainfall occurs. Our water balance approach is able to explain this observed runoff coefficient (i.e., discharge/precipitation) by distributing the effective rainfall (rainfall minus evapotranspiration) over the contributing and saturated areas that generate direct runoff. For our simulations the area that contributes runoff is 30% of the total basin area at the time that the soil is saturated. Thirty percent of the area does not contribute 30% of the rainfall as steamflow since runoff has to be adjested for the evaporation that occurs during the period. A portion of the discharge in August (as shown in Figure 3) when the runoff coefficient is the greatest, originates from interflow and base flow generated from the 70 % of the basin area where the rainfall infiltrates.

It is also interesting to note how minimal the scale effects are for the basin. Using similar parameters to Collick et al. (2008) (predicting discharge from small watersheds <500 ha) we were able to model the flow from the Blue Nile basin in Ethiopia with equal efficiency. The Basin as a whole and the small watersheds have nearly the same portions of contributing areas and hillslope areas with similar amounts of water needed before overland flow or interflow starts after the onset of the rains. However, the interflow component did show some scale effects. The interflow period last longer for the Blue Nile watershed than for small watersheds. This should have been expected since in the small catchments up to 20% of interflow was not recorded at the gage, and likely ended up as a regional flow , which would be measured at larger gauges. We are unable to study at scale effects of the sediment since the sediment concentrations were not analyzed in the small watersheds.

Despite the reasonable fit of the predicted and observed concentrations, processes governing the erosion and sedimentation dynamics are not fully understood in the Blue Nile, thus the sediment predictions in this paper should be considered tentative until more testing is done. It is interesting to note the decrease in observed stream sediment concentrations before the peak discharge occurs, and that the model captures the phenomenon is important, but other, more complicated process may play a role. For instance, it could be the result of relating the sediment concentration to the time when the watershed becomes covered by vegetation or when the watershed is fully wetted up and erodibility of all soils are decreasing. Based on watershed outflow concentrations, we cannot discriminate between these mechanisms since both signals appear at the same time because when interflow occurs the watershed is wet and vegetation begins to develop. However, the interflow explanation seems to be reasonable since during the rainy season high sediment concentrations are observed in the basin and relatively sediment free water is observed after the surface runoff has ended. Currently there are very few sediment models available and although the model described here can capture the trends, more research is needed to elucidate erosion processes, particularly gully erosion within the watershed (Daba et al., 2003; Billi and Dramis, 2003).

Acknowledgement

This research was funded through the CGIAR Challenge Program on Water and Food (CPWF) and carried under the auspices of the International Water Management Institute (IWMI) managed project, PN19 ‘Upstream-Downstream in the Blue Nile

APPENDIX A

DERIVATION OF INTERFLOW DISCHARGE FOR ZERO ORDER RESERVOIRS

The flux from a reservoir in general can be expressed as a function of the flux from the aquifer (Brutsaert and Nieber, 1977)

$$\frac{dQ}{dt} = -aQ^{b} \qquad (A1)$$

where *a* is a constant. Hillslopes can be modeled as zero order reservoir *b=0* (Steenhuis et al., 1999; Stagnitti et al., 2004) and regular groundwater outflow as a first order reservoir *b=1*

The flux equation is derived for a zero order reservoir as a function of the reservoir storage *S* the flux from the reservoir decreases linearly for a single storm, i.e.:

$$\frac{dQ_t}{dt} = -a_0 \qquad (A2)$$

Without loss of generality we can replace the time *t* with τ in Eq. A1 defined as the time after the storm has occurred. In addition, we have designated the flow Q_t is from the particular storm occurring at time *t*. Integrating with respect to *t* subject to the boundary condition that at time τ^* after the rain event the flux is zero (i.e., $Q=0$ at $\tau=\tau^*$). Integrating Eq. A2 with the boundary condition specified and τ as the time variable:

$$Q_t = a_0(\tau^* - \tau) \qquad (A3)$$

Integrating again from $\tau=0$ to $\tau = \tau^*$ we find the storage in the aquifer:

$$\int_0^{\tau^*} Q_t d\tau = SI_t = \frac{1}{2} a_o \tau^{*2} \qquad (A4)$$

Where $Perc^*_t$ is the amount of water added to the reservoir at time *t*. In order to conserve mass it is obvious from Eq. A3 that:

$$a_o = \frac{2Perc_t^*}{\tau^{*2}} \qquad (A5)$$

Combining Eqs. A5 and A3 results in the zero order flow equation for the discharge of the aquifer for a storm occurring at time *t*:

$$Q_t = 2Perc_t^*\left(\frac{1}{\tau^*} - \frac{\tau}{\tau^{*2}}\right) \qquad (A6)$$

The total flux is equal for a daily time step

$$BI_t = \sum_{\tau=0}^{\tau^*} 2Perc_{t-\tau}^*\left(\frac{1}{\tau^*} - \frac{\tau}{\tau^{*2}}\right) \qquad (A7)$$

REFERENCES

Ahmed A.A., 2003. Sediment Transport and Watershed Management Blue Nile System. Friend/Nile Project report. Sudan.

Awulachew, S. B., McCartney M., Steenhuis T. S.and Ahmed A. A. 2009. A Review of Hydrology, Sediment and Water Resource Use in the Blue Nile Basin. International Water Management Institute Working Paper: 131. In press

Ayenew T, Gebreegziabher Y. 2006. Application of a spreadsheet hydrological model for computing the long-term water balance of Lake Awassa, Ethiopia. *Hydrological Sciences Journal-Journal des Sciences Hydrologiques* 51 (3): 418-431

Billi P. and Dramis F. 2003. Geomorphological investigation on gully erosion in the Rift Valley and the northern highlands of Ethiopia. *Catena* 50: 353-368

Brutsaert, W., and Nieber J.L. 1977. Regionalized drought flow hydrographs from a mature glaciated plateau. *Water Resources. Research* ***13***: 637-643.

Collick A.S., Easton Z.M., Adgo E., Awulachew S.B. Zeleke G., and Steenhuis T S. 2008. Application of a physically-based water balance model on four watersheds throughout the upper Nile basin in Ethiopia. *In Eds.* W. Abtew and A. M. Melesse. *Proceedings of the 2008 workshop on the Nile Basin hydrology and ecology under extreme climatic conditions*. [submitted to special issue]

Conway D. 1997. A water balance model of the Upper Blue Nile in Ethiopia. *Hydrological Sciences Journal* **42**: 265–286.

Daba S., Rieger W; Strauss P. 2003. Assessment of gully erosion in eastern Ethiopia using photogrammetric techniques, *Catena,* 50: 273-291

Desta, G. 2003. Estimation of Runoff Coefficient at different growth stages of crops in the highlands of Amhara Region. MSc Thesis, Alemaya University, Ethiopia.

Feoli E., Vuerich L.G., and Zerihun W. 2002 Evaluation of environmental degradation in northern Ethiopia using GIS to integrate vegetation, geomorphological, erosion and socio-economic factors. *Agriculture Ecosystems & Environment* 91: 313-325

Haile M, Herweg K.and Stillhardt B., 2006: Sustainable Land Management: *A New Approach to Soil and Water Conservation in Ethiopia by Land resources Management and Environmental Protection Department.* Mekele Ethiopia.

Haregeweyn N and Yohannes F. 2003: Testing and evaluation of the agricultural non-point source pollution model (AGNPS) on Augucho catchment, western Hararghe, Ethiopia. *Agriculture Ecosystems & Environment* 99: 201-212 2003

Hengsdijk H., Meijerink G., Mosugu M., 2005. Modelling the effect of three soil and water conservation practices in Tigray, Ethiopia. *Agric. Ecosyst. Environ*. 105: 29-40.

Hurni, H., Tato K., and Zeleke G. 2005. The Implications of Changes in Population, Land Use, and Land Management for Surface Runoff in the Upper Nile Basin Area of Ethiopia Hans *Mountain Research and Development 25:*147–154

Johnson P.A., Curtis P.D. 1994. Water-balance of Blue Nile river basin in Ethiopia. *Journal of Irrigation and Drainage Engineering-ASCE* **120**: 573-590

Kebede S, Travi Y, Alemayehu T, Marc V. 2006. Water balance of Lake Tana and its sensitivity to fluctuations in rainfall, Blue Nile basin, Ethiopia. *Journal of Hydrology* **316**: 233–247.

Kim, U. and Kaluarachchi, J.J. 2008 Application of parameter estimation and regionalization methodologies to ungauged basins of the Upper Blue Nile River Basin, Ethiopia. J. Hydrology 362: 39-56

Legesse D., Vallet-Coulomb, C. and. Gasse F. 2003. Hydrological response of a catchment to climate and land use changes in Tropical Africa: case study South Central Ethiopia. *Journal of Hydrology*, 275: 67-85

Liu, B.M., Collick A.S., Zeleke G., Adgo E., Easton Z.M., and Steenhuis T.S. 2008. Rainfall-discharge relationships for a monsoonal climate in the Ethiopian highlands. *Hydrol. Proc*. 22: 1059-1067.

McHugh O.V. 2006. Integrated water resources assessment and management in a drought-prone watershed in the Ethiopian highlands. PhD dissertation, Department of Biological and Environmental Engineering. Cornell University Ithaca NY.

Mishra, A., Hata, T., Abdelhadi, A.W., Tada, A., and Tanakamaru, H. 2003. Recession flow analysis of the Blue Nile River. *Hydrological Processes,* 14: 2825-2835.

Mohammed., A., Yohannes, F., Zeleke, G, 2004.. Validation of agricultural non-point source (AGNPS) pollution model in Kori watershed, South Wollo, Ethiopia. *International Journal of Applied Earth Observation and Geoinformation* 6: 97–109

Mul, M. L., R. K. Mutiibwa a, S. Uhlenbrook and H. H.G. Savenije. 2008. Hydrograph separation using hydrochemical tracers in the Makanya catchment, Tanzania. *Physics and Chemistry of the Earth* **33**: 151–156.

Nyssen J., Poesen J., Gebremichael D., Vancampenhout K., D'Aes M., Yihdego G., Govers G., Leirs H., Moeyersons J., Naudts J., Haregeweyn N., Haile M., Deckers J. 2007. Interdisciplinary on-site evaluation of stone bunds to control soil erosion on cropland in Northern Ethiopia. *Soil & Tillage Research* 94: 151-163

Nyssen J., Haregeweyn N., Descheemaeker K., Gebremichael D., Vancampenhout K., Poesen J.,; Haile, M., Moeyersons J., Buytaert W., Naudts J., Deckers, J., Govers G. 2006. Modelling the effect of soil and water conservation practices in Tigray, Ethiopia (vol 105, pg 29, 2005*). Agriculture Ecosystems & Environment 114*: 407-411 JUN 2006

Okumu B.N., Russell N., Jabbar M.A., Colman D., Saleem M.A.M., Pender J. 2004. Economic impacts of technology, population growth and soil erosion at watershed level: The case of the Ginchi in Ethiopia *Journal of Agricultural Economics* 55: 503-523

Setegn, S G., Ragahavan S., Bijan D. 2008. Hydrological Modelling in the LakeTana Basin, Ethiopia using SWAT model. The Open Hydrology Journal 2: 49-62

Shiferaw B., and Holden S.T. 2000. Policy instruments for sustainable land management: the case of highland smallholders in Ethiopia. *Agricultural Economics* 22: 217-232

Sonneveld B.G.J.S. 2003. Formalizing expert judgments in land degradation assessment: A case study for Ethiopia. *Land Degradation & Development* 14: 347-361

Steenhuis T.S. and van der Molen W.H. 1986. The Thornthwaite-Mather Procedure as a Simple Engineering Method to Predict Recharge. *J. Hydrol*. **84**: 221-229.

Steenhuis T.S., Parlange J.-Y., Sanford W.E., Heilig A., Stagnitti F., Walter M.F. 1999. Can we distinguish Richards' and Boussinesq's equations for hillslopes?: The Coweeta experiment revisited. *Water Resources Research* **35**: 589-593.

Tamene L. and Vlek P. L. G. 2007. Assessing the potential of changing land use for reducing soil erosion and sediment yield of catchments: a case study in the highlands of northern Ethiopia. *Soil Use and Management* 2: 82-91

Thornthwaite C. W. and Mather J. R. (1955), The Water Balance*,* Publications in Climatology VIII(1): 1-104, Drexel Institute of Climatology, Centerton, NJ

Weiß M. and Schaldach. R. 2008. Current and future green and blue water fluxes in Africa under climate and land use change. Geophysical Research Abstracts, Vol. 10, EGU2008-A-06008, 2008 SRef-ID: 1607-962/gra/EGU2008-A-06008. EGU General Assembly 2008

Zeleke, G. 2000. Landscape Dynamics and Soil Erosion Process Modeling in the North-western Ethiopian Highlands. African Studies Series A 16, Geographica Bernensia, Berne.

Zeleke G. and Hurni H. 2001. Implications of Land Use and Land Cover Dynamics for Mountain Resource Degradation in the Northwestern Ethiopian Highlands. Mountain Research and Development 21:184–191

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Figure 4: Predicted and observed sediment concentration in the Abay (Blue Nile) at the Ethiopian Sudan border upstream of the Rosaries Dam. a) 1993 calibrated; b) 2003 validation; c) 2004 validation.

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Table 2: Statistical analysis of simulated and observed 10 day averaged discharge and sediment concentrations in the Blue Nile at the Ethiopia-Sudan border. Both the linear regression R-square and the Nash-Sutcliffe efficiency are calculated. The slope and the intercept of the linear regression are shown.

Table 1a: Model input values for surface flow components: The watershed is divided up in areas with different characteristics: exposed bedrock and saturated areas that contribute surface runoff or hillsides that produce recharge when the soil is above field capacity. Maximum storage of water is the amount of water needed from wilting point to become either saturated or to reach field capacity.

Description of area	*Maximum of Water storage, mm*	*Portion occupying in watershed*
contributing area rock outcrops	10	0.2
contributing area saturated bottoms	250	0.1
recharge area Hillside	500	0.7

Table 1b: Model input values for the baseflow and interflow parameters. SB_{max} is the maximum storage of the linear base flow reservoir; α is the time it takes in days to reduce the volume of the baseflow reservoir by a factor 2 under no recharge conditions, t* is the duration of the period after a single rainstorm until interflow ceases, C* and R* are the calibrated sediment concentration and discharge rate respectively just before interflow becomes significant.

Parameter description	*Value*	*Units*
Maximum storage capacity of linear reservoir, SB_{max}	20	mm
Half life, α	*35*	days
Interflow duration after rainfall. t*	140	days
Calibrated sediment concentration , C*	500	mg/l
Calibrated amount of runoff R*	1.4	mm/day

Table 2: Statistical analysis of simulated and observed 10 day averaged discharge and sediment concentrations in the Blue Nile at the Ethiopian Sudanese border. Both the linear regression R-square and the Nash-Sutcliffe efficiency are calculated. The slope and the intercept of the linear regression are shown.

	1993-1994		**2003**		**2004**	
	Flow	**Sediment**	**Flow**	**Sediment**	**Flow**	**Sediment**
Slope	0.99	0.84	0.87	0.88	1.19	0.95
Intercept	0.58	-19.72	1.20	-183.62	0.67	776.04
p-value (F-test)*	<0.001	<0.001	0.017	0.001	<0.001	0.002
$R^{2\dagger}$	0.98	0.81	0.45	0.74	0.80	0.74
NSE‡	0.98	0.75	0.42	0.60	0.73	0.69

*Test of significance for regression of observed and predicted response.

†Simple R-square of regression

‡ Nash –Sutcliffe efficiency defined as: $1-\left(\frac{\sum(Observed-Predicted)^2}{\sum(Observed-Observed\ Mean)^2}\right)$

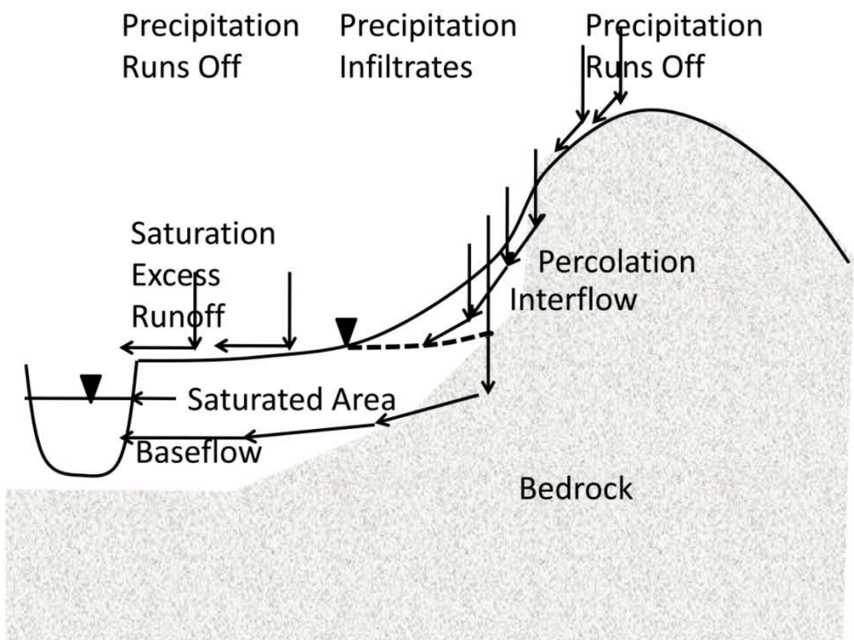


Figure 1: Simplified Hillslope

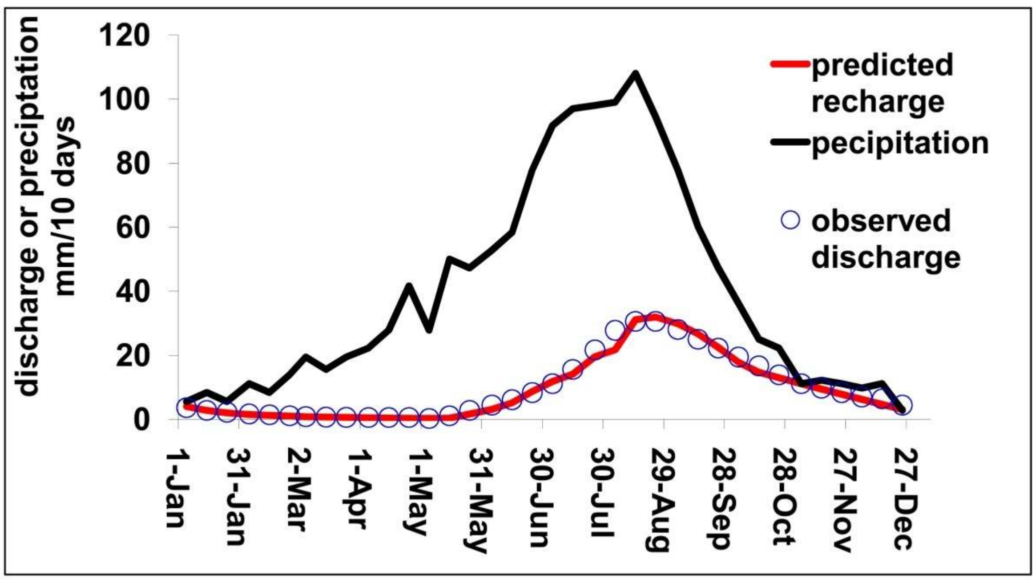

discharge or preciptation
mm/10 days
120
100
80
60
40
20
0
predicted
recharge
pecipitation
observed
discharge
1-Jan
31-Jan
2-Mar
1-Apr
1-May
31-May
30-Jun
30-Jul
29-Aug
28-Sep
28-Oct
27-Nov
27-Dec

Figure 2a

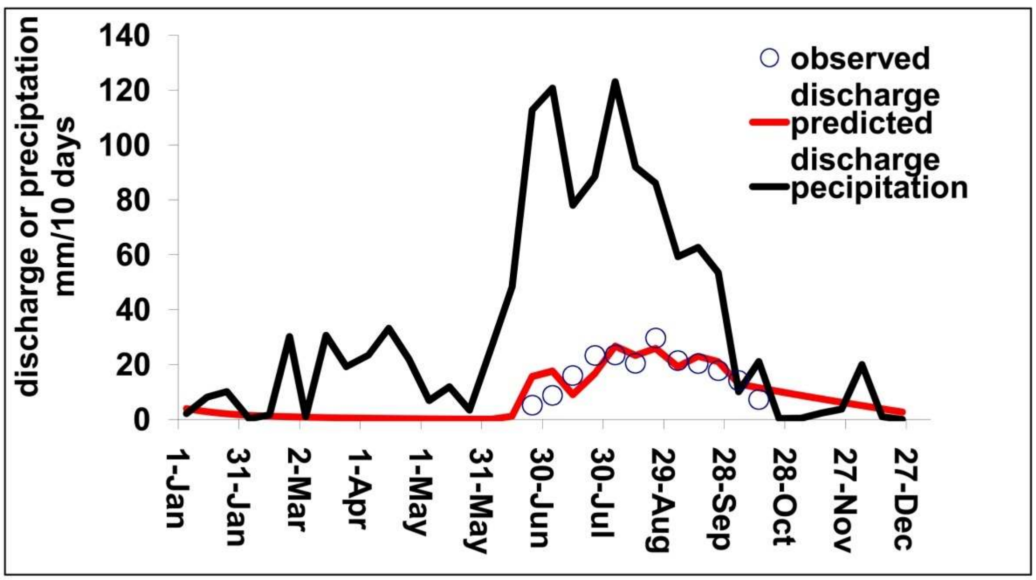
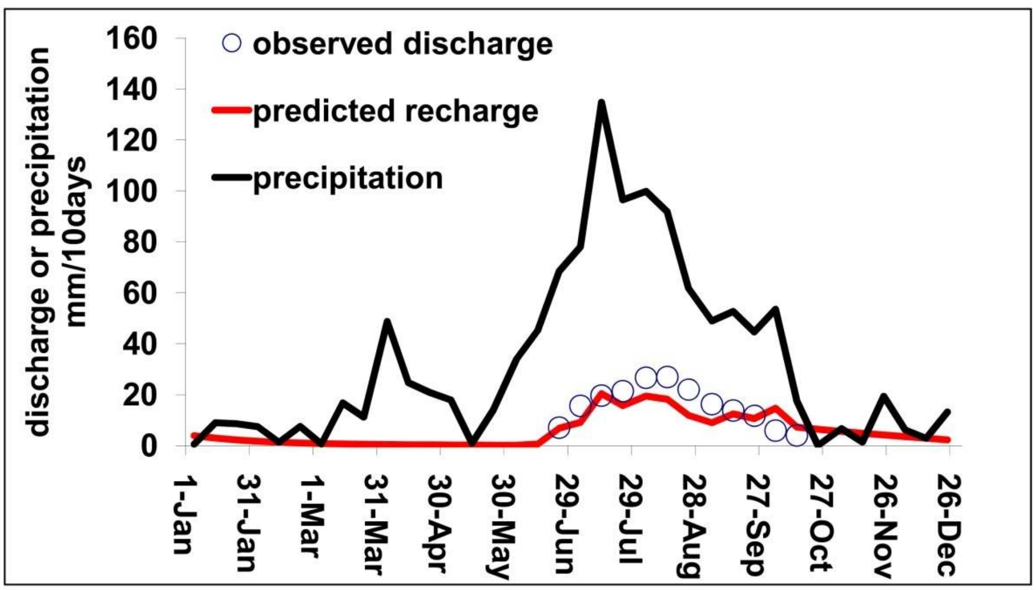

discharge or preciptation mm/10 days
140
120
100
80
60
40
20
0
observed discharge
predicted discharge
pecipitation
1-Jan
31-Jan
2-Mar
1-Apr
1-May
31-May
30-Jun
30-Jul
29-Aug
28-Sep
28-Oct
27-Nov
27-Dec

Figure 2b



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predicted recharge
precipitation
discharge or precipitation mm/10days
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140
120
100
80
60
40
20
0
1-Jan
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1-Mar
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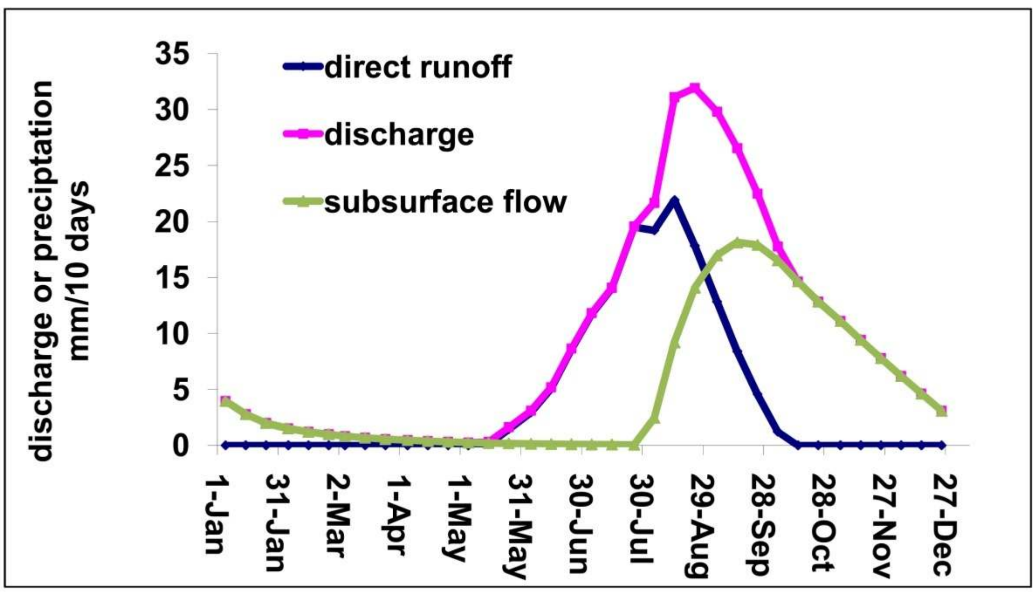
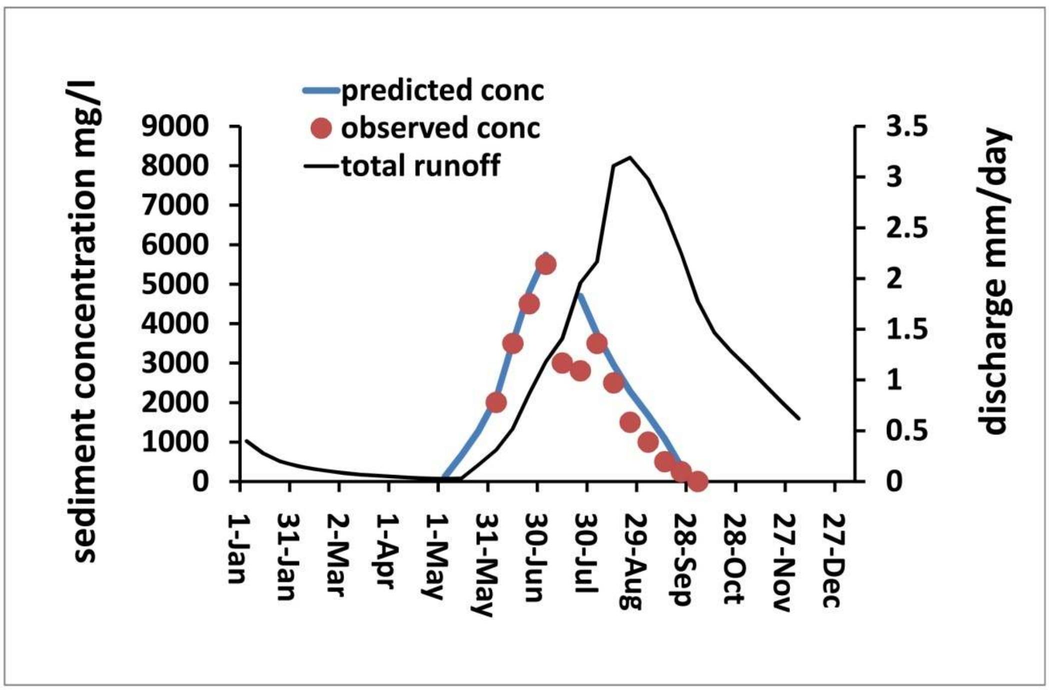
26-Nov
26-Dec

Figure 2c



direct runoff
discharge
subsurface flow
discharge or preciptation mm/10 days
35
30
25
20
15
10
5
0
1-Jan
31-Jan
2-Mar
1-Apr
1-May
31-May
30-Jun
30-Jul
29-Aug
28-Sep
28-Oct
27-Nov
27-Dec

Figure 3


predicted conc
observed conc
total runoff
sediment concentration mg/l
discharge mm/day
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8000
7000
6000
5000
4000
3000
2000
1000
0
3.5
3
2.5
2
1.5
1
0.5
0
1-Jan
31-Jan
2-Mar
1-Apr
1-May
31-May
30-Jun
30-Jul
29-Aug
28-Sep
28-Oct
27-Nov
27-Dec

Figure 4a

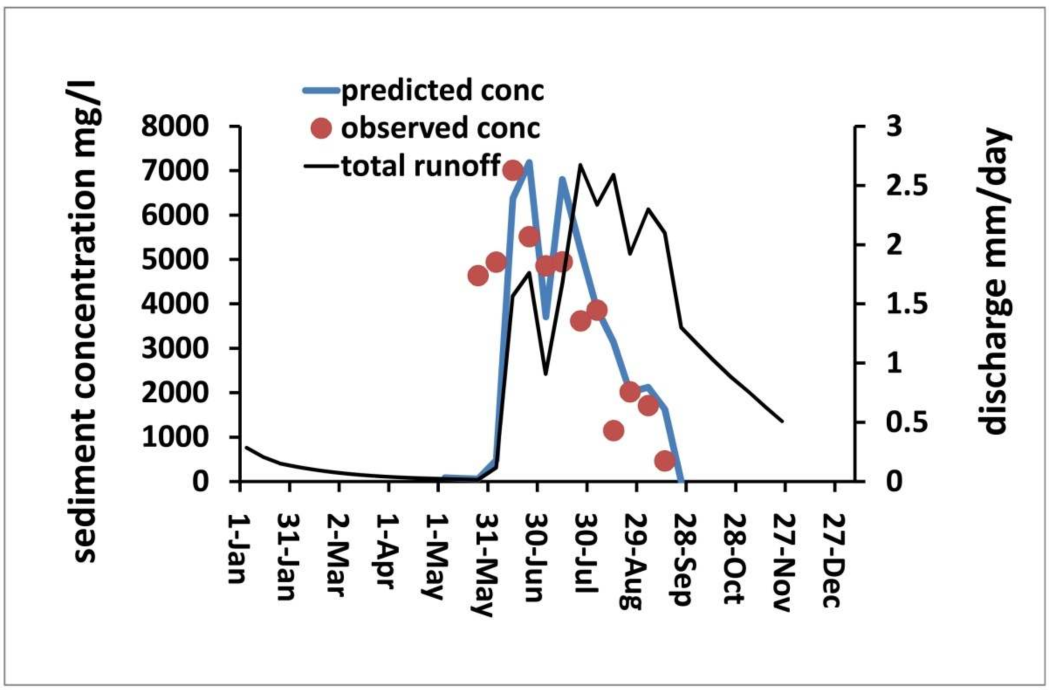
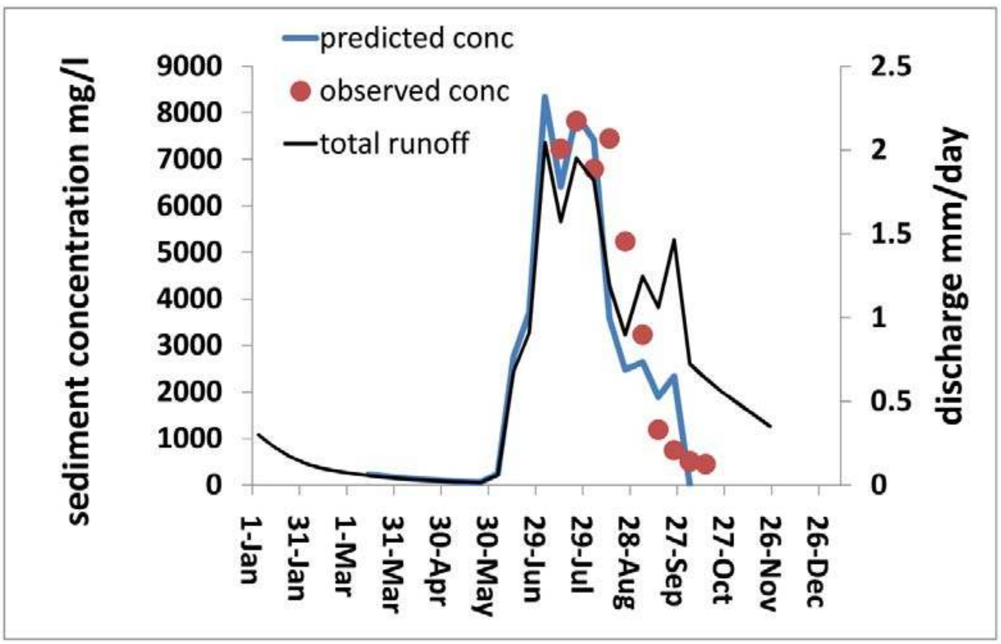

predicted conc
observed conc
total runoff
sediment concentration mg/l
discharge mm/day
8000
7000
6000
5000
4000
3000
2000
1000
0
3
2.5
2
1.5
1
0.5
0
1-Jan
31-Jan
2-Mar
1-Apr
1-May
31-May
30-Jun
30-Jul
29-Aug
28-Sep
28-Oct
27-Nov
27-Dec

Figure 4b


predicted conc
observed conc
total runoff
sediment concentration mg/l
discharge mm/day
9000
8000
7000
6000
5000
4000
3000
2000
1000
0
2.5
2
1.5
1
0.5
0
1-Jan
31-Jan
1-Mar
31-Mar
30-Apr
30-May
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29-Jul
28-Aug
27-Sep
27-Oct
26-Nov
26-Dec

Figure 4c