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Estimating Transmission Losses

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Abstract

Infiltration or transmission losses in ephemeral stream channels are an important part of the water balance in arid and semiarid regions. The governing equations for movement of flood waves subject to transmission losses are simplified through a time-averaging process to produce an ordinary differential equation describing transmission losses as a function of distance, upstream inflow, lateral inflow, channel dimensions, flow duration statistics, and effective hydraulic conductivity. The resulting equation has an analytic solution, and transmission losses throughout an entire channel network can be estimated by cascading the outflow from one or more upstream channel reaches as inflow to subsequent downstream reaches. The procedures maintain a mass balance, and thus provide time averaged, but spatially varying, estimates of potential groundwater recharge through ephemeral stream beds. Runoff volume-peak discharge relationships can also be used to estimate the impact of transmission losses on flood peaks, and thus improve flood peak estimates in ephemeral streams.

Introduction

In much of the world classified as arid and semiarid, increasing demands for water are resulting in increased competition for available water resources. This increased competition results in the need for improved means of quantifying components of the hydrologic cycle to better predict runoff, aquifer recharge, and downstream surface water yield. Abstractions of streamflow in ephemeral stream channels as infiltration losses to the channel bed and banks are called transmission losses.

These infiltration, or transmission, losses are an important component of the water balance in many arid and semiarid regions, because they influence water yield (e.g., Babcock and Cushing, 1941; Burkham, 1970; Renard, 1970), because they influence the shape of the runoff hydrograph (e.g., Renard and Keppel, 1966; Jordan, 1977; Thornes, 1977), because they influence runoff peak discharge (e.g., Renard and Keppel, 1966; Lane, 1982), and because they support riparian vegetation and recharge local aquifers (e.g., Babcock and Cushing, 1941; Renard, 1970; Wilson et al., 1980). Accurate estimation of transmission losses is thus seen as an important step in assessment of surface water supplies, assessment of groundwater recharge through stream channels, development of runoff hydrographs, and prediction of runoff peak discharge in many arid and semiarid watersheds.

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A broad spectrum of methods and procedures to estimate transmission losses has been developed. A simplified approach, developed to relate transmission loss rates to rates of inflow, resulted in development of inflow-loss rate equations (e.g., Burkham, 1970, SCS, 1972) for stream channel reaches. Volume-based equations include regression equations (Lane et al., 1971) relating inflow, outflow, and transmission loss volumes, and equations relating volume of alluvium in a channel reach to potential volume of transmission losses (Wallace and Lane, 1978). Storage routing equations, where channel reaches are treated as a cascade of leaky reservoirs, have been developed (e.g., Peebles, 1975) to approximate transmission losses in storage-routing procedures. The governing equations for movement of flood waves subject to transmission losses have been simplified through time-averaging procedures to develop ordinary differential equations describing transmission losses as a function of distance, upstream inflow, lateral inflow, channel dimensions, flow duration statistics, and effective hydraulic conductivity (Jordan, 1977; Lane, 1982).

Recently, a simplified transmission loss estimation procedure was developed (Lane, 1983), and incorporated as a chapter in the U.S. Soil Conservation Service National Engineering Handbook (SCS, 1972). This procedure has been incorporated in the channel component of a distributed simulation model (Lane, 1982), and has been used in flood frequency analysis (Lane et al., 1985). The purpose of this paper is to briefly describe the distributed model, incorporating transmission losses, to illustrate its application in flood frequency analysis, and to outline procedures to improve estimates of water yield and recharge potential for ephemeral stream channel systems.

Overview of the Distributed Model

Runoff from upland areas is computed using the SCS Runoff Equation (SCS, 1972) as:

$$V = \begin{cases} 0 & P > 0.2S \\ \dfrac{(P-0.2S)^2}{P + 0.8S} & P > 0.2S \end{cases} \tag{1}$$

wherein V = runoff volume (in. or mm), P = rainfall depth (in. or mm), and S = retention (in. or mm). The equation relating runoff curve number, CN, and retention, S, is given in English units as:

$$CN = \frac{1{,}000}{10 + S} \tag{2}$$

or

$$S = \frac{1{,}000}{CN} - 10 \tag{3}$$

The National Engineering Handbook (SCS, 1972) lists values of CN for four hydrologic soil groups, for various cover-land use complexes, and for three antecedent moisture conditions.

Given runoff volume, peak discharge is estimated using the equation

$$Q = C_5 \ V/D \tag{4}$$

where Q is peak discharge (in./hr or mm/hr), V is runoff volume from Eq. (1) (in. or mm), D is mean duration of flow or hydrograph base time (hr), and C_5 is a coefficient, or parameter, expressing hydrograph shape. Notice that Eq. (4) is of the form of the SCS peak discharge equation

$$Q = 484 \ VA/Tp \tag{5}$$

if time to peak, Tp, is assumed proportional to flow duration, D, flow duration is, in turn, proportional to watershed area, A, and the conversion factor, 484, is incorporated in the coefficient.

From analysis of runoff hydrographs from watersheds in Arizona, Murphey et al. (1977) found that mean flow duration was related to watershed area, A, as:

$$\bar{D} = C_1 \ A^{C_2} \tag{6}$$

where $\bar{D}$ is in. hr, and A is in. sq mi. The corresponding mean volume of runoff, $\bar{V}$, (in.) was related to drainage area, A, as:

$$\bar{V} = C_3 \ A^{C_4} \tag{7}$$

where C_1 through C_4 were parameters determined from hydrograph analysis.

In addition to the runoff curve number for upland runoff and the basin-average hydrograph parameters (C_1 - C_5), the distributed model requires the input of transmission loss parameters, as described below. Based upon the work of Jordan (1977), Lane (1982) approximated the rate of change in runoff volume with distance in a reach of infiltrating channel as:

$$\frac{dV}{dx} = - wc - wk \ V(x,w) + V_L/x \tag{8}$$

where V(x,w) is the volume of outflow (acre-ft or m^3) from a channel reach of length x (mi or km) and mean width w (ft or m), V_L is the volume of lateral inflow along the reach in the same units as V(x,w), and c and k are parameters. The solution to Eq. (8) is:

$$V(x,w) = a(x,w) + b(x,w)V_u + F(x,w)V_L/x \tag{9}$$

where $V(x,w) \geq 0$ is the outflow volume in acre-ft or m^3, and V_u is the upstream inflow in the same units. The relationships between a(x,w), b(x,w), F(x,w), and the parameters c and k are:

$$a(x,w) = \frac{a}{1-b}\big(1 - b(x,w)\big) \tag{10}$$

$$b(x,w) = e^{-kxw} \tag{11}$$

and

$$F(x,w) = \big(1 - b(x,w)\big)/kw \tag{12}$$

where, in English units, the parameter values are:

$$a = -0.00465\ KD \tag{13}$$

$$k = -1.09\ \ell n(1-0.00545\ K\bar{D}/\bar{V}) \tag{14}$$

$$b = e^{-k} \tag{15}$$

and K is the effective, steady-state hydraulic conductivity (in./hr or mm/hr). The parameter c, in Eq. (8), is expressed as $c = -ka/(1-b)$, and a, b, and k are defined above.

Eq. (9) is the solution to Eq. (8), and is the runoff volume at the end of a channel reach of length x and mean width w. Each channel reach can receive upstream input from an upland area, or from one or two upstream tributary channels. Lateral inflow to a channel reach is from one or two lateral contributing areas, and is assumed uniform along the channel reach. The channel network is represented by any number of channel reaches, and the watershed is modeled as the channel network and the upland and lateral flow areas contributing to the channel network. Eq. (1) is used to estimate upland runoff and lateral inflow; Eq. (9) is used to estimate runoff volume, and Eq. (4) is used to compute peak discharge.

Applications

The distributed model was applied to data from 8 small watersheds in southeastern Arizona to predict flood peaks, and from them, develop flood frequency analyses. The Walnut Gulch Experimental Watershed is a 58 mi^2 (150 km^2) watershed tributary to the San Pedro River in southeastern Arizona. It ranges in elevation from just over 4000 ft (1220 m) to about 6300 ft (1930 m) at the headwaters, and is predominantly a brush- and grass-covered rangeland watershed, with brush dominating in the lower elevations and grass dominating in the upper elevations. Of the watersheds listed in Table 3, 63.103 and 63.104 are located in the brush-dominated areas, 63.111 is located in the area dominated by grass cover, and watersheds 63.011 and 63.008 have mixed vegetation types. The Safford watershed (45.001) is a 0.81 mi^2 (2.1 km^2) watershed with sparse vegetation consisting of shrubs and some short grasses. Approximately 85% of the area is bare, and the watershed has been described as sparsely vegetated rangeland (Renard, 1970). The two watersheds, located in (High School Wash) and near (Big Wash) Tucson, Arizona, are described by Zeller (1979). High School Wash drains a 0.90 mi^2 (2.33 km^2) urbanized area of Tucson with approximately 29% impervious areas, and Big Wash drains a 2.75 mi^2 (7.12 km^2) desert brush area in the Tucson Mountains west of Tucson. Lengths of record, drainage area, and estimated 2-, 10-, and 100-yr flood peaks from observed data, and from the distributed model, are shown in Table 1. The relationship between data based (from frequency analysis of observed data) and simulated (from the distributed model) flood peaks is shown in Fig. 1. Notice that the distributed model explains about 85% of the variance in flood peaks ($R^2 = 0.85$). Relationships between peak discharge and watershed area are shown in Fig. 2. Notice that the magnitude of the flood peaks per unit area (cfs per mi^2 or cms per km^2) decrease rapidly with increasing drainage area.

Table 1.--Comparison of estimated flood peaks based on observed data and simulation results. (Note: 1 cfs/mi^2 = 0.0109 cms/km^2)

Watershed	Length of record (yr)	Drainage area (mi^2)	Estimated flood peaks (cfs/mi^2)					
			Observed data			Simulated data		
			2 yr	10 yr	100 yr	2 yr	10 yr	100 yr
Walnut Gulch								
63.103	17	.0142	620.	1460.	2960.	610.	1750.	3790.
63.104	17	.0175	710.	2110.	5160.	630.	1750.	3740.
63.111	20	.223	600.	1510.	3190.	370.	1040.	2230.
63.011	13	3.18	210.	850.	2520.	230.	940.	2890.
63.008	13	5.98	120.	390.	1050.	140.	390.	840.
Safford, AZ								
45.001	30	.81	100.	400.	1240.	110.	460.	1220.
Tucson, AZ								
High School Wash	8	.90	420.	930.	1690.	300.	1020.	2150.
Big Wash	11	2.75	80	640.	2480.	270.	780.	1520.

1. Observed flood peaks based on log-normal probability distributions fitted to observed data, except Tucson data (from Zeller, 1979), which were fitted with log-pearson Type III distribution.
2. Simulated flood peaks based on 2-, 10-, and 100-yr point rainfall depths for 60 min at Walnut Gulch (Osborn, 1983), Tucson (Reich, 1978), and Safford (NOAA Atlas-2; Miller et al., 1973).

The solid curves shown in Fig. 2 are least squares lines fitted to the data-based flood peaks shown by the individual data points. The dashed lines in Fig. 2 represent the relationships for the simulated flood peaks (individual points not shown in Fig. 2 to avoid confusion) from Table 1. Notice that the data-based simulated flood peaks induce similar flood peak-drainage area relationships, but that there is a great deal of variability of the data about the trend lines shown in Fig. 2 (i.e., values of R^2 varied from 0.5 to 0.7). Nonetheless, data such as shown in Fig. 2 are of significant value in establishing regional flood frequency relationships, in evaluating existing flood frequency relationships (i.e., Boughton and Renard, 1984), and in evaluating existing and proposed flood estimation procedures (i.e., Zeller, 1979; Lane et al., 1985).

A second important consequence of the relationships shown in Fig. 2 is a deterministic description and explanation of the phenomena responsible for empirical observations of decreasing water yield with increasing drainage area relationships in semiarid regions (i.e., Keppel, 1960; Renard, 1970). The fact that the distributed model mimics the observed peak discharge-drainage area relationships (Fig. 2) suggests that we are improving our ability to understand and predict runoff-drainage area relationships observed in semiarid regions such as southeastern Arizona. For example, the simulation model suggests that transmission losses reduce the 2-yr flood peak from 0.0142 mi^2 watershed 63.103 by about 2%, but reduce the 2-yr flood peak from the 5.98 mi^2

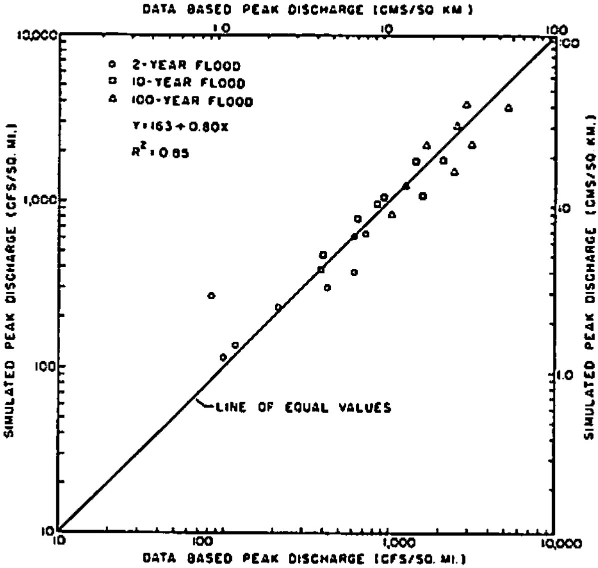

Figure 1. Comparison of data based and simulated flood peaks.

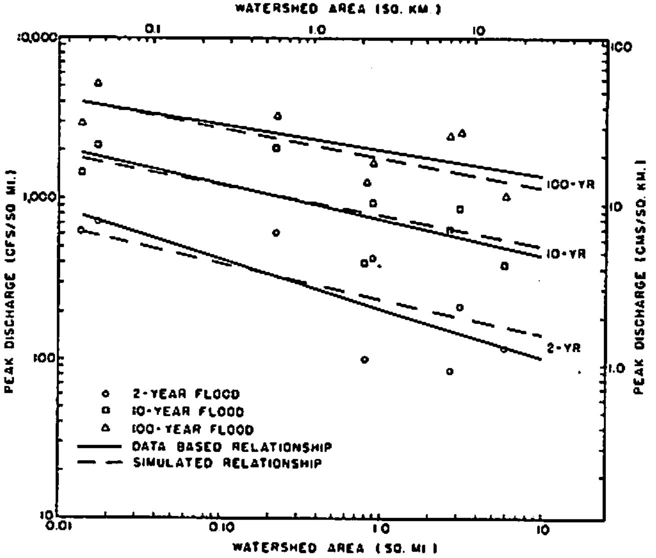

Figure 2. Comparison of data based and simulated flood peaks as a function of drainage area.

watershed 63.008 by about 30%. This improved ability will contribute to our ability to predict flood peaks, storm and annual water yield, and groundwater recharge through ephemeral stream channels.

Limitations

The simplified transmission loss and distributed runoff model described herein has been validated using runoff data from Arizona, Kansas, Nebraska, and Texas, and the flood peak estimation procedure has been validated using flood peak data from small semiarid watersheds in southeastern Arizona. However, as a simplified model, the procedures have obvious shortcomings in their inability to fully describe the dynamic nature of transmission losses and runoff and geographical limits on the parameter values. The transmission loss equations are limited to streamflow (from thunderstorm rainfall) as it occurs in ephemeral stream channels with infiltration losses, and the distributed model has not been evaluated for conditions other than those represented by the small (less than 10 mi^2 or 26 km^2) semiarid watersheds in southeastern Arizona.

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