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Spatial and temporal variability in streamflow generation on the West Fork of Walker Branch Watershed

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BSTRACT

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Spatially intensive measurements of streamflow were used to document the spatial and temporal ıriability in streamflow generation on the West Fork of Walker Branch Watershed, a 38.4 ha forested .tchment in Oak Ridge, Tennessee. The study focused on a 300 m section of a small stream, and covered wide range of flow conditions (Q_{weir}, streamflow at the basin outlet, varied from about 350 to 3500 l in^{-1}). There was enormous spatial variability in the stream inflow, down to the finest scale investigated eaches 20 m in length). Lateral inflow to longer reaches (60–130 m) was linearly correlated with Q_{weir} over .e full range of flows studied, making it possible to estimate the spatial pattern of stream inflow from easurement of Q_{weir} alone. The heterogeneous nature of the karstic dolomite bedrock was the dominant ›ntrol on the observed spatial variability in streamflow generation. This thesis is consistent with the results : field investigations using natural tracers, reported in a companion paper. Bedrock structure and :hology may affect streamflow generation directly (via water movement through fractured rock), and directly (by influencing the slope and thickness of the overlying soil). While the West Fork contains all ıe topographic and surface hydrologic features of larger basins (ridge tops, valleys, hollows, spurs, ›hemeral and perennial stream channels), it covers an area which is relatively small with respect to the ›drock heterogeneity. Therefore, while the hydrologic processes observed on the West Fork are no doubt 'pical of those occurring elsewhere in karst terrain, the particular patterns of spatial and temporal ıriability observed are somewhat specific to the study site.

NTRODUCTION

The study of streamflow generation is an important topic in environmental cience, both because of its direct relevance to the hydrologic cycle and

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because the water quality of streams is directly affected by the nature of the flowpaths supplying water to the streams. Consideration of spatial variability is a necessary component of streamflow generation studies, both in the field and on the computer. Since the number of sites at which measurements can be made is always limited in practice, it is useful to identify areas of high streamflow production so that measurements may be concentrated in those areas. In addition, it is essential to have some idea how representative a measurement is if it is used to infer the behavior of the surrounding area.

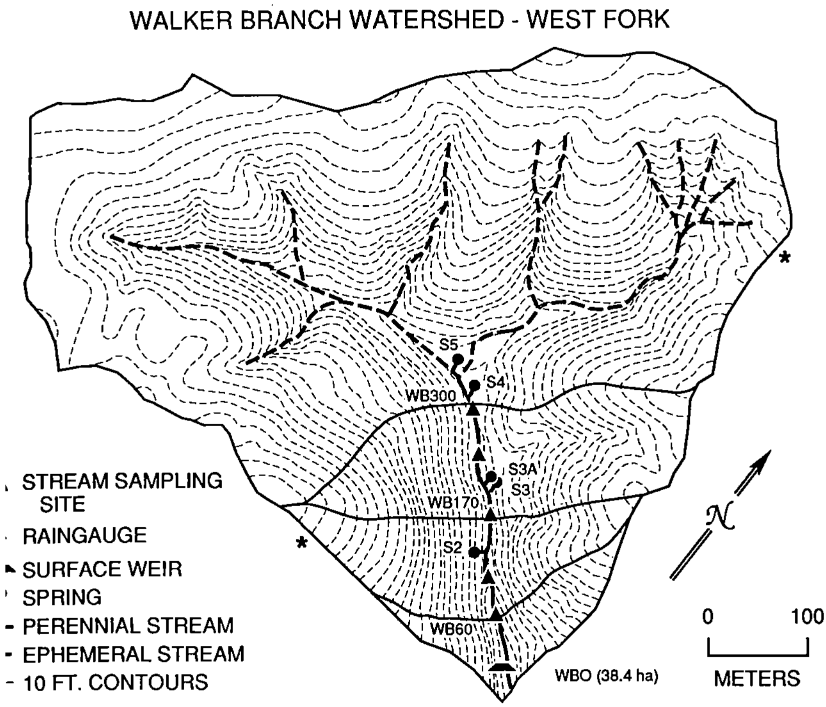
Also, comparison of the spatial pattern of stream inflow with that of other watershed parameters can elucidate the controls on streamflow generation. For a study site in southern England, Anderson and Burt (1978) found that all stream reaches of high lateral inflow were adjacent to hillslopes with concave, hollow-shaped topography. Concave topography led to convergent drainage pathways, which in turn led to saturated conditions and higher overall streamflow production in the hollows.

Spatial variability in streamflow generation is also a concern in modelling. Wood et al. (1988) proposed the notion of a 'Representative Elementary Area' (REA), the smallest basin size for which the variance of a basin output or response (e.g. volume of storm runoff for a particular rain event) is a minimum. The REA would be 'a fundamental building block for catchment modelling' (Wood et al., 1988, p. 31) since at this scale the particular pattern of heterogeneity in the study basin becomes unimportant, and the basin response can be analyzed in terms of the statistics of the underlying parameter distributions. Thus, an estimate of the REA would be useful in deciding the appropriate basin size for a modelling study.

This paper describes measurements of the spatial variability in streamflow generation on the West Fork of Walker Branch Watershed, a forested watershed in Oak Ridge, Tennessee. Chemical dilution stream gauging was used to determine streamflow at a number of points in the small stream draining the study site, over a wide range of flow conditions. A companion paper (Genereux et al., 1993) reports the results of simultaneous work with natural tracers.

STUDY SITE

The West Fork of Walker Branch has an area of 38.4 ha and an elevation range of 78 m (Fig. 1). Precipitation is measured with two weighing-type rain gauges on the ridge top; average annual precipitation is 140 cm. Streamflow at the basin outlet is monitored by means of a 120° V-notch weir with automatic stream stage recording (5 min interval). Vegetation is dominated by oak and hickory, with scattered pines on the ridges and mesophytic



g. 1. Contour map of the study site. Major stream sampling sites are indicated by triangles; stream sites e designated with the prefix 'WB' followed by their distance upstream of the weir in meters (e.g. WB300 the point 300 m upstream of the weir). The two unlabeled triangles represent WB100 and WB242. Solid es normal to the elevation contours and passing through WB60, WB170, and WB300 indicate the undaries of 'apparent contributing areas' (ACAs) for the stream reaches they define (that is, contributing eas defined in the usual way, on the basis of topography).

ardwoods, such as tulip poplar and beech, near the stream channels ohnson, 1989).

The West Fork is located on the southeast slope of Chestnut Ridge, one of any NE–SW-trending subparallel ridges in this area of the southern Appalchians. Bedrock consists of highly fractured dolomite (with some chert beds) f the Knox Group. Strata dip to the southeast at about 35° and strike along le long axis of Chestnut Ridge (roughly N 55° E), approximately normal to le study stream on the West Fork (Crider, 1981; Lee et al., 1984). The West ork is underlain by three different formations of the Knox; from oldest to oungest (i.e. going southeast along the perennial stream towards the weir) ley are the Chepultepec, the Longview and the Kingsport dolomites (Fig. 2). he Longview is much thinner than the other two formations and is more esistant to weathering because of its significantly higher chert content ietzke et al., 1989; Lietzke, 1990). All three formations are highly fractured;

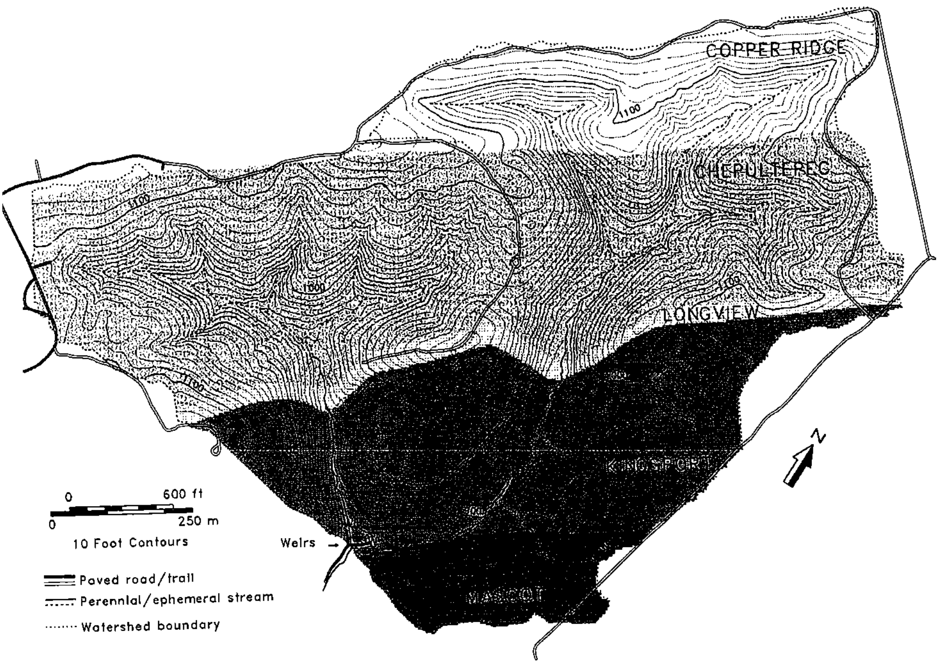


Fig. 2. Map of Walker Branch Watershed showing the different bedrock formations. The West Fork catchment shown in Fig. 1 is the triangular basin occupying the left side of this map (the East Fork, on the right, makes up the rest of Walker Branch).

a study involving measurement of 210 fractures in outcrops on Walker Branch (the West Fork and the adjacent 59.1 ha East Fork) found that fractures cluster in two common orientations, roughly perpendicular to, and parallel to, the local bedrock strike (Crider, 1981). In addition to being more numerous, fractures with these orientations were also found to be the only fractures 'solutionally enlarged to conduits in outcrops on the West Fork' (Crider, 1981, p.85). As discussed below and again later in the paper, these fractures are hydrologically significant in allowing groundwater movement across surface topographic divides.

Soils on the West Fork are mainly Ultisols (mostly of the Paleudult suborder), with one small band (< 5% of the watershed area) of Alfisols (suborder Hapludalf) along the east side of the lower 150 m of the perennial stream (Lietzke, 1990). Both Ultisols and Alfisols show the effects of clay translocation: an eluvial E horizon from which clay has been removed overlies an argillic Bt horizon where the clay particles have been deposited (Soil Conservation Service, 1975). The dominant clay mineral on the West Fork is kaolinite (10–40% of the clay fraction), with vermiculite and mica present in smaller amounts. The primary distinctions between Ultisols and Alfisols are emical and mineralogical, rather than structural or physical. According to ınning and Fanning (1989, p. 264), 'Most Alfisols differ from Ultisols in ᴌving a naturally higher base saturation (> 35% for the former, < 35% for e latter) — thus, commonly higher pH values — generally lower chromas ɩd yellower hues for the well-drained counterparts, and clay usually ntaining less 1:1 and more 2:1 layer silicate minerals'. Narrow zones of ɪtisols, soils which lack distinct pedogenic horizons, are found along the ream channels (both ephemeral and perennial).

In many places on the West Fork a thick layer of saprolite (residual clayey aterial formed in place by weathering of the dolomite bedrock) lies between e 1–2 m thick forest soil and the bedrock. This saprolite consists mainly of ıolinite (30–55%), mica (10–25%), vermiculite (5–20%) and iron oxide -8%) in the clay size ($<2\,\mu$m) fraction (Lee et al., 1984). Chert fragments e common, in places where a chert bed in the original bedrock has partially eathered but the surrounding dolomite has completely weathered to prolite. In places on the western and northern ridge tops around the West ork, the saprolite is nearly 30 m thick. Over much of the eastern side of the erennial stream valley (the same general area covered by Alfisols), there ɔpears to be little or no saprolite (small rock outcrops can be seen running ɔslope from the stream channel; well points have been driven to refusal ithin 2 m of the ground surface). The saprolite thickness is unknown or ɔorly known for much of the watershed. Previous studies have shown that perched saturated zone often develops above the soil/saprolite interface uring storms (e.g. Wilson et al., 1990).

While the saprolite may be much thicker than the overlying soil in some laces, it has a much lower maximum possible transmissivity than the soil maximum possible transmissivity' refers to the transmissivity of the layer ıturated over its entire thickness). The hydraulic conductivity of saprolite as measured in three different depth intervals (2.5–3, 6–12 and 21–30 m) at site about 6 km west of the West Fork on Chestnut Ridge (Ketelle and Huff, 984). Geometric mean conductivities were 6.1×10^{-8} m s^{-1} at 2.5–3 m (21 eld measurements), 2.0×10^{-8} m s^{-1} at 6–12 m (19 field measurements), and $.3 \times 10^{-10}$ m s^{-1} at 21–30 m (eight laboratory measurements). Assigning ıese three conductivity values to three somewhat arbitrarily chosen layers l.5–4.5, 4.5–16 and 16–30 m below ground surface) gives an estimate of 4.2 $\times 10^{-7}$ m^2 s^{-1} for the maximum possible transmissivity of the saprolite. Even the laboratory measurements of hydraulic conductivity for the deepest layer nderestimate the field scale value by a factor of 10 (which is possible), the ffect on overall transmissivity is minor (it increases to 5×10^{-7} m^2 s^{-1}). An verlying soil consisting of a 1 m thick B horizon (hydraulic conductivity of 0^{-5} to 10^{-4} m s^{-1} (Luxmoore et al., 1981)) and a 0.5 m thick A horizon (hydraulic conductivity of about 10^{-4} m s^{-1} (Peters et al., 1970)) would have a maximum possible transmissivity of 6×10^{-5} to 1.5×10^{-4} m^2 s^{-1}, 140–360 times larger than that of the 28.5 m thick saprolite.

Streamflow and rainfall data for the West Fork indicate that the watershed receives subsurface inflow from outside its topographic boundary. For the 12 years from 1969 through 1980, the average difference between annual rainfall and annual runoff at the weir was 34 cm (SD = 16 cm, n = 12). The best estimates of annual evapotranspiration (based on the average difference between rainfall and runoff for a number of watersheds in the Oak Ridge area) are about 73 cm (McMaster, 1967; Tennessee Valley Authority (TVA), 1972). Thus, the West Fork stream seems to be receiving about 1.5×10^5 m^3 of water each year (39 cm × 38.4 ha) from outside the West Fork boundary. Solutionally enlarged fractures and cavities in bedrock are the most likely conduits for this interbasin transfer.

METHODS

Chemical dilution gauging was used to determine the streamflow rate at a number of sites in the perennial stream channel; experiments spanned the range of Q_{weir} (streamflow at the weir) from 354 to 3457 l min^{-1}. The five main sites for streamflow determination are indicated with triangles in Fig. 1. The sites, each named with the prefix 'WB' followed by its distance upstream of the weir in meters, are WB300, WB242, WB170, WB100 and WB60. A flume installed in the stream channel at WB300 allowed us to check (or, occasionally, forego) chemical dilution streamflow measurements at that site. In addition to these five main sites, measurements were occasionally made at six additional sites (WB280, WB260, WB220, WB140, WB120 and WB80) in order to get more detailed information on the spatial structure of the stream inflow.

The chemical dilution methodology involves a one-dimensional steady-state analysis (e.g. Genereux and Hemond, 1990). A 50 l Mariotte bottle was used to inject a concentrated (2.7–3 mol l^{-1}) NaCl solution into the stream at a steady rate. The injection site varied, but was always 12–17 m upstream of the nearest measurement site. Since the channel was laterally well mixed (no vertical or horizontal tracer gradients), the streamflow Q at a stream measurement site was given by the simple steady-state relationship:

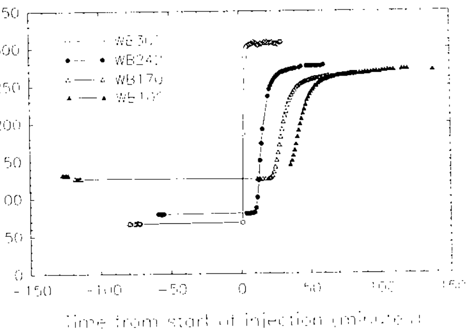
$$Q = Q_0 C_0 / C$$

where Q_0 and C_0 are the injection rate and concentration of the NaCl tracer solution, respectively, and C is the steady-state tracer Cl^- content of the streamwater at the site.

C values were estimated by two different means: field measurement of the ɔecific conductance (γ) of the streamwater and laboratory measurement of ıe Cl^- content (S) of the streamwater. Before beginning the NaCl injections ıe background specific conductance of the streamwater (γ_b) was determined t each measurement site. A battery powered hand-held conductivity meter Cole–Parmer Instrument Co. model 1481-60) with a gold dip cell was used; ıe meter automatically corrected the measured conductance to an equivalent alue at 25°C. Measurements were generally taken 30 s apart for several ıinutes in order to determine if there was any drift in γ_b. Prior to the three ighest flow experiments γ_b was slowly drifting upward at WB170, WB100 nd WB60. The γ_b drift rate at the three sites was noted and used to extrap- late γ_b forward in time to when γ_s (the steady-state plateau γ value) was ıeasured, in order to better estimate the background specific conductance nder the plateau. The drift in γ_b was associated with a slow decrease in Q_{weir} naximum rate of Q_{weir} decrease was -1.7% h^{-1}, during the highest flow xperiment on 18 March 1990). Strictly speaking, the chemical dilution ıethodology employed here requires that the streamflow be steady. In ractice, the small drop in streamflow observed during the highest flow xperiments introduces a negligible uncertainty into the calculated Q values see Appendix). After measuring γ_b at a particular site, water samples (two or hree) were collected for analysis of background Cl^- content (S_b); the nfiltered samples were collected by rinsing and filling small polyethylene ottles with streamwater. We did not attempt to measure drift in S_b because oth S_b (about 0.02 mmol l^{-1}) and changes in S_b associated with changes in treamflow (<0.01 mmol l^{-1}; Mulholland et al. (1990)) were much smaller han the steady-state chloride concentrations resulting from the tracer hloride additions (0.5–5.0 mmol l^{-1}).

NaCl injections were started after γ_b measurements were made and ackground water samples were collected. The specific conductance at each neasurement site rose and eventually leveled off during the injection (Fig. 3). Steady state was generally considered to have been attained when a constant reading (within the instrument resolution of 1 μS cm^{-1}) was obtained or $\geqslant 10$ min (or γ on the plateau was drifting at the same rate as γ_b). After the pecific conductance reached steady state three or four water samples were ollected for Cl^- analysis (S_s). Both $S_s - S_b$ and $\gamma_s - \gamma_b$ were used as measures of C in calculating Q values with eqn. (1).

Q_0 values for use in eqn. (1) were measured in the field with a stopwatch ınd graduated cylinder; C_0 for each tracer solution was determined from the nown amounts of NaCl and distilled water used to prepare the solution. In letermining the uncertainty in Q, the uncertainties in Q_0 and C_0 were ıegligible compared with the uncertainty in C (see Appendix).



. 3. Specific conductance data for the experiment of 4 October 1989. The steady NaCl injection was un at time zero.

Chloride concentrations (both S_s and S_b) were measured with an tomated ferricyanide method (U.S. Environmental Protection Agency PA), 1983), using a Technicon TRAACS 800 auto-analyzer. The steady-ıte (S_s) samples were diluted with doubly distilled water prior to analysis, bring their Cl^- contents into the range spanned by the calibration ındards. The difference $S_s - S_b$ was calculated and used directly as C in eqn.).

In using $\gamma_s - \gamma_b$ values in eqn. (1) to determine Q values, it is necessary to ›nvert the $\gamma_s - \gamma_b$ value from at least one measurement site to an equivalent l^- concentration. Measuring the specific conductance of the tracer solution $_0$) and expressing the tracer injection rate as $Q_0\gamma_0$ (instead of $Q_0 C_0$) would ɔt suffice because the relationship between γ and NaCl concentration is not ıear over the full range from γ_0 (roughly 1.7×10^5 μS cm^{-1}) to $\gamma_s - \gamma_b$ ypically 100–400 μS cm^{-1}); therefore $Q_0\gamma_0 \neq Q(\gamma_s - \gamma_b)$. Data from Jones 912) was used to convert the $\gamma_s - \gamma_b$ value from the first measurement site hat farthest upstream; either WB300 or WB242) to a Cl^- concentration. A near regression of Jones' five data points (at 25°C) in the range $9 < \gamma < 880\,\mu$S cm^{-1} gave the following relationship:

$$' = 8.91 \times 10^{-6}\gamma - 2.97 \times 10^{-5} \quad (r = 0.99997) \tag{2}$$

'here C is concentration in mol l^{-1} and γ is specific conductance in μS cm^{-1}. ıfter using eqn. (2) to convert the $\gamma_s - \gamma_b$ value from the first site to a Cl^- oncentration, and using eqn. (1) to calculate Q at that site, streamflow values t the other sites downstream were calculated by using $\gamma_s - \gamma_b$ values directly ı eqn. (1):

$$Q_j = Q_1(\gamma_s - \gamma_b)_1/(\gamma_s - \gamma_b)_j \tag{3}$$