Comparative geomorphic analysis of surficial deposits at three central Appalachian watersheds: implications for controls on sediment-transport efficiency

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Abstract

Factors that control the routing and storage of sediments in the Appalachian region are poorly understood. This study involves a comparative geomorphic analysis of three watersheds underlain by sandstones and shales of the Acadian clastic wedge. These areas include the Fernow Experimental Forest, Tucker County, West Virginia; the North Fork basin, Pocahontas County, West Virginia; and the Little River basin, Augusta County, Virginia. GIS-based analyses of surficial map units allow first-order approximation of valley-bottom storage volumes. Volume estimates are examined in tandem with morphometric analyses and bedrock-channel distribution to make inferences regarding controls on sediment-transport efficiency in the central Appalachians.

The Fernow and North Fork areas are characterized by V-shaped valleys with mixed alluvial-bedrock channel reaches distributed throughout the drainage network. In contrast, the Little River valley is notably wider and gravelly alluvial fill is abundant. Comparator watershed parameters for the Fernow, North Fork, and Little River areas include, respectively: (1) basin area = 15.2 km^2, 49.3 km^2, 41.5 km^2; (2) basin relief = 0.586 km, 0.533 km, 0.828 km;

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(3) drainage density = 4.2 km^{-1}, 3.3 km^{-1}, 4.7 km^{-1}; (4) ruggedness = 2.5, 1.7, 3.9; (5) Shreve magnitude = 139, 287, 380; (6) total valley bottom area (km^2) = 0.76 km^2, 1.86 km^2, 3.09 km^2; (7) average hillslope gradients = 17.2°, 18.4°, 22.1°; (8) total debris-fan surface area = 0.113 km^2, 0.165 km^2, 0.486 km^2; and (9) debris-fan frequency = 2.0 km^{-2}, 1.0 km^{-2}, 2.8 km^{-2}.

Valley-bottom storage volumes were estimated using map polygon areas and surface heights above channel grade. The Little River contains significantly higher volumes in floodplain, terrace, and fan storage compartments; total valley-bottom volumes are approximately twice that of the Fernow and North Fork areas combined. Unit storage volumes for the Fernow, North Fork, and Little River are 5.2 x 10^4 m^3 km^{-2}, 5.5 x 10^4 m^3 km^{-2}, and 1.6 x 10^5 m^3 $km^{-2,}$ respectively.

A conceptual model postulates that valley-width morphometry and style of hillslope delivery are the primary factors controlling sediment-transport efficiency. Steep, debris-flow-prone hillslopes at the Little River deliver high volumes of gravelly sediment at magnitudes greater than channel-transport capacity. Stream-power patterns are complex, as low-order tributaries are under capacity, and high-order tributaries over capacity with respect to sediment load. Aggraded alluvial fill insulates valley-floor bedrock from vertical erosion, while valley widening dominates. Valley-width expansion creates a positive response via increased storage capacity and lower unit stream power. Conversely, the Fernow and North Fork are characterized by diffusive mass wasting on hillslopes with incremental bedload transport to higher-order tributaries. Hillslope delivery rates are balanced by the rate of channel export. Mixed alluvial-bedrock reaches provide the optimal channel configuration for active valley-floor incision. Low valley-width expansion promotes high unit stream power and vertical erosion processes. The model implies that the Fernow and North Fork have been more effective at sediment transport

during the late Quaternary. Given similar climatic and tectonic settings, bedrock lithofacies variation is likely the primary factor modulating transport efficiency.

Keywords: Surficial mapping; Sediment storage; Sediment-transport efficiency; Appalachians

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1. Introduction

Study of the production, transport, and storage of sediment in drainage basins is essential for understanding their evolution and geomorphic behavior. Sediment routing and storage models for mountain watersheds provide a valuable technique for assessing the impacts of complex variables on fluvial systems (Schumm, 1977; Dietrich and Dunne, 1978; Bull, 1991). In addition, geomorphic analysis establishes the fundamental framework for monitoring the response of forest ecosystems to extrinsic variables such as climate and anthropogenic activity (Swanson, 1980; Swanson and Franklin, 1988).

The central Appalachian Physiographic Province represents an unglaciated, humid-mountainous landscape that has a long history of catastrophic flooding and debris flow (Mills et al., 1987). Forested drainage basins export sediment by colluvial and alluvial processes in high-gradient channel systems. Mechanisms for routing and storage of surficial sediments in this region are poorly understood and have received little attention in the literature. This study involves a comparative geomorphic analysis of three watersheds underlain by interbedded sandstones and shales of the Acadian clastic wedge. These areas include the Fernow Experimental Forest, Tucker County, West Virginia; the North Fork basin, Pocahontas County, West Virginia; and the Little River basin, Augusta County, Virginia (Figure 1). The Little River was subject to catastrophic flooding, slope failure, and debris flow as triggered by a convective storm during June of 1949. The classic paper by Hack and Goodlett (1960) documented the geomorphic response to this event.

Although sediment yield data are presently not available for the study areas, GIS-based analysis of surficial map units provides a proxy for assessing sediment-transport efficiency. This paper focuses on valley-bottom storage compartments and presents a comparison of sediment storage volumes. We derive a conceptual model that relates local aggradation to hillslope delivery mechanisms and stream-power distribution. Results have important implications for regional controls on sediment-transport efficiency in small mountain catchments (<60 km^2).

2. Study areas

2.1. Physiographic setting

The Fernow Experimental Forest lies in the Allegheny Mountain section of the unglaciated Appalachian Plateau province, 40 km west of the Allegheny structural front (Figure 1). This research facility serves as the Timber and Watershed Laboratory for the USDA Northeastern Forest Experiment Station. The Fernow occupies 19 km^2 of Monongahela National Forest and elevation ranges from 533 to 1,112 m AMSL. The North Fork area also lies in the Allegheny Mountain section, but only 4 km west of the Allegheny structural front (Figure 1). The 50 km^2 area has elevations ranging from 853 to 1,386 m AMSL. The Little River basin occupies 40 km^2 in the Valley and Ridge province of central Virginia, 40 km east of the Allegheny structural front (Figure 1). Mountain slopes are steep with elevations ranging from 500 to 1340 m AMSL.

The climate of all three areas is classified as humid continental (Lockwood, 1985). Average annual precipitation ranges from 1450 mm at the Fernow area to 1000 mm at the Little River. Regional weather systems are directed primarily from the west, with mid-latitude frontal systems and large-scale cyclonic disturbances common (Lee et al., 1977; Hirschboeck, 1991).

The region is sporadically prone to torrential rainfall associated with the extratropical-phase of hurricanes and terrain-locked convective clusters (Michaels, 1985; Colucci et al., 1993; Smith et al., 1996). Site soils are quite similar (Mesic Typic Dystrochrepts) and commonly form on bedrock hillslopes mantled with colluvium and residuum (Losche and Beverage, 1967; Hockman et al., 1979; Flegel, 1999). Each area is heavily forested by deciduous, coniferous, and mixed-forest communities (Braun, 1950; Hack and Goodlett, 1960; Core, 1966).

2.2. Surficial geology

Landforms in the central Appalachians are divided into hillslope and valley-bottom features. The hillslope regime can be further subdivided into noses, side slopes, and hollows (Hack and Goodlett, 1960; Taylor et al., 1996). Valley bottoms include channels, floodplains, fans, and terraces. Colluvium and residuum are the most widespread surficial deposits, commonly occurring as a veneer of gravel diamicton on hillslopes (Mills and Delcourt, 1991).

Zero-order hillslope hollows can be significant sediment storage components in mountain watersheds (Dietrich and Dunne, 1978) and form an important facet of Appalachian landscapes (Hack and Goodlett, 1960). Local topographic convergence directs colluvium into hollows, with transport rates proportional to degree of convergence (Reneau et al., 1990). Hollows accumulate hillslope regolith via diffusive mass-wasting processes such as creep, tree throw, and bioturbation. The sediment mass is subsequently delivered to higher-order tributaries by a spectrum of hydraulic processes ranging from streamflow to hyperconcentrated flow to debris flow (Pierson and Costa, 1987). Once in the primary fluvial system, turbulent channel flow exports sediment out of the watershed as dissolved, suspended, and traction load. Each of these processes routes sediment into and out of storage, depending on energy level of the system.

Jacobson et al. (1989a) argued that catastrophic flooding and debris flow are the most effective geomorphic agents in the central Appalachian region. High-magnitude precipitation events and attendant debris flow processes are the driving mechanisms for transfer of surficial material between hillslope and valley storage compartments (Springer et al., 2001; Eaton et al., 2003a). Cenderelli and Kite (1998) documented the efficiency with which Appalachian debris flows route low-order tributary deposits to higher-order valley bottoms. The occurrence of debris flow, in turn, dramatically influences the spatial and temporal distribution of channel alluvium (Benda and Dunne, 1987; 1997). A geomorphically significant debris-flow event impacted the Little River basin on June 17-18, 1949 (Stringfield and Smith, 1956; Hack and Goodlett, 1960). The watershed received greater than 175 mm of rain during a 24 hour period. The intense rainfall associated with a multi-cell convective storm triggered more than 100 debris slides, dramatically modifying valley-bottom topography. Debris fans and slide scars generated by that event are a conspicuous component of the present-day landscape (Osterkamp et al., 1995; Taylor, 1999).

Debris fans are ubiquitous throughout the central Appalachian region, and are associated with low-order drainage basins in erosion-dominated landscapes (Kochel, 1990; Mills, 2000). These landform elements are characterized by small-scale (< 1 km^2), irregularly shaped deposits that commonly occur at valley intersections. Debris fans are constructed primarily by a combination of debris-slide and debris-flow processes in steep mountain watersheds. Sedimentary facies are characterized by poorly sorted, crudely-stratified diamictons, with lesser amounts of clast-supported gravels (Kite, 1987; Kochel, 1992). Similar fan occurrences were documented in other regions by Benda and Dunne (1987), Wells and Harvey (1987), and Kellerhals and Church (1990).

2.3. Bedrock geology

Each of the study areas is underlain primarily by sandstones and shales of the Acadian clastic wedge, covering 80% to 100% of total watershed area (Figure 2). This mega-facies ranges in age from Devonian to Mississippian and includes the Foreknobs, Hamsphire, and Price (Pocono) formations (Boswell, 1988). These strata are interpreted as progradational, shallow-marine to nonmarine sequences derived from eastern tectonic highlands during the Acadian orogeny (Boswell et al., 1987; Dennison, 1988). Although there is local variation, the structure of each area is remarkably similar in that folds are broad and open, with dips less than 10 degrees. The similarity of mapped lithostratigraphy was the common variable used in selecting the three respective watersheds for comparative analysis.

In the central Appalachians, varying lithologic resistance to erosion is the main influence on topographic development of mountainous landforms. Erosionally resistant sandstones of the Price (Pocono) Formation are notable ridge formers in the study areas (Figure 2). High relief and altitude are invariably associated with resistant lithology (Hack, 1979; Germanoski, 2001; Mills, 2003). Quartzo-feldspathic sandstones of the Acadian clastic wedge support numerous mountain slopes throughout this region, providing a dynamic geomorphic environment associated with slope failure, debris flow, and flooding. In addition, sandstone lithologies comprise the bulk of framework gravel contained in colluvial diamicton and channel alluvium.

3. Research approach

Watershed sediment budgets are constructed by quantifying the principal processes responsible for production and transport of surficial material. The task involves monitoring hillslope and valley-bottom transport processes, with recognition of the linkages between transfer and storage elements (Dietrich and Dunne, 1978; Trimble, 1983; Reid and Dunne, 1996). Essential components of sediment budgets include quantification of transport processes, storage elements, and time-averaged sediment yields (Dietrich et al., 1982). Sediment budgets in their pure form are time consuming and costly, requiring many years of data collection. Process-based surficial mapping provides a relatively rapid, cost-effective method for delineating storage compartments and routing mechanisms in mountain watersheds (Taylor, 1999). The spatial and temporal distribution of sediment storage compartments provides insight into the relative efficiency with which watersheds export sediment.

The present study focuses on the spatial variation of sediment storage at the three study areas described above. Valley-bottom compartments are the focus of the analysis as they can be mapped in detail and reflect the ability of watersheds to export sediment. Temporal rates of sediment transfer require long-term monitoring and are beyond the scope of this work. Detailed surficial mapping, combined with GIS-based analyses, allow first-order approximation of valley-bottom storage volumes. Volume estimates are examined in tandem with morphometric analyses and bedrock-channel distribution to make inferences regarding controls on sediment-transport efficiency.

Surficial mapping methodology follows the protocol established by Taylor et al. (1996). Map units were identified at a scale of 1:9,600 and delineated on the basis of age, origin, landform, and material (Table 1; Figure 3). Surficial data were compiled using county soil surveys, natural exposures, topographic analysis, and aerial photography. Spatial data were manually digitized and incorporated into a GIS database using a combination of AutoCAD, Idrisi, and ArcView.

A standard suite of morphometric parameters (Horton, 1945; Strahler, 1957) were also computed using 10-m digital elevation models, vectorized topographic maps, and GIS techniques. The channel network was manually digitized from 7.5-minute map sheets, with channels identified by the contour-crenulation method (Strahler, 1957).

4. Results of comparative geomorphic analysis

4.1. Morphometric parameters

Table 2 summarizes the results of morphometric analyses conducted at each of the study areas. The data indicate that the Little River watershed is morphologically distinct compared to the Fernow and North Fork areas. By comparison, Little River is a rugged, high-relief watershed with a dense channel network, wide valley bottom, and a high number of intersecting tributary nodes. Its ruggedness is also manifested in hillslope and channel gradients, displaying significantly steeper slopes compared to those at the Fernow and North Fork (z-test rejects null hypothesis at $\alpha = 0.05$).

4.2. Storage compartments

Sediment-storage compartments are identified on the basis of landform, material, and deposit geometry (Table 1). Hillslope compartments include ridge-top residuum (Qr), nose-side slope colluvium (Qc1), and hollow colluvium (Qc2). Valley-bottom compartments include channel alluvium (Hch), floodplain alluvium (Hfp), terrace alluvium (Qt), and fan sediments (Qf). Colluvial aprons (Qap) lie at the base of side slopes and represent a zone transitional to the valley-bottom regime ("footslope" deposits of Hack and Goodlett, 1960). Figure 4 summarizes the areal distribution of surficial map units at each of the study areas. Hillslope colluvium (Qc1 and Qc2) accounts for 80 percent of all map polygons. Total valley-bottom compartments account for only 5 to 10 percent of total watershed area. The Little River contains a significantly greater percentage of valley-bottom area compared to the Fernow and North Fork. The former area also exhibits consistently higher areal percentages of floodplain, terrace, and fan compartments.

4.3. Valley-bottom storage budgets

4.3.1. Bedrock-channel maps

Distribution of bedrock-floored channel reaches, drilling records, and detailed surficial maps were used to estimate valley-bottom storage volumes at each of the study areas. Figure 5 shows map distribution of bedrock and alluvial reaches in the respective channel systems. The Fernow and North Fork areas are characterized by V-shaped valleys with bedrock-channel exposure throughout the drainage network (Figure 6A). In contrast, the Little River valley is notably wider and gravelly alluvial fill is abundant in fourth-order and higher tributaries (Figure 6B). Reaches floored by bedrock are limited to lower-order channels in the steepest headwater zones (Figure 5). Drilling logs were obtained for the southeastern corner of the Little River area, near the mouth of the watershed (Unpublished Data, Headwaters Conservation District, Verona, Virginia). Drilling data support the map observations, documenting five to ten meters of alluvium in the lower channel reaches (Figure 5).

4.3.2. Debris fan distribution

Fan map units are delineated according to fan-surface morphology and height above channel grade (Table 1). Debris fans mapped in the study areas include both simple and compound forms, the latter of which are associated with complex map patterns and inset fan-terrace relationships (Taylor, 1998; 1999). Fan terraces are preserved segments that have been otherwise dissected by tributary channels following deposition. Debris fans are preserved at select tributary junctions, with greater than 75% occurring at the intersections of first- or second-order channels with higher order trunk streams (Figure 7). Critical quantitative parameters derived from systematic fan mapping are listed in Table 3.

Mapping and morphometric data (Tables 2 and 3, Figures 4 and 7) indicate that compared to the other two areas, Little River is a steep, rugged watershed with significantly higher drainage density and higher percentage of valley-bottom area. Accordingly, Little River also has the highest volume of fan deposits in storage. Comparatively, the Little River consistently shows greater values in the categories of valley-bottom area, number of fans, and total fan area.

These results suggest that compared to the other areas, geomorphic conditions at the Little River are more conducive to storage of sediment in fans. The Little River is associated with the highest values of maximum valley width, drainage density, tributary-junction frequency, fan frequency, and average fan area. Wide valleys, coupled with a high number of tributary intersections, provide numerous storage sites for fan deposits. The less frequent occurrence of debris fans at the Fernow and North Fork suggests that these fluvial systems are more effective at routing sediments out of the watershed.

4.3.3. Storage volume estimates

Valley-bottom storage volumes were estimated using map polygon areas and surface height information. Figure 8 is a generalized cross section illustrating the methodology employed in volume calculations. Bedrock-channel exposure at the Fernow and North Fork areas indicates that floodplain and terrace heights above channel grade are viable approximations of deposit thickness. Map polygon volumes were derived from the product of polygon area and surface height. In cases where surface heights were estimated within a range, the midpoint of the range was used in the volume calculation. The procedure was repeated for each valley-bottom polygon within the respective watershed, the sum of which represents total "surface volume" above channel grade (Figure 8). In the case of the Little River, an additional "fill volume" was calculated to account for the wedge of sediments preserved below channel grade. Drilling data provide an estimate of fill thickness at the mouth of the watershed. The sediment wedge is assumed to thin uniformly upstream, approaching zero thickness at the first point of bedrock-channel exposure (Figure 5C). The kriging algorithm in Surfer for Windows (Golden Software, 1997) was used to contour upper and lower surfaces on the alluvial wedge, with the "fill volume" calculated by numerical subtraction of the bounding grids.

Figure 9 summarizes the results of valley-bottom storage analysis. The Little River contains significantly higher volumes in floodplain, terrace, and fan storage compartments; the total valley-bottom volume is approximately twice that of the Fernow and North Fork areas combined. Unit storage volumes were calculated by dividing total valley-bottom storage by drainage area. Unit values for the Fernow, North Fork, and Little River are 5.2×10^4 m^3 km^{-2}, 5.5×10^4 m^3 km^{-2}, and 1.6×10^5 m^3 km^{-2}, respectively. Thus, unit valley-bottom storage at the Little River is three times that of the Fernow or North Fork. The distributions of total valley-bottom area (Table 2), bedrock-channel relationships (Figure 5), and unit storage volume (Figure 9) suggest that the Little River is over transport capacity with respect to sediment load, while the Fernow and North Fork areas are at equilibrium or slightly below capacity.

4.4. Residence time

Dating of surficial deposits in the Appalachians is problematic and persists as an elusive facet of geomorphic study. Vegetative growth patterns provide an important dating tool for historic sediments (Yanosky, 1982; Osterkamp et al., 1995), however the ages of older deposits are largely unknown. Radiocarbon techniques are of limited value due to poor preservation of organic matter (Mills and Delcourt, 1991). Thermoluminescence (Shafer, 1988), magnetostratigraphic (Jacobson et al., 1988; Springer et al., 1997), and cosmogenic isotope (Pavich et al., 1985, 2005; Granger et al., 1997; Matmon et al., 2003) techniques provide results holding some promise; however, they have not been widely applied in the Appalachians. Relative-age dating techniques were utilized in several studies (Mills, 1988; Mills and Allison, 1995; Engel et al., 1996; Mills, 2005), although the discontinuous nature of surficial deposits and varying lithologic composition make regional correlation difficult without numerical calibration.

Recent studies have addressed recurrence intervals for debris flow in the central Appalachians. Based on radiocarbon dating, Kochel (1987) estimated recurrence intervals of 3,000 to 4,000 years on Holocene fans in the Virginia Blue Ridge. He postulated that incursion of tropical cyclones at the Holocene interglacial transition triggered debris-flow delivery of periglacial colluvium to fans. Eaton et al. (2003a, 2003b) presented dates for prehistoric debris flows in the Virginia Blue Ridge with radiocarbon ages ranging from 2,080 to greater than 50,000 years. Their work supports Kochel's (1987) 3,000 year recurrence interval estimates, but suggests that Pleistocene periglacial climates were also capable of spawning debris flows. This latter point is also supported by work to the south in the Smoky Mountains of Tennessee (Schultz et al., 2000). Kochel (1990) surmised that these recurrence estimates provide an approximation of the time required for colluvial replenishment of hillslope source areas.

Given the lack of numerical chronologies, it is only possible to speculate on the residence time of valley-bottom deposits at the study areas. Several models have been proposed suggesting that hillslope colluvium in the central Appalachians dates to the last glacial maximum, with rates of regolith production enhanced by periglacial processes (Mills, 2005). Behling et al. (1993) obtained radiocarbon dates from a surficial sequence in the Pendleton Creek basin of northern West Virginia. They found that hillslope colluvium is between 17,000 and 22,000 years old, while floodplain alluvium is entirely Holocene in age (<10,000 ka). Jacobson et al. (1989b) found a similar chronology for colluvial and alluvial deposits in the South Branch of the Potomac. Considering the proximity of these dated sites to the present study areas, it is anticipated that similar chronologies exist. The absence of well-developed soil horizons and clast weathering rinds suggests that site valley-bottom deposits are no older than Late Pleistocene. Kochel (1990) noted that 1949 debris-flow deposits at the Little River were unaffected by record discharges in a 1985 flood event. Many fans are armored with coarse bouldery debris that will likely remain in storage until comminuted by weathering (Taylor, 1999). This observation, coupled with the high number of compound fans and terraces, suggests that valley-bottom deposits at the Little River may be older than those at the Fernow or North Fork.

Matmon et al. (2003) measured ^{10}Be in alluvium from select watersheds in the southern Appalachians. Their analysis yielded sediment generation rates ranging from 46 to 100 t km^{-2} yr^{-1}, equivalent to model erosion rates between 17 and 37 m/m.y. The Matmon et al. (2003) results match well with Cenozoic erosion estimates presented by other workers in the region (e.g., Hack, 1979; Granger et al., 1997). Assuming similar sediment generation rates for the sites considered in this study, sediment storage volumes presented on Figure 9 require residence times ranging from 1400 to 9400 years (ρ = 2.7 g cm^{-3}). Additional work is required to definitively establish age of surficial deposits.

5. Discussion

5.1. Stream-power distribution

Bull (1979, 1988, 1991) promoted the analysis of driving and resisting forces as a method for understanding watershed dynamics. Geomorphic work occurs when driving forces exceed resisting thresholds. Driving forces are controlled primarily by gravity and climate, manifested quantitatively as stream power. Bagnold (1966) defined stream power as the kinetic energy available for erosion and transport:

$$\Omega = \gamma Q S \tag{1}$$

where Ω is stream power per unit length of channel (watts = Nm/s), γ is specific weight of the fluid (~9800 N/m^3), Q is stream discharge (m^3/sec), and S is energy slope (approximated by channel slope in m/m). Unit stream power is further defined in terms of power per unit area across the channel bed, and quantifies the rate of loss of potential energy as water flows down gradient (Ferguson, 2005):

$$\omega = (\gamma Q S)/w = \gamma D V S = \tau V \tag{2}$$

where ω is unit stream power (watts/m^2), w is flow width (m), D is mean flow depth (m), V is mean velocity (m/sec), and τ is shear stress acting tangential to the channel boundary (N/m^2). Discharge increases as a power function of drainage area in a watershed (Dunne and Leopold, 1978). A general relationship is described by the equation:

$$Q = c A_{\mathrm{d}}^{n} \tag{3}$$

where A_{d} is drainage area (km^2). The parameters c and n are determined by hydrologic factors such as season, vegetation, surficial materials, and bedrock geology. By substitution of equation 3 into equation 2, potential unit stream power at a given watershed position depends on drainage area, channel slope, and flow width:

$$\omega = (\gamma (c A_{\mathrm{d}}^{n})\, S)/w \tag{4}$$

Thus for a given drainage area, an increase in channel slope or reduction of flow width produces a proportionate increase in unit stream power, the energy available for valley-floor erosion. Accordingly, time-averaged transport in wider valleys results in lower net stream power, promoting sediment storage in floodplains and terraces (Miller, 1990; Magilligan, 1992). Baker and Costa (1987) concluded that drainage basins with areas between 10 and 50 km^2 tend to maximize stream power.

Bull (1979, 1991) reasoned that the net balance between stream power and resisting power drives aggradation and degradation in watershed systems. Stream power represents the actual energy available for sediment transport; whereas resisting power represents the critical threshold energy needed to initiate transport. As illustrated above, stream power is a direct function of discharge and channel slope. Resisting power represents channel elements that control energy expenditure such as bed roughness, sediment volume, clast diameter, and woody debris. Energy is expended in the fluvial system by maintaining fluid flow against resisting elements and transporting sediment. Hence, the ratio of stream power to resisting power affects watershed response according to the following relationships:

1. stream power < resisting power = aggradation,
2. stream power > resisting power = degradation, and
3. stream power = resisting power = graded .

The critical power threshold may change abruptly with the volume and size of sediment load delivered from the hillslope subsystem. Bedrock-floored channels suggest that fluvial systems are under capacity with respect to sediment loads, while alluvial-dominated channels imply stream power deficits (Howard and Kerby, 1983; Montgomery et al., 1996). High sediment supply results in alluvium-covered reaches; reduced supply or increased transport capacity induces erosion to bedrock (Montgomery et al., 1996).

Storage volume estimates and bedrock-channel distribution permit qualitative assessment of stream power conditions at the study areas. The Fernow and North Fork are characterized by mixed alluvial-bedrock reaches with unit storage volumes on the order of 5×10^4 m^3 km^{-2} (Figure 6A). Comparatively, the Little River is dominated by alluvial channel reaches and unit storage volumes are on the order of 2×10^5 m^3 km^{-2} (Figure 6B). Map data indicate that the Fernow and North Fork are associated with average alluvial thicknesses ranging from two to three meters, while drilling data at the Little River documents up to ten meters in the main channel. These data record a greater volume of surficial deposits in storage at the Little River, suggesting that the Fernow and North Fork have been more effective at sediment transport during the late Quaternary. Mixed alluvial-bedrock reaches in the latter two channel systems imply relative equilibrium between stream power and critical power. Little River stream power patterns are more complex, as low-order tributaries are under capacity, and high-order tributaries over capacity with respect to sediment load. The following is a summary of evidence suggesting net stream-power deficit at the Little River:

1. valley-bottom storage volume greater than can be transported by individual high-discharge flood events,

2. floodplain width significantly less than total valley-floor width,

3. pervasiveness of alluvial-dominated reaches in higher order tributaries, with noted absence of bedrock exposure in channel bottoms,

4. abundance of debris-flow dominated fan deposits, and

5. three-fold increase of unit valley-bottom storage compared to the Fernow and North Fork.

5.2. Storage factors

5.2.1. Valley-bottom accommodation

Taylor (1998, 1999) developed a model for sediment accommodation in Appalachian landscapes. Drainage density and valley-width geometry are interpreted as the primary morphologic parameters providing storage capacity at the study areas. These dependent variables are in turn controlled by climate and bedrock geology (mechanical resistance to erosion). Figure 10 is a plot of valley width as a function of channel distance from drainage divide. The Fernow is associated with the narrowest valley bottoms and lacks significant correlation between the two parameters ($R<0.3$). In contrast, the North Fork and Little River areas display a positive linear relation with slopes of 0.01 and 0.02, respectively ($R = 0.87$ and 0.91). Slope of the linear regression represents the rate of valley width increase per unit channel distance, and is termed the "valley-width expansion rate". The Little River exhibits the highest values of maximum valley width, valley-width expansion rate, drainage density, and Shreve magnitude (Table 2). Increased width factors, coupled with a high number of tributary intersections, provide a large degree of accommodation space for valley-bottom sediment storage. High valley-width expansion rates also promote net reduction of unit stream power across the valley floor. The North Fork and Fernow areas are more topographically constricted, affording less accommodation space. Narrow valley bottoms generate high shear stress values during large-magnitude flood events, stimulating sediment transport and removal (Miller, 1990).

5.2.2. Woody debris

Coarse woody debris and poorly sorted gravel combine to create complex roughness elements in forested mountain watersheds (Swanson et al., 1982; Nakamura and Swanson, 1994). Kochel et al. (1987) documented the role of woody debris as a catalyst for local channel aggradation at the Little River. Woody debris is incorporated into the channel system through bank erosion, undercutting, and tree fall during high-discharge events. Once in the channel proper, woody debris dramatically increases hydraulic roughness, promoting a reduction in flow velocity. Kochel et al. (1987) introduced the term "pseudo-terrace" to describe channel reaches associated with local gravel aggradation upstream of woody debris dams. Surfaces of the aggraded reaches are commonly higher than the adjacent floodplain, resulting in an inverted channel profile. Given the wide valley bottom at the Little River, it is likely that woody debris dams play an important role in driving channel abandonment and lateral migration. The high valley-width expansion rate and woody debris content greatly increase the conditions for net aggradation. Woody debris has less influence on narrow V-shaped valleys as fallen trees are more likely to bridge above the channel, with negligible effect on channel roughness.

5.3. Valley erosion mechanisms

Hancock et al. (1998) summarized the physical processes by which rivers erode rock. The dominant processes operating at the study areas include abrasion and bed quarrying. Abrasion involves rock-surface erosion by forcible impact of entrained sediment. Bed quarrying results from plucking of preconditioned bedrock blocks under fluvial shear stress. Quarrying is enhanced by structural and mineralogical heterogeneities in the valley wall and floor (Wohl, 1998).

Rates of fluvial incision are partly governed by style of hillslope sediment delivery to the channel system. Incision rates are maximized by the optimum sediment flux that provides gravel tools for valley-floor abrasion (Bull, 1991). If hillslope supply is greater than channel transport capacity, then aggradation insulates the bedrock substrate from abrasion and plucking processes. Thus the rate of fluvial incision is inversely proportional to the volume of stored alluvial sediments in the channel system (Pazzaglia et al., 1998). A moderate supply of sediment to the channel, below transport capacity, provides the optimal tools necessary for abrasion, plucking, and bedrock incision (Sklar and Dietrich, 1998). Aggraded valley bottoms tend toward low rates of vertical incision and increased rates of valley widening (Bull, 1991).

The above erosion model can be used to infer that the potential for vertical bedrock incision is maximized at the Fernow and North Fork watersheds. Mixed alluvial-bedrock reaches provide the optimal channel configuration for active valley-floor incision. Low valley-width expansion rates promote higher unit stream power and vertical erosion processes. In contrast, higher valley-width expansion rates at the Little River imply that lateral erosion and aggradation is a dominant process in high-order tributaries. The transition from mixed alluvial-bedrock to alluvial reaches at the Little River indicates that low-order tributaries are sites of vertical incision while the main valley floor is largely insulated from erosion. Debris flow is likely the dominant erosion process in low-order channels (Taylor, 1999). From the above relationships, we infer that the Fernow and North Fork expend a large proportion of stream power incising bedrock, while the Little River dissipates energy primarily by reworking stored sediments. Assuming average regional erosion rates of approximately 30-40 m/m.y. (Hack, 1979; Granger et al., 1997; Matmon et al., 2003), both vertical and lateral incision processes must operate on time scales of 10^6 to 10^7 years to create significant change in valley morphology (i.e., topographic modification on the order of 10^2 m).

In a related study, Taylor (1999) presented the results of systematic clast-size sampling at each of the study areas. An association of poorly sorted, large-diameter clast populations with mixed bedrock-channel reaches implies that local bed quarrying is the dominant erosion process in this substrate type. Sandstone slabs dislodge from the channel bottom and become incorporated into the local gravel population, resulting in large average clast diameters and high standard deviations. The angular nature of large-diameter sandstone slabs suggests that they have not been transported far from their bedrock source and are likely derived within the channel environment. In higher-order tributaries of the Little River, the alluvial-dominated substrate is up to ten meters thick, insulating valley-floor bedrock from channel erosion processes. High-discharge flood events rework clast populations in alluvial reaches, with progressive mechanical abrasion and selective sorting in the downstream direction. Hence, the main stem of the Little River is able to internally organize the sediment load with respect to gradient and stream-power distribution. The Fernow and North Fork watersheds, conversely, are marked by mixed bedrock-alluvial reaches throughout the channel network. Local slab input impedes the effectiveness of mechanical sorting.

In studies of varying source-area lithology, other researchers have ascribed differences in valley-width morphometry to mineralogic composition and erosional resistance (Kelson and Wells, 1989; Magilligan, 1992; Pazzaglia et al., 1998; Montgomery and Gran, 2001). Erosionally resistant bedrock is generally associated with steep, rugged watersheds and narrow valleys. The relationship between sandstone lithofacies and resistant mountain slopes in the Appalachians is well known (Cooper, 1944; Hack, 1979; Mills, 2003). Higher valley-width expansion rates at the Little River imply a less-resistant bedrock lithology conducive to lateral incision and valley widening; compared to the other two sites. This interpretation is counterintuitive to lithofacies analysis that documents a high percentage of thickly-bedded sandstone intervals supporting Little River hillslopes (Taylor, 1997; 1999). Thus a conundrum exists in that wide valley bottoms imply non-resistant bedrock conditions, while steep slopes and high sandstone content suggest resistant. Two hypotheses are presented to address this apparent contradiction. The anomalous widening may be a product of inherent, long-term watershed evolution during the late Cenozoic. In this hypothesis, consistently steep, debris-flow-prone slopes yield sediment volumes greater than the channel network can export. The result is a long-term channel response in the form of lateral migration and sidewall erosion. Apparent width could be further enhanced by infilling of V-shape valleys and aggradation up adjacent hillslopes. An alternate hypothesis invokes structural weaknesses along fracture-controlled lineaments. Notable lineaments occur in the Little River watershed, subparallel with the main trend of the lower trunk channel (Figure 2) (Hack and Goodlett, 1960; Taylor, 1999). It is possible that deep-seated fractures initially provided a less resistant boundary condition predisposed to lateral incision and valley widening. Both hypotheses are viable explanations of the anomalous valley-width expansion, however neither is testable given the present data set.

5.4. Transport efficiency model

Comparative geomorphic analysis of the study areas yields a conceptual model for controls on sediment-transport efficiency in the central Appalachians. The model postulates that valley-width morphometry and style of hillslope delivery are the primary factors controlling the efficiency of sediment export. Bedrock geology and climate independently modulate these factors on time scales of 10^4 to 10^6 years. Tectonic influence is considered negligible, with post-orogenic isostacy balanced by denudation rates in this unglaciated, passive-margin setting (Matmon et al., 2003). Figure 11 is a diagrammatic illustration of model concepts. Table 4 provides a relative ranking of critical geomorphic factors upon which the model is based.

Taylor (1997; 1999) concluded that sandstone lithofacies variations in the Acadian clastic wedge ultimately control the style of surficial processes at the respective study areas. The Little River is a steep, high-relief basin with high drainage density and bedrock dominated by thickly bedded, amalgamated sandstone bodies (sandstone:shale ratio = 2.7; percent sandstone in section = 71). Steep, colluvium-covered hillslopes result in debris flow as the primary delivery mechanism. Conversely, the Fernow and North Fork areas are less rugged with a greater percentage of shale bedrock (sandstone:shale ratio = 1.2 and 1.8, respectively; percent sandstone in section = 41 and 47, respectively). Gentler hillslopes promote diffusive mass wasting and sediment transport through normal streamflow.

The Fernow and North Fork areas are characterized by mixed alluvial-bedrock channels suggesting that net stream power is equal to or greater than resisting power (Figure 11 A, B). Diffusive mass wasting is the primary hillslope-delivery mechanism, with incremental bedload transport to higher-order tributaries. Relatively low unit storage volumes imply that hillslope delivery rates are balanced by the rate of channel export. Mixed alluvial-bedrock reaches provide the optimal conditions necessary for active valley-floor incision. Low valley-width expansion rates promote high unit stream power and vertical erosion processes. The model suggests that the Fernow and North Fork have been relatively effective at sediment transport during the late Quaternary.

In contrast, the Little River is dominated by alluvial channel reaches and high unit storage volumes. Steep, debris-flow-prone hillslopes deliver high volumes of bouldery sediment at magnitudes greater than the channel transport capacity (Figure 11C). Bedrock-channel patterns suggest a complex mode of stream power distribution, with low-order tributaries under capacity with respect to sediment load, and high-order tributaries over capacity. Headwater channels are subject to episodic debris-flow scour and vertical incision. The main valley bottom aggrades as surficial deposits are stored in fans, floodplains, and terraces. Aggraded alluvial fill insulates valley-floor bedrock from vertical erosion, while valley widening dominates. Valley-width expansion creates a positive response via increased storage capacity and lower net stream power across the valley floor. Valley aggradation is further enhanced as discharge is diminished due to baseflow infiltration into gravelly alluvial fill. The net result is reduced transport efficiency at the watershed scale. Similar conditions for debris-flow-driven aggradation have been documented at other forested mountain watersheds (Kelsey, 1980; Benda, 1990; Miller and Benda, 2000).

5.4.1 Model implications

Hack (1960) applied the concept of dynamic equilibrium to landscape models for the central Appalachians. He postulated that erosional forces and resisting framework are in a steady state condition with topographic elements mutually adjusting and subject to constant rates of erosion. The landscape evolves over time in an equilibrium state, with system adjustments occurring in response to changes in independent variables such as bedrock type, tectonic deformation, and climate. Hack's (1960) model implies that under a given set of independent conditions, equal rates of erosion occur throughout the watershed network and in differing bedrock terrains, regardless of variation in mechanical strength. The balance of driving and resisting forces is manifested by topographic form of the landscape via adjustment of slope angles; steep and high relief in resistant rocks, low gradient and relief in weak rocks.

Recent cosmogenic isotope work on alluvial deposits in the Appalachians yields insight into the applicability of the dynamic equilibrium model to watershed processes in the region. As discussed above, Matmon et al. (2003) analyzed ^{10}Be content in valley-bottom alluvium of 27 watersheds in the southern Appalachians, with drainage areas ranging from 1-330 km^2. They observed that model erosion rates were highly variable in drainage basins < 50 km^2 in area (range: 17-37 m/m.y.). This variability implies that Hack's dynamic equilibrium may not applicable at the headwater scale; however the model is supported by spatially homogenous erosion rates derived for larger basins (> 100 km^2). The wholesale applicability of dynamic equilibrium to the Appalachians was further questioned by Pavich et al. (2005). New ^{10}Be

exposure data from the Susquehanna basin indicate local bedrock incision rates of 1000 m/m.y. during the last glacial maximum (~30 ka to 10 ka). The authors concluded that various disequilibrium processes have affected portions of the Appalachians over the past 20 Ma.

The transport-efficiency model presented in this study allows for spatially variable rates of erosion and aggradation in the channel system, as controlled by style of hillslope delivery, valley morphology, and sediment storage volumes. Results for the Little River basin suggest that non-equilibrium, transient conditions exist. The combination of up-gradient incision and down-gradient aggradation is not sustainable over the long term, and is implicitly inconsistent with the dynamic equilibrium concept. Hence, the transport-efficiency model is in part supported by the results of Matmon et al. (2003) and Pavich et al. (2005). Additional cosmogenic isotope work and detailed surficial chronologies at a wide range of spatial scales are needed to conclusively evaluate the applicability of steady-state equilibrium models in the central Appalachians.

6. Conclusions

The Little River basin is unique in terms of watershed morphology, channel-reach characteristics, and valley-bottom sediment storage. Comparative geomorphic analysis with two other watersheds set in similar bedrock terrains yields considerable insight into process-response mechanisms governing sediment-transport efficiency in the central Appalachians. Our conceptual model relates style of hillslope delivery and valley-width morphometry to sediment storage and valley-erosion dynamics. Steep, debris-flow-prone slopes, high drainage density, and high valley-width expansion rates promote sediment storage. Debris flows route high volumes of sediment to valley bottoms at magnitudes greater than the channel transport capacity. Aggraded alluvial fill insulates valley-floor bedrock from vertical erosion, stimulating lateral incision. Valley aggradation is further enhanced as discharge is diminished due to baseflow infiltration into gravelly alluvium. The net result is reduced transport efficiency at the watershed scale. Controlling factors are independently modulated by bedrock geology and climate at time scales of 10^4 to 10^6 years. Given similar climatic and tectonic settings, bedrock lithofacies variation is likely the primary factor controlling hillslope processes, valley accommodation space, and sediment transport efficiency in the central Appalachians.

Field-based hydraulic and sediment yield measurements are needed to assess the validity of the transport-efficiency model developed here. Cosmogenic isotope work and other chronologic analyses would be particularly useful in further evaluating the applicability of steady state, dynamic equilibrium concepts to the central Appalachians.

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Table 1
Principal surficial map units recognized at the Fernow, North Fork, and Little River study areas

Map Unit Label	Map Unit Description	Age	Origin (Process)	Landform	Material (Texture)
Qr	Quaternary Residuum	Quaternary (Undiff.)	Residuum	Ridge-Veneer	Cobble- to Boulder-Diamicton with Silty Loam Matrix
Qc1	Quaternary Colluvium (Side Slopes)	Quaternary (Undiff.)	Colluvium	Nose-Side Slope Veneer	Cobble- to Boulder-Diamicton with Silty Loam Matrix
Qc2	Quaternary Colluvium (Hollows)	Quaternary (Undiff.)	Colluvium	Hollow Veneer	Cobble- to Boulder-Diamicton with Silty Loam Matrix
Hch	Holocene Channel Alluvium	Holocene	Alluvium	Channel and Narrow Floodplain	Cobbles-Boulders and Pebbly Loam (rounded to subrounded)
Hfp1	Holocene Floodplain Alluvium (0.5 to 1.0 m surface)	Holocene	Alluvium	Floodplain	Cobbles-Boulders and Pebbly Loam (rounded to subrounded)
Hfp2	Holocene Floodplain Alluvium (1.0 to 2.0 m surface)	Holocene	Alluvium	Floodplain	Cobbles-Boulders and Pebbly Loam (rounded to subrounded)
Hfp2A	Holocene Floodplain Alluvium (1.0 to 2.0 m surface)	Holocene	Alluvium	Floodplain	Sandy Loam
Hfp2B	Holocene Floodplain Alluvium (1.0 to 2.0 m surface)	Holocene	Alluvium	Floodplain	Clayey Loam
Qt1	Quaternary Terrace Alluvium (2.0 m surface)	Quaternary (Undiff.)	Alluvium	Terrace (Floodplain?)	Cobbles-Boulders and Pebbly Loam (rounded to subrounded)
Qt2	Quaternary Terrace Alluvium (2.0 to 4.0 m surface)	Quaternary (Undiff.)	Alluvium	Terrace	Cobbles-Boulders and Pebbly Loam (rounded to subrounded)
Qt3	Quaternary Terrace Alluvium (4.0 to 6.0 m surface)	Quaternary (Undiff.)	Alluvium	Terrace	Cobbles-Boulders and Pebbly Loam (rounded to subrounded)
Qt4	Quaternary Terrace Alluvium (6.0 to 8.0 m surface)	Quaternary (Undiff.)	Alluvium	Terrace	Cobbles-Boulders and Pebbly Loam (rounded to subrounded)
Hf	Holocene (Historic) Fan Deposits (undissected)	Holocene	Alluvium - Debris Flow(?)	Fan	Cobbles and Boulders, Gravel Diamicton
Qf	Quaternary Fan Deposits (undissected)	Quaternary (Undiff.)	Alluvium - Debris Flow(?)	Fan	Cobble- to Boulder-Diamicton with Silty Loam Matrix (subangular to rounded)
Qf1	Quaternary Fan-Terrace Deposits (2.0 to 4.0 m surface)	Quaternary (Undiff.)	Alluvium - Debris Flow(?)	Fan	Cobble- to Boulder-Diamicton with Silty Loam Matrix (subangular to rounded)
Qf2	Quaternary Fan-Terrace Deposits (4.0 to 6.0 m surface)	Quaternary (Undiff.)	Alluvium - Debris Flow(?)	Fan	Cobble- to Boulder-Diamicton with Silty Loam Matrix (subangular to rounded)
Qf3	Quaternary Fan-Terrace Deposits (6.0 to 8.0 m surface)	Quaternary (Undiff.)	Alluvium - Debris Flow(?)	Fan	Cobble- to Boulder-Diamicton with Silty Loam Matrix (subangular to rounded)
Qf4	Quaternary Fan-Terrace Deposits (8.0 to 10.0 m surface)	Quaternary (Undiff.)	Alluvium - Debris Flow(?)	Fan	Cobble- to Boulder-Diamicton with Silty Loam Matrix (subangular to rounded)
Qf5	Quaternary Fan-Terrace Deposits (>10.0 m surface)	Quaternary (Undiff.)	Alluvium - Debris Flow(?)	Fan	Cobble- to Boulder-Diamicton with Silty Loam Matrix (subangular to rounded)
Qap	Quaternary Apron Deposits	Quaternary (Undiff.)	Colluvium	Apron	Cobble- to Boulder-Diamicton with Silty Loam Matrix

Table 2
Summary of morphometric data for at the Fernow, North Fork, and Little River areas

Morphometric Parameter[a]	**Fernow**	**North Fork**	**Little River**
Drainage Order (Strahler)	5	5	6
Basin Area (km^2)	17.62	49.27	41.48
Drainage Density (km^{-1}) (C_t/A)	4.24	3.26	4.66
Maximum Basin Elevation (m)	1099.7	1386	1340.2
Minimum Basin Elevation (m)	513.6	853	512
Basin Relief (km)	0.586	0.533	0.828
Ruggedness (D_dH_b)	2.486	1.737	3.856
Total No. of First-Order Tributaries (Shreve Magnitude)	139	287	380
Total No. of Stream Segment Intersections	158	284	377
Stream Segment Intersection Frequency (km^{-2}) (N_s/A)	8.97	5.76	9.09
Stream Frequency (km^{-2})	12.14	7.79	12.08
Maximum Valley Width (m)	120	180	290
Total Valley Bottom Area (km^2)	0.76	1.86	3.09
Average Hillslope Gradient (dec. deg.)	17.2	18.4	22.1

[a] A = Basin area; N_o = Number of stream segments in a given Strahler order; Hb = Basin relief; Dd = Drainage Density; C_t = Total stream channel lengths; N_t = Total number of stream channels; F = Stream frequency; N_s = Total number of stream segment intersections.

Table 3
Results of debris-fan analysis at the Fernow, North Fork, and Little River areas

Analytical Parameter	Fernow	North Fork	Little River
Total Number of Fans	36	51	116
Total Fan Surface Area	0.113	0.165	0.486
Ratio: Total Fan Area / Total Basin Area	0.006	0.003	0.012
Fan Frequency (km^{-2})	2.0	1.0	2.8

Table 4
Site summary and relative ranking of critical geomorphic factors

Geomorphic Factor	Study Area		
	Fernow	North Fork	Little River
Bedrock: Sandstone Content*	Medium	Medium	High
Physiographic Setting	Plateau	Plateau	Valley and Ridge
Relief	Medium	Low	High
Drainage Density	Medium	Low	High
Slope	Medium	Medium	High
Bedrock-Channel Reaches	Medium	Medium	Low
Valley-Width Expansion Rate	Low	Medium	High
Erosional Accommodation	Low	Medium	High
Debris Flow Occurrence	Low	Low	High
Debris Fan Occurrence	Low	Low	High
Unit Valley-Bottom Storage	Low-Medium	Low-Medium	High

* From Taylor (1997; 1999)

Figure Captions

Fig. 1. Location map showing outcrop belt of the Acadian clastic wedge and physiographic provinces of the central Appalachians. Study locations include: F = Fernow Experimental Forest; NF = North Fork basin, LR = Little River basin (after McClung, 1983; Kulander and Dean, 1986).

Fig. 2. Generalized bedrock geology of the Fernow, North Fork, and Little river study areas. Map unit labels are as follows: Dfk = Foreknobs Formation, Mg = Greenbrier Group, Mmc = Mauch Chunk Group, and |Ppv = Pottsville Group.

Fig. 3. Portion of the surficial geology map for the Little River area, Augusta County, Virginia. Features were originally mapped at a scale of 1:9,600 (Taylor and Kite, 1998). Refer to Table 1 for an expanded explanation of map units. The contour interval is 40 ft.

Fig. 4. Bar graph showing areal distribution of surfical map units at each study area. Surficial map unit codes are as follows: Qr = residuum (ridge tops), Qc1 = colluvium (side slopes, noses), Qc2 = colluvium (hollows), Hch = channels (alluvium/bedrock), Hfp = floodplain alluvium, Qt = terrace alluvium, and Qf = debris fan deposits. T.V.B. refers to total valley bottom area and is equal to the sum of Hch, Hfp, Qt, and Qf. Map units represent the primary sediment-storage compartments recognized in this study.

Fig. 5. Bedrock exposure along primary stream channels at the Fernow, North Fork, and Little River study sites. Drilling records at Little River are from the Headwaters Conservation District, Verona, Virginia (unpublished data, 1963).

Fig. 6. A. Photo showing mixed alluvial-bedrock channel reach at the North Fork study site. Note the bedrock exposure in the channel bottom and perennial discharge. This view is also representative of the typical channel morphology at the Fernow area. Horizontal field of view is 6 meters. B. Photo showing an alluvial-dominated channel reach at the Little River study area. Note the abundance of bouldery gravel, high volume of coarse woody debris, and lack of bedrock exposure. Horizontal field of view is 17 meters.

Fig. 7. Frequency distribution of debris-fan occurrence at a given stream tributary junction type (Strahler order) for the study areas. The junction type code refers to Strahler stream order intersections (e.g. 1-4 = 1^{st} order – 4^{th} order intersection). Zero-order tributaries are hollows (*sensu* Hack and Goodlett, 1960) without well-defined channels.

Fig. 8. Generalized cross-sections showing valley-bottom surficial units and the methodology employed to determine sediment storage volumes. Surface heights are given in parentheses; t is the estimated thickness relative to active channel grade. Where surface heights are bracketed by ranges, the intermediate value of the range is used as thickness in volume calculations (unit volume = polygon area x thickness). "Surface volume" equals the sum of surficial unit volumes above channel grade; "fill volume" equals the volume of sediment between active channel grade and the lower bedrock interface. The fill volume at the Little River was estimated by comparison of two 30-m grid surfaces, at channel grade and bedrock interface, respectively. The volume algorithm of Surfer for Windows (Golden Software, 1997) was utilized in the latter analysis. Refer to Table 1 for explanation of map unit abbreviations.

Fig. 9. Estimated valley-bottom storage volumes for the Fernow, North Fork, and Little River study areas. The methodology for calculating "surface volume" and "fill volume" is illustrated in Figure 9. Total valley-bottom storage is equal to the sum of "surf vol" and "fill vol". Unit storage represents total storage volume per unit watershed area.

Fig. 10. Linear regression summary from scatter plot of longest channel distance from divide (m) vs. valley width (m). Note the constricted valley width trend at the Fernow area and the relatively high rate of valley-width expansion at Little River.

Fig. 11. Longitudinal profiles illustrating transport efficiency models for the Fernow, North Fork, and Little River study areas.

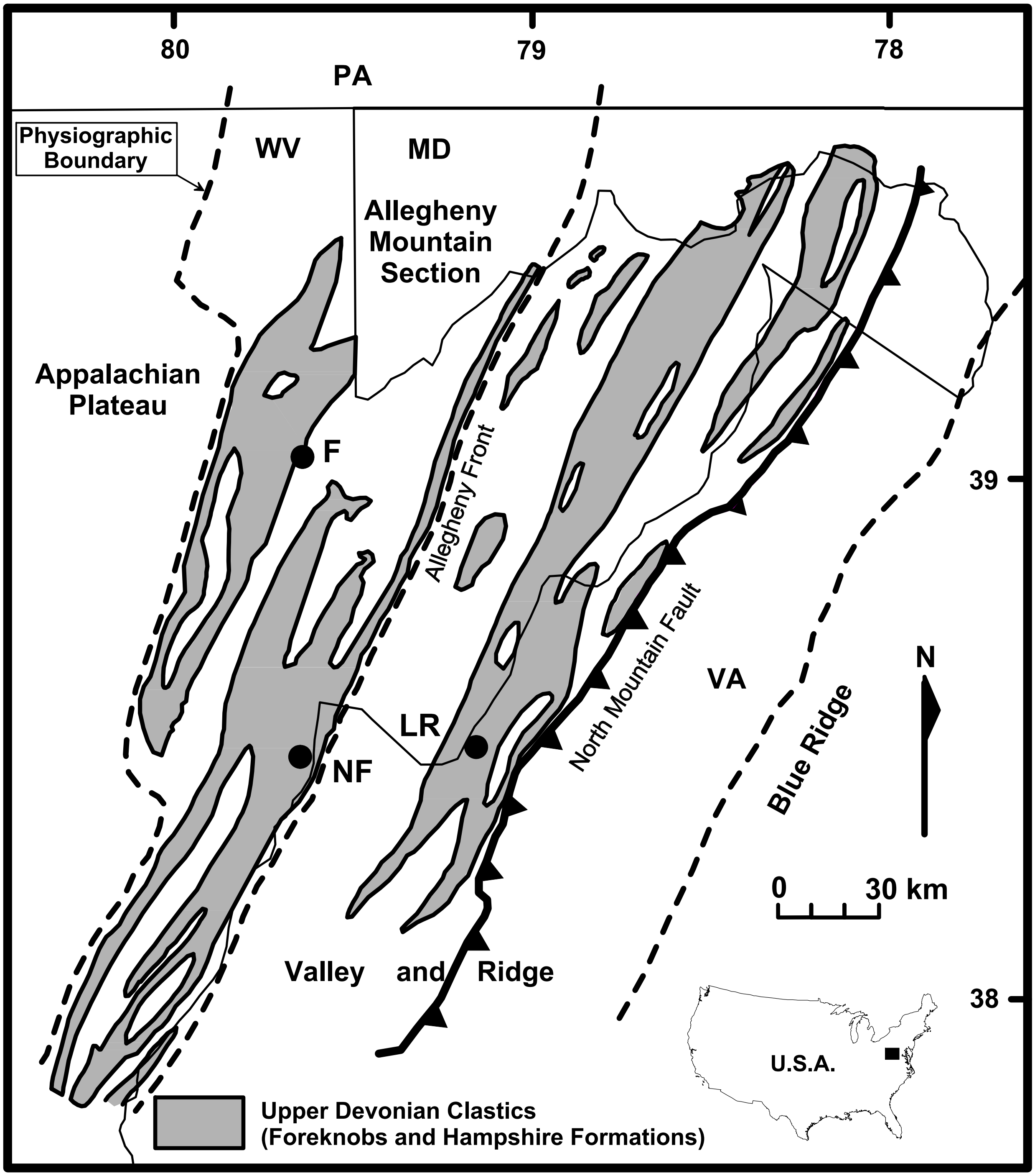
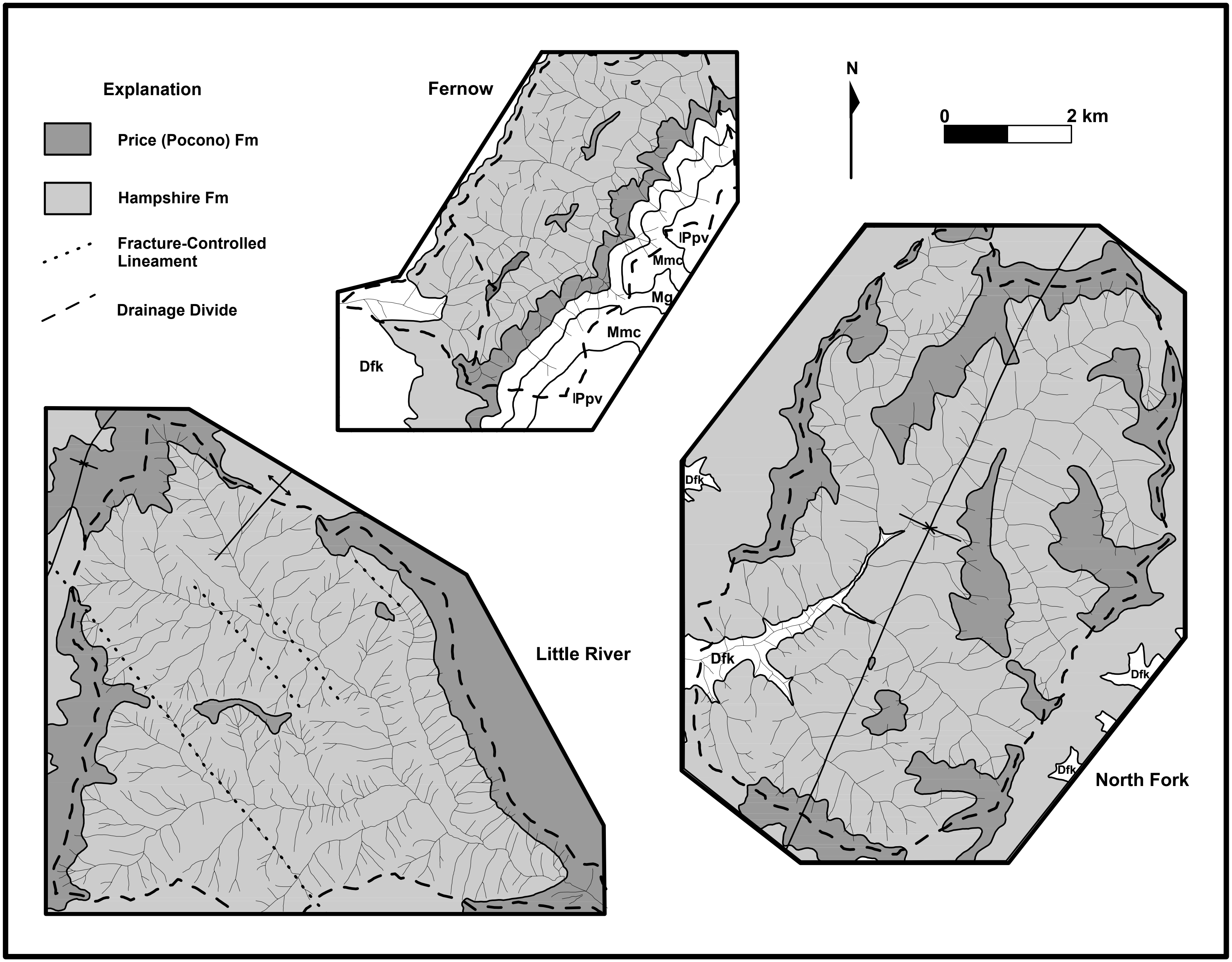


Figure 1 (Taylor and Kite)



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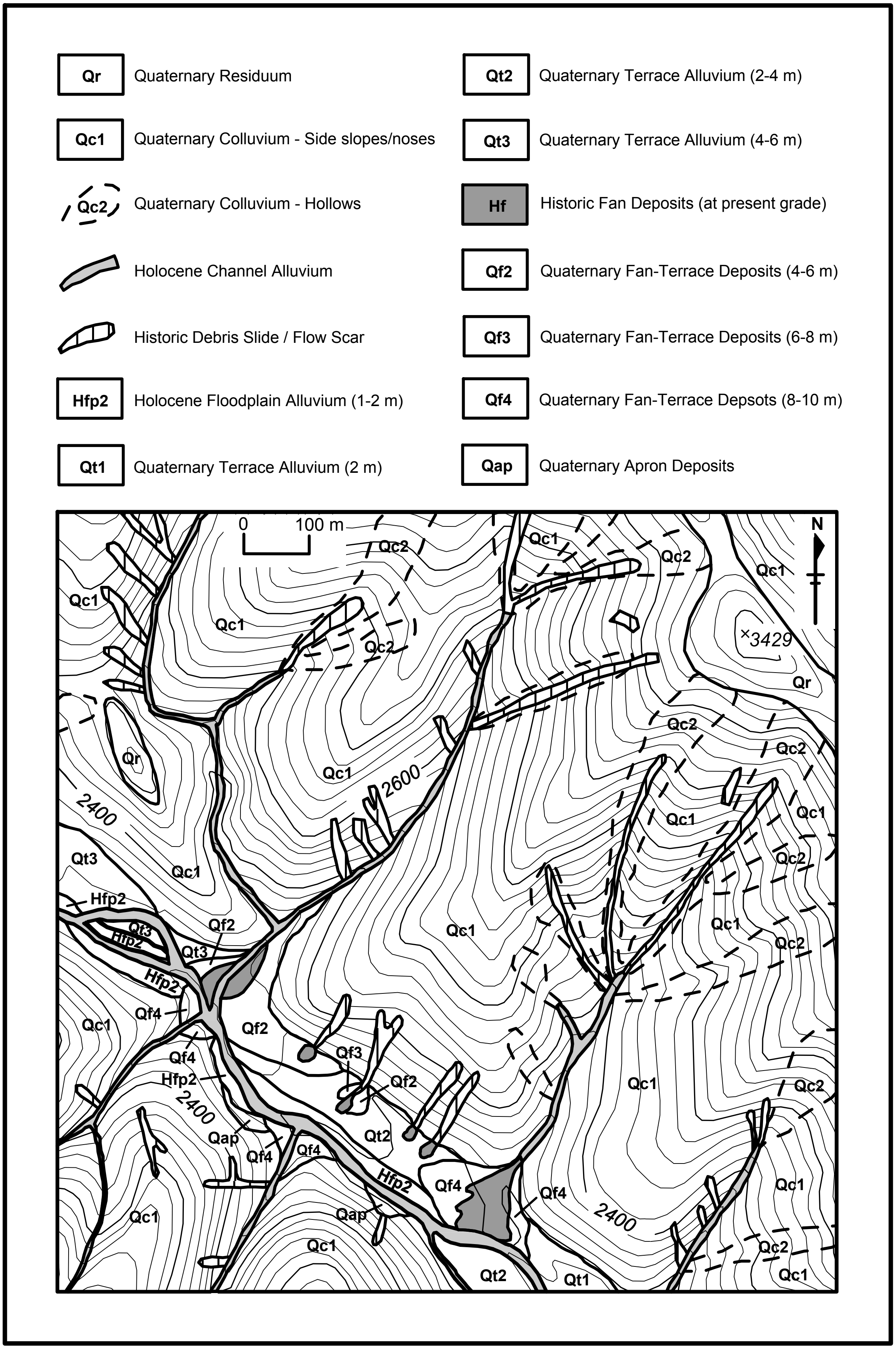

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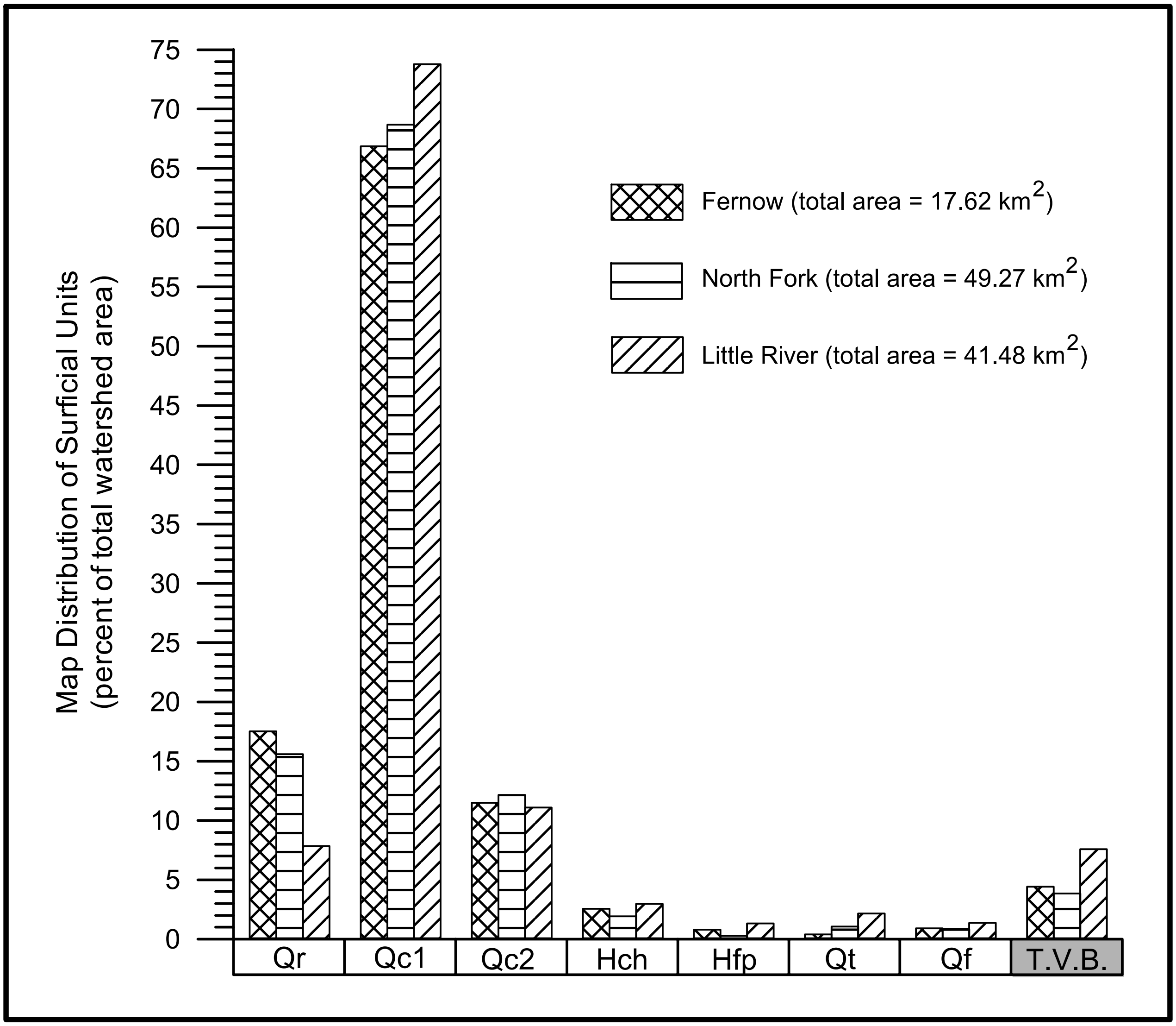

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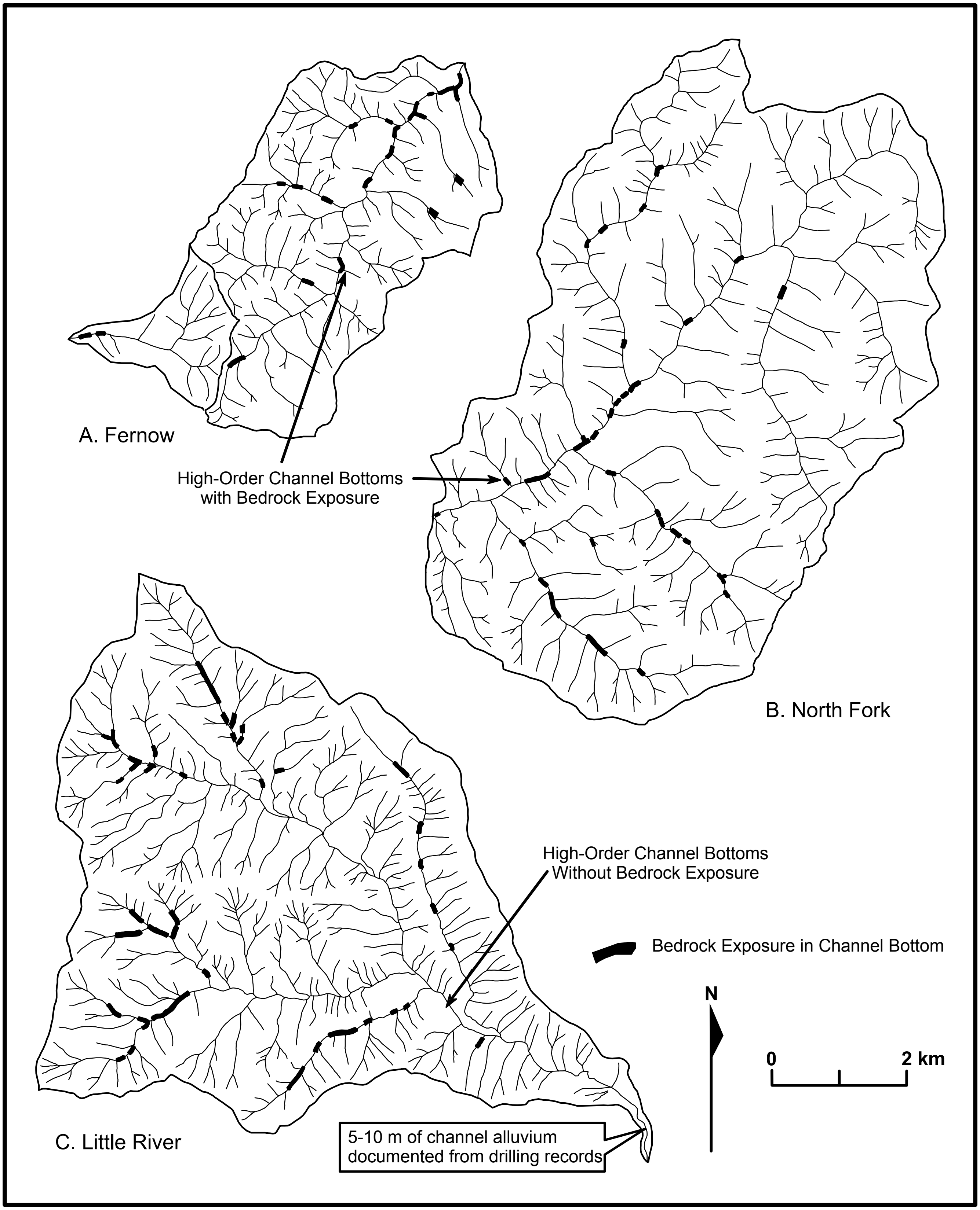
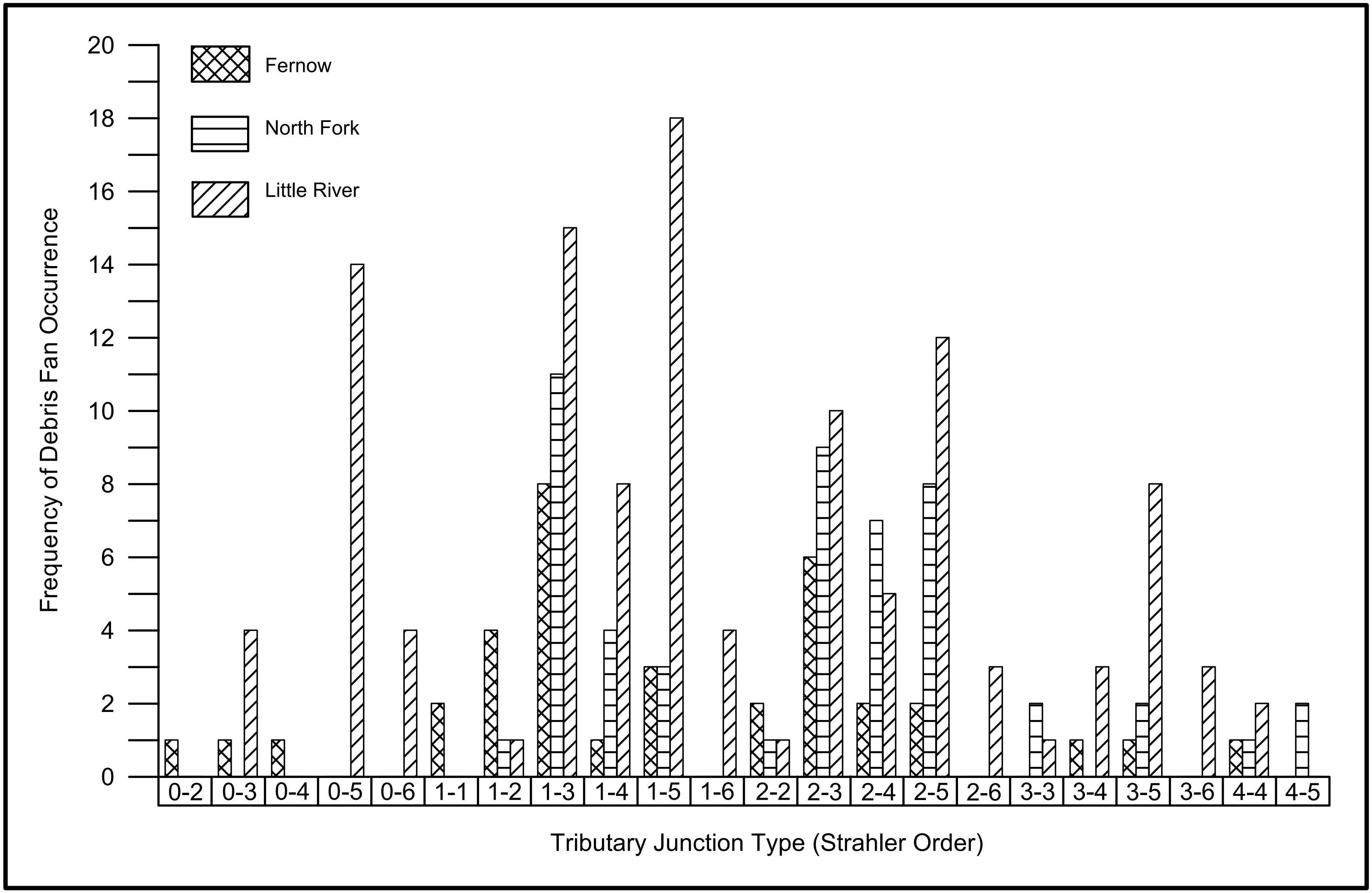


Figure 5 (Taylor and Kite)

Figure 6 (Taylor and Kite)



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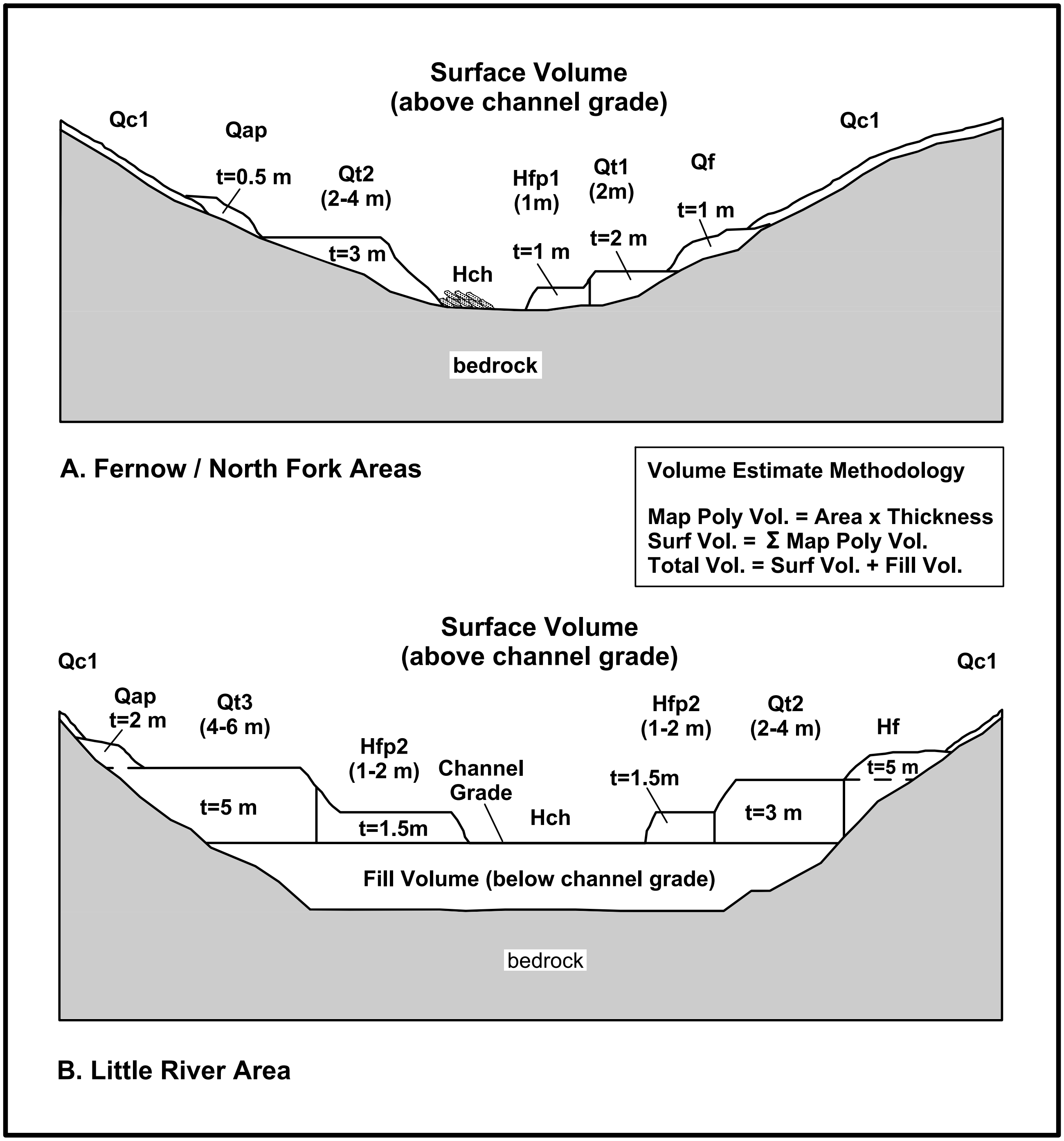
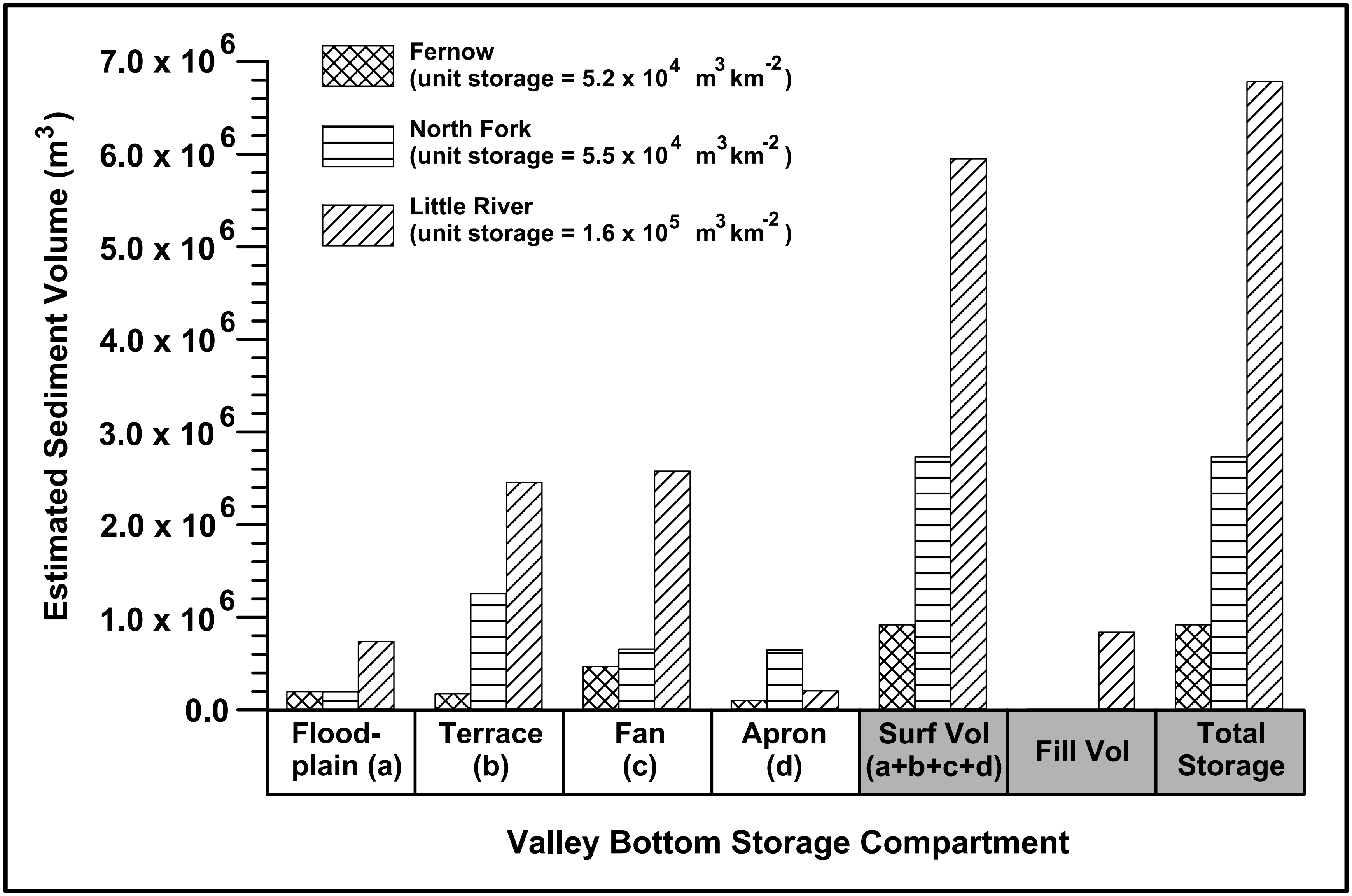


Figure 8 (Taylor and Kite)



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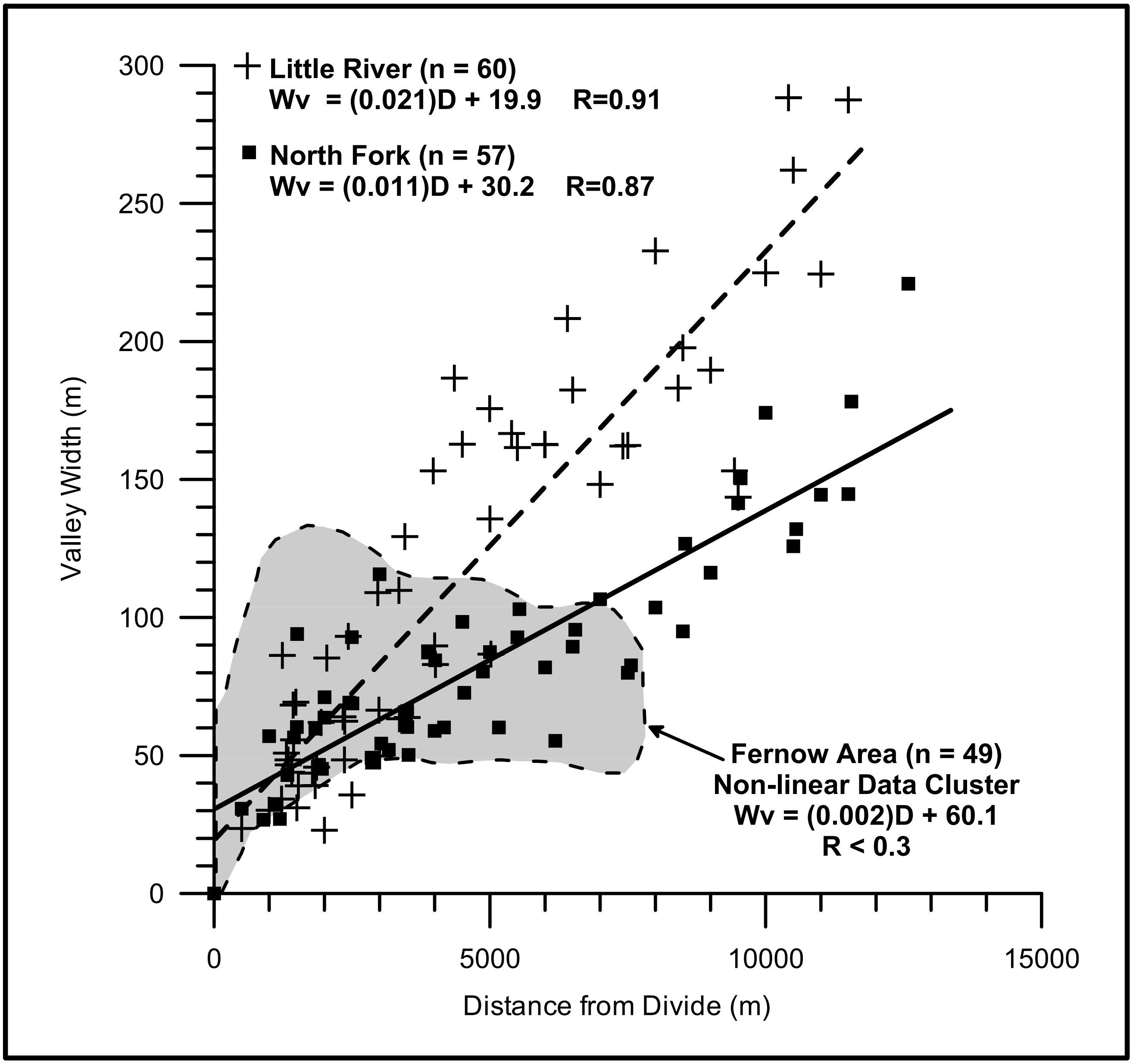

Figure 10 (Taylor and Kite)

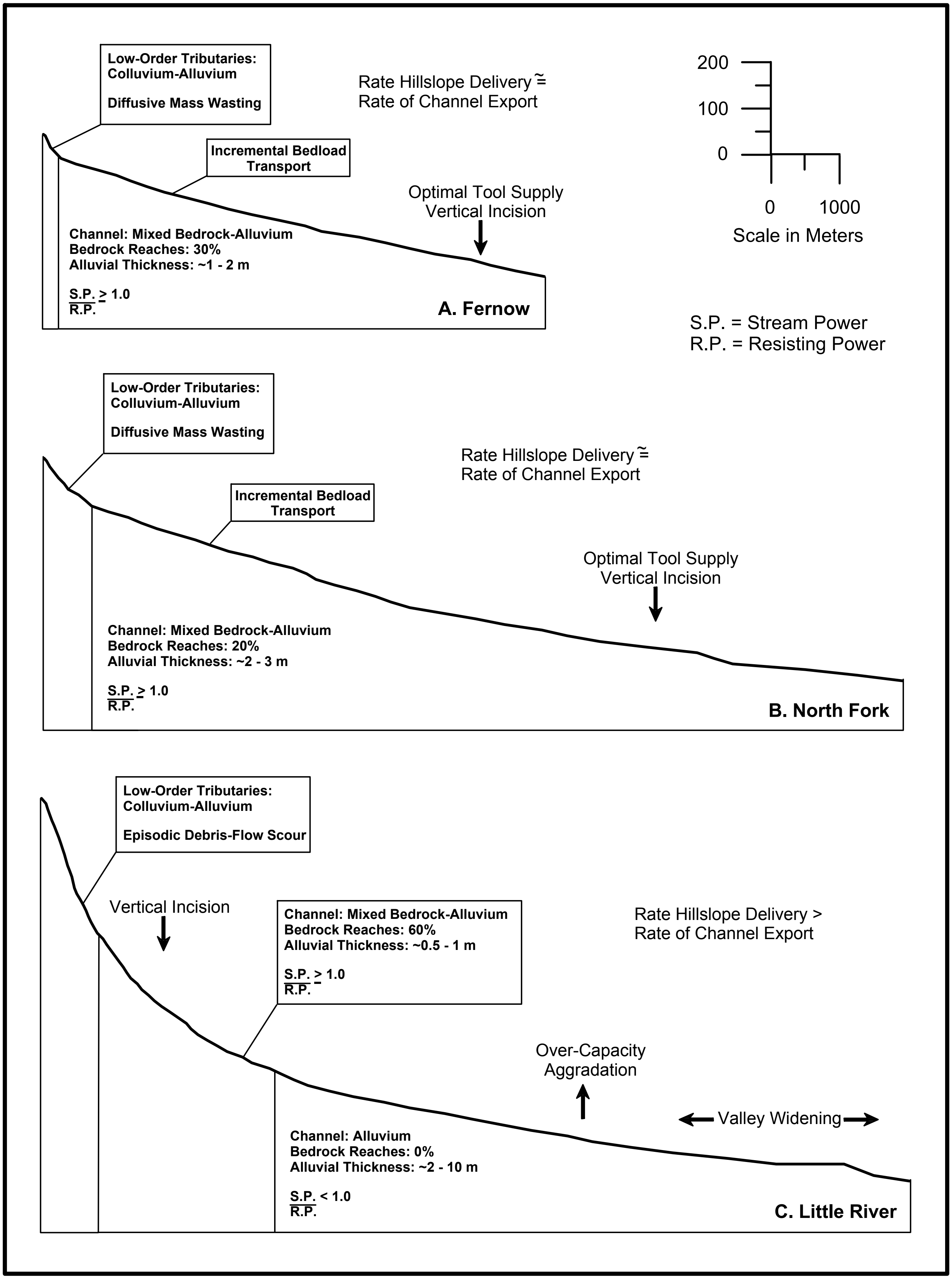

Figure 11 (Taylor and Kite)