University of Washington
Department of Civil and Environmental Engineering

# THE FREQUENCY AND EXTENT OF HYDROLOGIC DISTURBANCES IN STREAMS IN THE PUGET LOWLAND, WASHINGTON

Christopher P. Konrad

Water Resources Series
Technical Report No.164
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The frequency and extent of hydrologic disturbances in streams in the Puget Lowland, Washington

Christopher P. Konrad

A dissertation submitted in partial fulfillment of the requirements for the degree of

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Graduate School

This is to certify that we have examined this copy of a doctoral dissertation by

Christopher P. Konrad

and have found that it is complete and satisfactory in all respects, and that any and all revisions required by the final examining committee have been made.

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Stephen J. Burges

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Abstract

The frequency and extent of hydrologic disturbances in streams in the Puget Lowland, Washington

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Co-chairs of the Supervisory Committee:
Research Associate Professor Derek B. Booth
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Hydrologic changes resulting from urban development degrade the biological conditions of stream ecosystems by modifying annual and inter-annual stream flow patterns. In urban streams, discharge is less than mean annual discharge on more days of the year, discharge exceeds the magnitude of frequent floods for a shorter duration of time, and the peak discharge rate of the annual maximum flood is less variable than in suburban streams. These hydrologic changes may cause increases in the frequency and extent of disturbances in urban streams.

Floods and drought are common forms of disturbance in stream ecosystems. The biological effects of these hydrologic disturbances depend on their spatial extent and frequency. The extent of seasonal drought was documented in 59 Puget Lowland stream basins. Streams draining 1.2 $km^2$ had a 50% probability of being dry during summer base flow conditions (ephemeral). The length (km) of perennial streams in a basin varied as a linear function of drainage area ($km^2$), $L = 0.4 A + 0.8$ with a root mean square error

of 0.04 km.  While urban development did not influence the extent of perennial streams in the basins, it may reduce the period of continuous flow during winter and spring in ephemeral streams.

The spatial extent of bed material entrainment during floods was documented at seven gravel bars in three Puget Lowland streams using bed tags, which are metal washers placed in the stream bed.  Partial entrainment ($PE_{bar}$), which is the fraction of a bar's surface disturbed in a flood, was estimated by $PE_{bar} = 12.5(\tau_0* - 0.045)$ with a root means square error of 0.099, where $\tau_0*$ is the total boundary shear stress applied by the flood divided by the product of the median of the particle-size distribution of the surface material on the gravel bar and its buoyant specific weight.  Frequent and extensive flood disturbance is likely in urban and other gravel-bed streams where the magnitudes of floods are greater than the magnitude of longer-duration intermediate flows, represented by the discharge exceeded 5% of the time, that control the strength of the stream bed.

# Chapter 1: Hydrologic disturbances in Puget Lowland streams

Urban development in the Puget Lowland, Washington modifies hydrologic processes that regulate the transport of water through hillslopes to streams. I examine whether the resulting stream flow patterns influence the frequency and spatial extent of two types of physical disturbances in stream ecosystems: desiccation of streams during the summer dry season and entrainment of the surface material of the stream bed during winter floods.

Physical disturbances such as floods, drought, windstorms, fire, and landslides change the biologic conditions of a place (Pickett and White, 1985). Organisms must flee or are removed from their habitats; new substrates are introduced for habitation; and the changes in environmental conditions resulting from the disturbance initiate growth, reproduction, and migration of organisms. While biological conditions in streams typically recover rapidly after a disturbance (Stehr and Branson, 1938; Fisher et al., 1982; Bayley and Osborne, 1993), spatial and temporal patterns of physical disturbances may have persistent influences on the biological conditions of ecosystems including the types and health of organisms, the size of their populations, and the trophic structure of communities (Allan, 1996). Differences in the frequency and extent of disturbances between stream ecosystems are therefore likely to manifest as differences in their biological conditions.

The influence of stream flow patterns on disturbance patterns serves as a possible nexus between urban development and degradation of stream ecosystems in the Puget Lowland region. If the hydrologic changes resulting from urban development increase the frequency and extent of disturbance in streams, there are likely to be biological effects that will persist as along as stream flow patterns are altered. Total restoration of pre-development stream flow patterns is an infeasible objective for urban streams in the Puget Lowland since hydrologic restoration would require extensive changes in the extent and style of human occupation in urban areas. There may be opportunities, however, to focus management actions on those hydrologic changes that are the most ecologically deleterious.

Since physical disturbances may have a dominant influence on the biological conditions of ecosystems, I consider whether disturbance patterns in streams are controlled by stream flow patterns, which, in turn, are modified as a result of urban development. If specific stream flow patterns control disturbance patterns and are influenced by urban development, then restoration of these stream flow patterns may be a necessary component for recovering the biologic conditions of stream ecosystems in urban areas. Furthermore, restoration of these stream flow patterns would provide ecologically relevant hydrologic objectives for stormwater management projects.

Given the resiliency of lotic communities to hydrologic fluctuations (McElarvy et al., 1989; Boulton et al., 1992) as well as their rapid recovery after individual disturbances, I presume that stream flow patterns have their greatest biological influence over time-scales longer than a storm or season. In Chapter 2, I analyze how the hydrologic effects of urban development, at the scale of single storms and seasons (Leopold, 1968), manifest as stream flow patterns over annual and inter-annual time scales. Hydrologic effects at these longer time scales are evident as changes over time in streams flow patterns during periods of urban development and as differences in stream flow pattern between streams with different levels of urban development.

Hydrologic changes over time in a stream and differences between streams may influenced by variation in natural physiographic factors (e.g., rainfall patterns, surficial geology, drainage area). The influences of natural physiographic factors on stream flow patterns are compared to the influences of urban development to assess whether the hydrologic effects of urban development can be distinguished from the effects of physiographic variation.

The variation in stream flow patterns between streams resulting from natural physiographic factors may account for the differences in the structure and composition of lotic communities (Odum, 1956; Poff and Ward, 1989; Poff and Allan, 1995). As a result, biological conditions in streams may vary with stream flow patterns rather than land use per se and, in particular, streams with low levels of development may have degraded biological conditions if their stream flow patterns are characteristic of higher levels of development. The influences of annual and inter-annual stream flow patterns are compared to the biological conditions of streams using the Benthic Index of Biological Integrity (Karr and Chu, 1999).

After identifying differences in urban and suburban stream flow patterns, I analyze patterns of hydrologic disturbances in Puget Lowland streams. Hydrologic disturbances occur when either a drought or a flood cause biological changes in a stream. During low flow periods, portions of a stream channel may be dry and unsuitable for aquatic organisms. As a result, these organisms must flee downstream or into the hyporheic zone. Fish stranded in pools and sessile organisms (e.g., periphyton) may perish during droughts. During high flow periods, stream flow may entrain portions of the stream bed surface washing organic debris, periphyton, and other benthic organisms downstream. Fish and invertebrates may seek refuge in areas with slower currents or in the hyporheic zone (Giller and Malmqvist, 1998). While the biological effects of a flood or a drought may be short-lived (Stehr and Branson, 1938), the patterns of disturbance over time may have persistent effects on lotic communities.

Two spatially extensive measures of low-flow disturbance patterns are introduced in Chapter 3: the drainage area of perennial first-order streams and the length of perennial streams in a basin. The measures are used to test the low-flow disturbance hypotheses that (1) the drainage area required to generate perennial flow is greater in urban areas than in suburban areas, and (2) the length of perennial streams is shorter in urban basins than in suburban basins. These effects are anticipated to result from a reduction in subsurface flow and ground water elevations in urban areas.

A method for estimating the spatial extent of bed disturbance during floods in gravel-bed stream is developed in Chapter 4. Experiments at seven gravel bars using “bed tags”, which are metal washers inserted into the stream bed, were used to document spatial and temporal patterns of stream bed disturbance. The fraction of the bar’s surface entrained during a flood, or “partial entrainment”, is related to a dimensionless shear stress which is the ratio of a flood's peak applied shear stress to the median of the particle-size distribution of the surface material on the bar.

The general relationship between dimensionless shear stress and partial entrainment allows the extent of bed disturbance during a flood to be estimated at a site on a gravel-bed stream. It is used in Chapter 5 to test the hypothesis that stream bed disturbance during floods is under hydrologic control. Stream bed disturbance is expected to be more frequent and extensive in urban streams than in suburban streams as a result of the hydrologic effects of urban development.

# Chapter 2: Stream flow patterns in the Puget Lowland, Washington

Urban development in the Puget Lowland, Washington modifies hillslope hydrologic processes, including the production of runoff, when trees and soils are cleared, the land surface is graded, and roads and drainage systems are constructed. The changes in hillslope hydrologic processes caused by urban development produce characteristic stream flow patterns at a storm-scale such as increased peak discharge rate for a given amount of rainfall (Leopold, 1968). The effects of storm-scale hydrologic changes on the biological conditions of urban streams are difficult to deduce since the biological effects of single storms are transient (Fisher et al., 1982). Over annual or multiple-year time scales, however, changes in stream flow patterns may have a persistent influence on the biological conditions of urban streams.

The basic premise of this analysis is that stream flow patterns have a strong influence on the biological conditions of streams and, as a corollary, anthropogenic changes in stream flow patterns degrade lotic ecosystems in urban areas of the Puget Lowland. Stream flow patterns are presumed to have the greatest influence on the biological conditions of streams over annual and inter-annual time scales. Three hydrologic measures are presented here that represent the storm and base flow patterns over these longer time scales: (1) the fraction of a year that the daily mean discharge rate exceeds the annual mean discharge rate ($T_{Qmean}$); (2) the fraction of a multiple year period that the discharge rate of a specified flood quantile is exceeded ($T_{X\ yr}$ is the cumulative duration that stream flow exceeds the discharge of a flood occurring on average 1/X times per year); and (3) the coefficient of variation of the annual maximum flood ($CV_{AMF}$). These measures represent stream flow patterns occurring over progressively shorter periods of time from $Q_{mean}$, which is typical of winter base flow in Puget Lowland streams, to annual maximum flood. They are analyzed in light of natural physiographic variability and urban development using discharge records from stream gages in the Puget Lowland region.

The relationship between stream flow patterns and biological conditions of streams is evaluated by comparing the three hydrologic measures to the benthic index of biological integrity (B-IBI) developed for Puget Lowland streams (Kleindl, 1995; Karr and Chu, 1999; Morley, 2000). Biological conditions in streams are expected to vary with stream flow patterns rather than strictly with land use. For example, streams may have very different biologic conditions, in spite of similar levels of urban development, if they have different stream flow patterns. Conversely, biological conditions may be similar in streams with different levels of urban development if the streams have similar flow patterns.

## 2.1. A review of the hydrologic effects of urban development

Human activities in a landscape modify a wide range of hydrologic processes, including overland and shallow subsurface flow, evapotranspiration, groundwater recharge, and stream flow (Hoover, 1944; Savini and Kammerer, 1961; Sawyer, 1963; Harris and Rantz, 1964). This section describes changes in hillslope hydrologic processes caused by urban development and the resulting stream flow patterns.

### 2.1.1. Urban effects on hillslope hydrologic processes

Construction of roads and buildings change hydrologic processes of forested hillslopes, beginning when vegetation and soil are removed from the land surface. Clearing forests and grading the land surface reduces: the depth of the soil column, depression storage (including that provided by wetlands), forest canopy interception, evaporation, and transpiration. Roofs, roads, and lawns with shallow soils generate more runoff during storms and deliver the runoff more quickly to constructed drainage networks than would forested hillslopes. Furthermore, the drainage networks in urban areas (road gutters, storm sewers, culverts) route water more rapidly downstream to receiving water bodies such as streams than would the wetlands in forests. As a result of these changes, urban hillslopes produce more storm flow as a fraction of rainfall and at greater peak rates than forested hillslopes (Burges et al., 1998). Water stored on hillslopes is also depleted more rapidly in urban catchments, leading to rapid recession of storm flow after rain ceases.

Burges et al. (1998) examined hydrologic differences between two zero-order catchments in the Puget Lowland: Klahanie, a 0.17-$km^2$ residential catchment with approximately 30% of the drainage area covered by impervious surface and much of the rest covered by lawn; and Novelty Hill, a 0.37-$km^2$ catchment covered by a second-growth Douglas fir – western hemlock forest. They found that runoff from Klahanie constituted 44 to 48% of annual precipitation whereas runoff from Novelty Hill was only 12 to 30% of annual precipitation. Peak rates of runoff were higher from Klahanie than those from Novelty Hill, with the greatest differences in storm flow production occurring under relatively dry antecedent conditions (e.g., during summer and early autumn in the Puget Lowland).

Differences in storm flow production reflect differences in the dominant forms of runoff processes. Overland flow was a dominant form of runoff for Klahanie (29% of precipitation) but was relatively minor for Novelty Hill (4% of precipitation), where shallow subsurface flow dominated (Wigmosta and Burges, 1997). Burges et al. (1998) attribute the changes in runoff production processes and the resulting increase in storm flow production primarily to a reduction of soil column depth in the residential area and, secondarily, to the addition of impervious surfaces.

The effects of land use on the allocation of rainfall to different hillslope hydrologic processes (i.e., runoff, evapotranspiration, and recharge) vary with season and by year depending on rain storm and temperature patterns in the Puget Lowland (Fritschen et al., 1977; Dinicola, 1990; Bauer and Mastin, 1997; Burges et al., 1998). Burges et al. (1998) calculated lower total recharge during a wet winter at the Klahanie catchment than at Novelty Hill. In contrast, Bauer and Mastin (1997) suggest that recharge rates in glacial till-mantled forests and pastures of the Puget Lowland depend largely on infiltration capacity of till and are independent of annual precipitation and vegetation. They observed that soil immediately above till is saturated during most of the winter regardless of whether a hillslope is forest or pasture. Their conclusion may not apply in urban areas, however, where hillslope are effectively shorter due to roads and constructed drainage networks, the remaining thin soil column drains quickly between storms, and surface depressions have been graded flat. Recharge is likely to be low during summer in forested and urban areas of the Puget Lowland because of low rainfall and high evapotranspiration. In spite of seasonal and annual variability, however, increased runoff from urban hillslopes can be expected to be balanced by decreased groundwater recharge and evapotranspiration at annual scales (ASCE Task Committee, 1975).

### 2.1.2. Urban effects on stream flow patterns

The changes in hillslope hydrologic processes caused by urban development produce characteristic changes in stream flow patterns. While the hydrologic effects of urban development were incorporated in early efforts to model runoff production in cities (e.g, Horner and Flynt, 1936), it was not until the 1960's that the changes in stream flow patterns resulting from urban development were described generally in terms of increased storm flow volume, peak discharge rate, and recession rate and decreased time to peak discharge (Carter, 1961; Harris and Rantz, 1964). At annual scales, urban stream flow patterns are characterized by more frequent storm peaks, particularly during periods of dry antecedent conditions of hillslopes. Urban effects on stream flow over longer time scales are evident in changes in flood magnitude, with small, frequent floods exhibiting the largest relative increase in magnitude (James, 1965; Hollis, 1975; Bailey et al, 1989).

The effects of urban development on base flow are not as consistent or evident as the effects on storm flow production (ASCE Task Committee, 1975). Base flow can be expected to decrease if urban development reduces ground water recharge and, accordingly, groundwater elevations. Sawyer (1963) attributed a difference in base flow for two adjacent basins on Long Island, New York to lower recharge of shallow, unconfined aquifers as a result of urban development in one of the basins.

The evidence for decreased base flow, however, is equivocal. Base flow during a dry season may be higher as a result of deforestation, particularly where riparian vegetation is cleared (Hoover, 1944; Keppeler and Ziemer, 1990). In urban areas, water imported from other basins and used for landscape irrigation can also increase dry season flow (Harris and Rantz, 1964). Water use in urban areas can also influence stream flow patterns where streams are impounded or ground water is extracted.

Urban stream flow patterns are typically characterized in terms of parameters such as flood magnitude and sediment transport rates. These parameters do not indicate the biological effects of hydrologic changes in urban streams. Alternative hydrologic measures, described in this chapter, characterize stream flow patterns in terms that are ecologically relevant. These measures provide a link between urban development and stream degradation such that they should be robust predictors of stream conditions, where stream flow patterns are important, and can serve in efforts to identify hydrologic mechanisms of stream degradation.

## 2.2. Ecological influences of stream flow patterns

Stream flow is a primary element of stream ecosystems with many of its characteristics (e.g., discharge rate, frequency, duration, seasonal fluctuations) potentially affecting the biological conditions of streams (Shelford and Eddy, 1929; Odum, 1956; Horwitz, 1978; Fisher et al,. 1982; Schlosser, 1985; Newbury, 1988; Power et al., 1988; Resh et al., 1988; Poff and Ward, 1989; Death and Winterbourne, 1995; Richter et al., 1996; Poff et al., 1997). While lotic communities may be insensitive to small fluctuations in stream flow that routinely occur from day to day, hydrologic variability over time-scales longer than a storm or season does appear to have an ecological influence. For example, Poff and Allen (1995) found that streams in Wisconsin and Minnesota with highly variable daily discharge and base flow, over multiple-year periods, had fish assemblages with lower taxonomic diversity and more trophic and habitat generalists than streams with less hydrologic variability.

Hydrologic variability in the extreme, floods and droughts, has an immediate effect on biological conditions of streams including reductions in the taxonomic diversity and population levels of fish and macroinvertebrates, periphyton biomass, and the trophic structure of lotic communities (Stehr and Branson, 1938; Douglas, 1958; Anderson and Lehmkuhl, 1968; Fisher et al., 1982; McAuliffe, 1984; Schlosser, 1985; McCormick and Stevenson, 1991; Boulton et al., 1992, Bayley and Osborne, 1993; Closs and Lake, 1994; Dieterich and Anderson, 1995; Wooton et al., 1996). Biological conditions in streams re-establish quickly, often within months, after hydrologic disturbances (Stehr and Branson, 1938; Fisher et al., 1982; Power and Stewart, 1987; DeBray and Lockwood, 1990; Boulton et al., 1992; Bayley and Osborne, 1993; Jones et al., 1995). The rapid succession of lotic communities limits the period of time during which the biologic conditions of a stream are affected by individual flood or drought events.

A pattern of disturbance can have persistent biological effects, for example, if disturbances recur before biological conditions have recovered from the previous event. In this case, a biological community will remain in an early successional state. Thus, the pattern of disturbances is likely to have a persistent influence on the biologic conditions of streams even if the effects of individual disturbances are transient. Likewise, more frequent (e.g., daily) or lower magnitude (e.g., changes in base flow levels) hydrologic variability that does not disturb a stream community may nonetheless influence its composition, for example, by affecting the area of habitat available to aquatic organisms, the downstream flux of nutrients, or the velocity of flow.

Given the magnitude of hydrologic changes resulting from urban development, urban stream flow patterns are likely to affect the biological conditions of streams. Orser and Shure (1972) documented lower population densities of dusky salamander (*Desmognathus fuscus fuscus*) with increasing levels of urban development for five streams near Atlanta, Georgia. They indicated that floods had likely scoured individuals downstream, reducing salamander populations and creating unstable age structures.

The influence of stream flow patterns, however, is mediated by non-hydrologic factors such as habitat diversity (Gorman and Karr, 1978; Gurtz and Wallace, 1984) and biotic interactions (McAuliffe, 1984; Feminella and Resh, 1990; McCormick and Stevenson, 1991; Wootton et al., 1996). As a result, the biological consequences of specific stream flow patterns are not fixed but depend on the ecological context of those patterns. In any event, stream flow patterns particularly over annual and multiple-year periods may have general effects on lotic communities.

## 2.3. Analysis of Puget Lowland stream flow patterns

The objective of this analysis is to identify stream flow patterns that may account for the degradation of the urban stream ecosystems. Such stream flow patterns should change during periods of urban development, indicating the potential for a biological change over time, though no historical data on the biological conditions of Puget Lowland streams are presented here. Moreover, biological conditions in streams should vary with these stream flow patterns regardless of the level of urban development in their basins.

Natural (i.e., non-anthropogenic) hydrologic variability in space and time presents a formidable obstacle to this analysis because hydrologic changes over time and differences between basins may not be the results of urban development. Furthermore, some types of natural hydrologic variability (e.g., differences in rainfall patterns from year-to-year) are not expected to have significant or persistent biological effects. As a result, the analysis attempts to distinguish stream flow patterns modified by urban development from natural hydrologic variability.

Rainfall patterns, drainage area, geologic materials, soils, topography, hypsometry, basin shape, and other physiographic factors govern runoff production from a stream basin. Stream flow patterns will vary over time in a stream or between two streams as a consequence of changes or differences in natural physiographic factors. Physiographic factors can be controlled in experiments focused on small, contiguous stream basins, such that comparisons between streams illustrate the hydrologic effects of land use (e.g., Hoover, 1944). However, neither changes in stream flow patterns over time at a stream nor differences between stream flow patterns for stream basins covering large, heterogeneous areas can be attributed to differences or changes in land use without considering the influence of natural physiographic factors.

The discharge of a stream depends on its drainage area with larger basins producing more discharge. One approach for comparing stream flow patterns between streams of different sizes is to normalize discharge rates, or volumes, by drainage area. While normalized discharge rates are presented here, discharge rates, volumes, and flood peak distributions are not simple linear functions of drainage area (Miller et al., 1971; Dunne and Leopold, 1978; Pilgrim et al., 1982; Smith, 1992). As a result, differences in stream flow patterns can be expected when comparing streams of different size. Moreover, area-based normalization schemes cannot account for many other physiographic factors influencing stream flow.

While physiographic variability may be an obvious issue for analyzing differences between streams, it is just as vexing when analyzing hydrologic changes over time. Comparisons of stream flow patterns “before” and “after” a land use change will reflect differences in weather patterns that are not likely to be an effect of land use changes. Variability in conditions (e.g., antecedent soil moisture and depression storage, net rainfall rates and depths), at temporal scales ranging from storms to a year, influence stream flow observed at any point in time. As a consequence, differences in stream flow patterns over time may not be attributable to differences in land use.

The biological conditions of streams should vary with stream flow patterns rather than the level of urban development in stream basins per se to the extent that stream flow has a primary influence on stream ecosystems. Changes in ecologically relevant flow patterns should be evident during periods of urban development if hydrologic modification is a primary cause of degradation in urban stream ecosystems. Furthermore, natural variation in space and time in these flow patterns should be less than the variation produced by urban development. The three hydrologic statistics of annual and multiple-year stream flow patterns are analyzed in light of these criteria.

### 2.3.1. Puget Lowland streams

The hydrologic effects of urban development in the Puget Lowland are characterized in terms of differences in flow patterns between urban and suburban streams, and changes in flow patterns in urban and suburban streams, during the latter half of the 20th century. The locations of the streams are shown in Figure 2.1. Differences between urban and suburban stream flow patterns were analyzed for the period of record from Water Years (WY) 1989 to 1998 (i.e., 1 October 1988 to 30 September 1998). Changes over time in stream flow patterns were analyzed by comparing hydrologic measures for the period from WY 1960 through WY 1969 to the period from WY 1989 through WY 1998. Table 2.1 lists the drainage area, road density, and the agency operating each stream gage.

The stream basins span the range of urban development found in the Puget Lowland as indicated by road densities (i.e., the total length of road in a stream basin divided by its drainage area). Road densities were determined using a geographic information system by delineating watersheds from USGS 7.5 minute topographic maps, with elevation contour intervals of 5 m or 20 ft, and calculating the total road length within each watershed using vector representations of roads. The road data were provided by King, Snohomish, Pierce, and Kitsap counties and include interstate and state highways and county roads ca. 1990. The road data do not include logging and service roads or private driveways.

In some cases, I have classified as “urban” or “suburban” to facilitate an analysis or simplify its results. “Urban” streams are defined here arbitrarily as having road densities > 6 km/km$^2$; “suburban” streams are defined as having road densities < 6 km/km$^2$. Within each category, however, stream basins span a gradient of urban development. Flett, Juanita, Leach, Mercer, and Miller Creeks have the highest road densities (>9.0 km/km$^2$). Forest cover in these basins is limited to steep slopes and narrow riparian corridors. Clover, Des Moines, Hylebos, Swamp, North, and the Cedar River tributary 0308 have lower road densities, ranging from 7.4 to 7.9 km/km$^2$, but are classified as urban. Suburban stream basins with moderate road densities (4.0 to 5.0 km/km$^2$) include May, Big Bear, Covington, Jenkins, and Soos Creeks, which have few commercial developments and low density residential developments along with pastures and forests. Suburban streams with the lowest road densities (0 to 4.0 km/km$^2$), include Big Beef, Evans, Huge, Issaquah, Newaukum, Novelty Hill and Rock Creeks, which all have extensive forests and pastures in their basins.

### 2.3.2. Differences in storm flow between streams due to physiographic factors

Basin area, geology, and topography are dominant factors influencing storm flow production in Puget Lowland streams. The influenceof these physiographic factors are illustrated by the stream flow responses in various streams to a large storm from 3 to 5 April 1991. The peak discharge rate in many streams in the region was approximately equal to the mean annual flood. Recorded three-day rain depths were 100 mm at Novelty Hill (Burges et al. 1998) and 120 mm at the Seattle-Tacoma International airport, with 60 to 70 mm falling on 4 April 1991. Hydrographs of area-normalized daily discharge for the period from 1 to 15 April 1991 are shown in Figure 2.2 for 18 Puget Lowland streams. The influence of basin area is illustrated by comparing stream flow patterns for three "nested" catchments: Novelty Hill (drainage area 0.37 km$^2$), Evans Creek, (drainage area of 37 km$^2$) and Bear Creek at the Union Hill Road (drainage area of 127 km$^2$). The basins of all of these streams have low levels of urban development (e.g., road densities range from 0 to 4.6 km/km$^2$).

Comparisons of the hydrographs of Novelty Hill to Evans Creek and of Evans Creek to Bear Creek show that smaller streams have higher area-normalized peak discharge rate and storm flow recession rate and a lower area-normalized base flow rate than the larger streams (Figure 2.2a). The time to peak discharge rate is also shorter for smaller streams.

The differences in stream flow patterns between smaller and larger streams parallel the expected differences between a stream with a higher level of urban development to one with a lower level of urban development. It is not clear if the hydrologic differences resulting from basin area are ecologically relevant. For example, macroinvertebrate and fish assemblages may be less diverse in small streams (Giller and Malmqvist, 1998). In any event, differences in drainage area may confound an analysis of the hydrologic effects of urban development. Accordingly, the influence of drainage area on the proposed hydrologic measures is analyzed here, with a goal that a robust measure of the hydrologic effects of urban development should be relatively insensitive to drainage area to the extent it does not influence the biological conditions of a stream.

Physiographic differences between stream basins, however, present opportunities to isolate the biological effects of stream flow patterns from other urban influences. In particular, differences in geologic and topographic conditions of streams in the Puget Lowland lead to a wide range of stream flow patterns at moderate levels of urban development. Much of the Puget Lowland region is underlain by glacial till, particularly on plateaus and uplands. Many valley bottoms are filled with glacial outwash deposits. ine grained (sand, silt, and clay) lacustrine deposits are common at lower elevations where glacial deposits have been eroded.

Stream flow patterns from Jenkins Creek (Figure 2.2b) exemplify the influences of glacial outwash deposits that readily infiltrate stormwater and lakes that attenuate flood waves. With both extensive outwash deposits and many lakes in its basin, Jenkins Creek has the lowest peak and highest base flow discharge rate during April 1991 of the streams in Figure 2.2. Physiographic conditions thus buffer the hydrologic effects of urban development on Jenkins Creek, such that its stream flow patterns are characteristics of streams with substantially lower levels of development.

Basin elevation is another physiographic factor that obfuscates the hydrologic differences between urban and suburban streams. For example, May Creek (Figure 2.2c) had the highest area-normalized, mean daily discharge rate of all of the streams in Figure 2.2 except for a tributary to Miller Creek and Des Moines Creek above the Tyee Pond (Figure 2.2e), which have very small drainage areas (<3 $km^2$) with high levels of urban development. The high area-normalized discharge rate in May Creek can be attributed to runoff from the part of the basin formed by the southern flank of Cougar Mountain. The elevation of this area (between 200 and 400 m above sea level) is higher than most of the Puget Lowland and receives more precipitation than the region in general. Quick-response runoff production may also be higher in May Creek because of shallow bedrock forming steep and impermeble hillslopes.

Huge Creek also had an anomalously high peak discharge rate during the April 1991 storm given its low level of development (Figure 2.2b). Huge Creek is located on the Kitsap Peninsula which receives more rainfall than, though the same storms as, other parts of the Puget Lowland. Huge Creek also has a smaller drainage area (16.6 $km^2$) relative to many of the streams considered in the analysis. The combination of these two conditions is likely the reason for the relatively high peak discharge rate during the April storm.

Physiographic factors in the May and Huge Creek basins promote “flashy” runoff production in ways that are analogous to the hydrologic effects of urban development, including high peak discharge rates relative to base flow, rapid storm flow recession, and relatively short duration storm flow. Given the high peak discharge rates and rapid storm flow recessions, stream flow patterns over annual and multiple-year time scales in May and Huge Creek are likely to be similar to those in streams with higher levels of urban development. These creeks are used to examine whether suburban stream flow patterns can be distinguished from urban stream flow patterns even when the suburban streams are “flashy” at a storm-scale. They illustrate the least difference that can be expected between urban and suburban stream flow patterns in the Puget Lowland.

Jenkins Creek, in contrast, with low peak discharge rates and gradual storm flow recessions, has one of the most attenuated stream flow patterns for suburban streams in the Puget Lowland. The difference between May and Jenkins Creeks, which have similar drainage areas and road densities, demonstrates the range in the hydrologic response of suburban Puget Lowland streams that cannot be explained solely in terms of land use.

### 2.3.3. Differences in storm flow patterns due to urban development

The characteristics of urban storm flow patterns are illustrated by the hydrographs for 1 to 15 April 1991 in Figure 2.2b and 2.2c. .Shortly after rain began falling on 3 April, urban streams (Leach, Miller, Mercer, and Swamp Creeks) rose rapidly. These streams attained high peak discharge rates early in the storm, and fell quickly when the rain ended on 5 April. One day later on 6 April, the area-normalized daily discharge rates for urban streams were lower than the rates for any of the suburban streams except Big Bear at Union Hill Road.

The rapid recession of storm flow in urban streams reflects less storage of water in the soils and surface depressions in urban basins and the rapid delivery of any stored water to stream channels via overland pathways, pipes, or open channels. The lower storm flow recession rates in suburban streams corresponds to the slow release of a greater supply of water stored in the soil and surface depressions on hillslopes and the slow delivery of the stored water to stream channels via relatively long subsurface pathways. The differences in storm flow recession rates are a reliable indicator of urban streams for this storm. Recession rates vary over time after a storm and between storms, so they do not have a characteristic value for a stream. Differences in recession rates do, however, produce differences in annual and inter-annual stream flow patterns that are summarized by the three hydrologic measures described below.

In spite of the differences in the rate of change in discharge between urban and suburban storm flow patterns, the area-normalized peak daily discharge rates for the urban streams are not greater than those of many suburban streams (e.g., May, East Fork Issaquah, and Canyon Creeks). The storm flow response and recession of the urban streams occur at time scales less than a day. Consequently, daily mean discharge rate data do not fully resolve differences between urban and suburban streams. On 5 April, the peak 15-minute discharge rate was 44% higher than the daily mean discharge rate in Miller Creek (7.1 $m^3/s$ versus 4.9 $m^3/s$), 31% higher for Huge Creek (5.1 $m^3/s$ versus 3.9 $m^3/s$), and only 15% higher for May Creek (12.8 $m^3/s$ versus 11.1 $m^3/s$). These data show area-normalized peak discharge rate for a single storm is not a reliable basis for distinguishing between urban and suburban streams where basins vary in their size, location, geology, or topography.

### 2.3.4. Differences in base flow due to urban development

Base flow is the portion of stream flow supplied by steady ground water discharge. The magnitude of base flow in a stream varies seasonally, though much more gradually than storm flow, as groundwater discharge to the stream increases during winter and declines during summer. In ephemeral reaches, groundwater levels decline during summer below the surface of the stream bed and surface flow ceases.

The effects of urban development on base flow in Puget Lowland streams are more ambiguous than the effects on storm flow. For April 1991, stream flow declines to a relatively constant or base level by 6 April in smaller urban streams (Figure 2.2c and 2.2e) and by 13 April in the intermediate-size urban streams (Figure 2.2b). Area-normalized discharge is higher in the suburban streams during the base-flow period from 13 to 15 April.

Higher area-normalized wet-season base flow in suburban streams is likely a result of greater subsurface flow generated by water stored in the soil column, stream banks, and other shallow aquifers. In contrast, area-normalized wet-season base flow in urban streams is lower due to less soil storage capacity, lower surface infiltration rates, less aquifer recharge, and extended drainage networks in urban areas.

In the late spring and summer, discharge of urban streams declines to dry-season base levels earlier than suburban streams. For the period from 1 May to 1 September 1998, discharge rates from basins with higher levels of urban development (Leach, Mercer, Miller, and Swamp Creeks) decline to a steady level by 31 May whereas base flow from less urban basins (Huge, Jenkins, May, and Rock Creeks) continues to decline throughout the summer (Figure 2.4). The earlier recession of urban stream flow to base levels would be difficult to identify with a simple average measure such as mean dry-season discharge rate because of periodic storms during the summer when more runoff is produced from urban basins than suburban basins.

Later in the summer, there are no evident differences in area-normalized discharge between urban and suburban streams. Figure 2.3 shows the mean discharge rate for August 1994 plotted by drainage area (log scale) for 19 urban and 19 suburban streams in the Puget Lowland. August 1994 was a dry month and discharge was at its lowest level for the period from WY 1989 to 1998 in many streams in the region. The mean discharge rate for August 1994 varies over an order of magnitude in the neighborhood of any given drainage area. There is no systematic difference in the discharge rate between urban and suburban streams. Instead, the variation of dry-season discharge reflects the influences of physiographic factors (e.g., rainfall, geology, vegetation, and topography) not controlled for in this analysis.

### 2.3.5. Differences in annual and inter-annual stream flow patterns due to urban development

Flow patterns at the scale of a storm or season illustrate hydrologic differences between urban and suburban streams, but they do not represent the differences over longer periods of time, which are likely to be ecologically relevant. This section shows how the storm and seasonal-scale differences between urban and suburban streams manifest as differences at annual and inter-annual time scales. Annual mean runoff depth (i.e., annual discharge divided by drainage area) during WY 1989 to WY 1998 was significantly lower on average ($p < 0.01$ using a Student's t-test for samples with unequal variances, Helsel and Hirsch, 1993, p. 126) for 11 urban streams (1.2 mm/day or 0.44 m/year) than 12 suburban streams (1.8 mm/day or 0.66 m/year) in the Puget Lowland (Table 2.2). The lower annual discharge stands in contrast to increased runoff production observed for urban hillslopes and higher storm flow observed in urban streams. Lower recessional and wet-season base flows must offset the higher storm flow production in these urban streams.

Annual and inter-annual stream flow patterns can be examined using flow-duration curves. A flow-duration curve shows the fraction of a period of record that a given discharge rate that was exceeded in a stream, providing the complement of the cumulative distribution function of discharge (Vogel and Fennessey, 1994). While flow-duration curves show the duration of the full distribution of flows in a stream, they do not indicate the sequence of those flows.

Two types of flow-duration curves are used here to compare urban and suburban stream flow patterns: period of record curves, and median annual flow-duration curves. Period of record (POR) flow-duration curves represent the fraction of a period of record that a given discharge rate is exceeded. A POR flow-duration curve represents the full range of stream flow from the largest floods to the lowest flows over multiple-year periods, but it does not distinguish annual variability from inter-annual variability. The discharge at any quantile (i.e., fraction of time) of a POR flow-duration curve will vary depending on the specific period of record used to construct the curve. In particular, extremely wet or dry periods will cause the tails of the curve to shift up or down, respectively.

POR flow-duration curves cannot be used to analyze land use changes over time in the Puget Lowland because there are not long records available before and after the period of urban development during which climatic conditions were stationary. Even comparisons of contemporary stream flow patterns using POR flow-duration curves may be dubious since the period when the gage was operating varies from stream to stream. POR curves in this analysis are based on 15-minute discharge data.

Median annual (MA) flow-duration curves are less sensitive to period of record of the discharge data than POR flow-duration curves, but they do not show the extreme high and low discharge rates for a stream. A MA flow-duration curve is constructed, first, by calculating individual flow-duration curves for each year in the period of record for a stream. Then, the median discharge rate of the annual flow duration curves at each quantile is used to form the MA flow-duration curve. A MA flow-duration curve represents the discharge rate with a 50% probability of being exceeded for a given duration in any year (Vogel and Fennessey, 1994).

MA flow duration curves are used here to analyze changes in stream flow patterns over time and in the comparison of contemporary stream flow patterns since they are less sensitive to data gaps at different times among the streams during the periods of analysis. The MA flow-duration curves in this analysis are derived from mean daily discharge rate data.

POR and MA flow-duration curves correspond closely for all quantiles except short duration, high flows. Figure 2.5 shows a series of four flow-duration curves for Huge, Miller, May, and Mercer Creeks during WY 1989 to 1998. The discharge rates for the MA and POR flow-duration curves are similar except for those exceeded less than 1% of the time. The minimum and maximum annual flow-duration curves illustrate the large inter-annual variation in stream flow at all quantiles in these streams.

MA flow-duration curves are shown in Figure 2.6 for 9 streams spanning a range of development in the Puget Lowland. These streams have drainage areas ranging from 17 to 37 $km^2$. Each curve is based on median annual discharge rates for 8 to 10 years during the period from WY 1989 to WY 1998, except for Rock Creek. While the Rock Creek record spans only 4 years, it includes WY 1996 and 1997 which were wet years with large storms throughout the region and WY 1998 which was a dry year. The curves are truncated at the discharge rate exceeded 1% of the time since these short duration flows are not well represented by mean daily discharge.

MA flow-duration curves have been normalized by drainage area in Figure 2.7 to facilitate comparisons between streams of different sizes. A difference in area-normalized discharge between urban and suburban streams is most evident over the range of discharges exceeded 20 to 40% of the time: suburban streams (Jenkins, Rock, and May Creeks) had the highest area-normalized discharge rates in these intermediate quantiles while urban streams (Leach and Miller Creeks) had the lowest discharge rates.

The discharge rates in this quantile range represent storm flow recessional and wet-season base flows, which are expected to be lower in urban streams than in suburban streams. The result agrees with the differences in storm flow recession observed the in the April 1991 hydrographs (Figures 2.2a to 2.2e).

Urban stream basins produce storm flow for only brief periods during storms. As a result, area-normalized discharge rates are higher in urban streams than in suburban streams only for flows exceeded less than 1% of the time. Given the “flashiness” of urban streams, the discharge rate for short duration quantiles (e.g., less than 1% of the time) should be estimated using hourly or 15-minute stream flow data rather than daily mean discharge rates.

Examples of flow-duration curves constructed with daily mean and 15-minute mean discharge rates are illustrated in Figure 2.8 for Miller and May Creeks. The discharge rate exceeded 1% of the time for Miller Creek is 12% higher (2.0 $m^3/s$ compared to 1.8 $m^3/s$) when the flow-duration curve is based on 15-minute data rather than daily mean discharge data. The difference is not as pronounced for May Creek except for the maximum observed discharge rate. As a result, the comparison of high flows (i.e., short duration) between urban and urban streams requires high-resolution (e.g., hourly or shorter) stream flow data and a consistent period of record.

High-resolution discharge data are not necessary, however, to discern differences in annual stream flow patterns of urban and suburban streams. In particular, the area-normalized discharge rate in the urban streams shown in Figure 2.7 is lower than the discharge in suburban streams for the discharges exceeded 20 to 40% of the year which represents periods of storm flow recession and wet-season base flow. While these discharges are not associated with extremely high or low flows, they represent common hydrologic conditions during the winter and spring in the Puget Lowland that may nonetheless have a significant ecological role.

## 2.4. Measures of annual and inter-annual stream flow patterns

The differences between urban and suburban stream flow patterns over annual and multiple-year periods are examined here using three measures: (1) the fraction of a year that the daily mean discharge rate exceeds the annual mean discharge rate ($T_{Qmean}$); (2) the cumulative fraction of a multiple year period that the discharge rate of a specified flood quantile is exceeded ($T_{X\ yr}$ is the cumulative duration that stream flow exceeds the discharge of a flood that is exceeded on average 1/X times per year); and (3) the coefficient of variation of the annual maximum flood ($CV_{AMF}$). These measures are applied to discharge records for most of the Puget Lowland streams listed in Table 2.1 for the period from WY 1989 to 1998 to characterize differences between urban and suburban stream flow patterns. The first two measures are also applied to discharge records from streams gaged during WY 1960 to 1969 to compare how the measures are influenced by changes in land use and physiographic conditions (i.e., storm patterns) over time. The analysis of changes in stream flow patterns over time includes six urban streams (Juanita, Leach, Mercer, Swamp, and Clover Creeks) and six suburban streams (Huge, May, Rock, Newaukum, Soos, and Issaquah Creeks)

Differences in both $T_{Qmean}$ and $T_{X\ yr}$ between urban and suburban streams are expected because of the differences in peak discharge and recession rates, and the lack of differences in annual discharge for these two groups of streams. Likewise, differences in the $CV_{AMF}$ between urban and suburban streams are expected given the results of rainfall-runoff modeling by James (1965) and the observations of Hollis (1975), which both showed a greater relative increase in the magnitude of small, frequent floods than the relative increase of large, infrequent floods in response to urban development.

### 2.4.1. Fraction of time that mean daily discharge rate is exceeded

Comparisons of flow-duration curves between streams and between different time periods for individual streams are facilitated by normalizing discharge by a reference discharge that accounts for broad differences in drainage area and rainfall totals. Saville and Watson (1933) used mean discharge rate as the basis for non-dimensional flow-duration curves in a regional analysis of North Carolina streams. Likewise, the mean discharge rate for individual water years can be used to normalize discharge when comparing flow duration curves for wet and dry years in a single stream. However, flow duration curves normalized by annual mean discharge do not account for many of the physiographic factors influencing intra-annual patterns of stream flow generation. Morgan (1936, p 425) observed that the distribution of daily flows relative to the mean "is affected by topography, arrangement of tributaries with regard to time of concentration of surface-flow, geologic structure, soil, vegetation, weather, and human developments related to flow of water."

The hydrologic effects of urban development are evident in flow duration curves normalized by mean discharge, even amidst the variability generated by physiographic differences among the basins in the Puget Lowland. The POR flow-duration curves for nine streams have been normalized with respect to the mean discharge rate in Figure 2.9. In urban streams, the fraction of time that the mean discharge rate is exceeded, $T_{Qmean}$, (i.e., $Q_{daily}/Q_{mean}$ exceeds 1) generally is less than 30%, while $T_{Qmean}$ is generally greater than 30% in suburban streams. The lower values of $T_{Qmean}$ in urban streams are a result of more rapid storm flow recession and lower wet-season base flow. The difference in $T_{Qmean}$ between urban and suburban streams corresponds to the observation of lower discharge in urban streams for the 20 to 40% quantiles of area-normalized flow duration curves.

The patterns in $T_{Qmean}$ observed for the nine streams above was analyzed using stream flow records for 23 Puget Lowland streams for the period from WY 1989 to 1998. The fraction of the year that daily mean discharge rate ($Q_{daily}$) exceeded the annual mean discharge rate ($Q_{mean}$) was determined for each year of record for each stream. $T_{Qmean}$ was calculated as the average annual fraction that $Q_{daily} > Q_{mean}$.

$T_{Qmean}$ generally varies inversely with urban development among Puget Lowland streams (Figure 2.10). The mean value of $T_{Qmean}$ during WY 1989 to 1998 for 11 urban streams was 0.29 while it was 0.34 for 12 suburban streams (Table 2.2). The difference is statistically significant ($p < 0.01$ using Student's t-test of samples with equal variance). Suburban streams had values of $T_{Qmean}$ greater than or equal to 0.32 with the exception of Huge Creek. Urban streams had values of $T_{Qmean}$ less than or equal to 0.31 with the exception of Clover Creek. $T_{Qmean}$ is lower in urban streams because of increased storm flow volume, rapid recession rates, and lower wet-season base flow.

Huge and Clover Creeks stand out from the other data points in Figure 2.10. Huge Creek has a relatively low value of $T_{Qmean}$ (26%) among suburban streams. Two factors may account for the low value of $T_{Qmean}$ in Huge Creek: it has a smaller drainage area (17 $km^2$) and receives greater rainfall during storms. As noted in the earlier analysis of stream hydrographs, these characteristics promote a "flashy" storm flow pattern in Huge Creek with high peak discharge rates and rapid recession rates. In contrast, Clover Creek has a high value of $T_{Qmean}$ (41%) in spite of a moderate level of urban development (road density 6.7 $km/km^2$). The high value of $T_{Qmean}$ for Clover Creek likely results from the influence of lakes and permeable glacial outwash soils in its basin, which attenuate higher flows and sustain discharge during dry periods. Clover Creek is also the largest urban streams analyzed here.

Larger streams typically have more attenuated stream flow patterns than smaller streams and, as a consequence, higher values of $T_{Qmean}$ (Figure 2.12). The mean value of $T_{Qmean}$ for large (drainage greater than 30 $km^2$) streams is 0.35 and significantly greater than the mean value for smaller (drainage area < 30 $km^2$) streams which was 0.28. Given the observed and potential influence of physiographic factors (e.g., geology, storm patterns, drainage area), $T_{Qmean}$ may be a reliable indicator of urban development only for comparison between stream basins with similar physiographic conditions. However, an analysis of the mean values of $T_{Qmean}$ between urban and suburban streams with drainage areas greater than 20 $km^2$ still indicates significantly lower values in urban streams ($p < 0.01$ based on Student's t-test of samples with unequal variance).

In a stream with stable land use, $T_{Qmean}$ varies little from year to year. The coefficient of variation for annual values of $T_{Qmean}$ during the period of 1989 to 1998 was less than 17% for all of the streams except a small tributary to Miller Creek (Table 2.3). Since $T_{Qmean}$ does not display high inter-annual variability, it can be estimated reliably from a relatively short (e.g., ~ 10 years) stream flow record.

$T_{Qmean}$ for a stream changes over a period of urban development. Annual values of $T_{Qmean}$ for Mercer Creek illustrate a systematic decline during a period of urban development from 1960 to 1998 (Figure 2.10). Changes in the value of $T_{Qmean}$ over time were analyzed for 10 Puget Lowland streams with stream flow records spanning multiple decades. The analysis compares values between an earlier period (A) which spans WY 1960 to WY 1969 for most streams and a later period (B) which spans from WY 1989 to WY 1998. Table 2.4 provides the number of years of record for each period, the value of $T_{Qmean}$ and its coefficient of variation.

Differences in $T_{Qmean}$ between these periods were examined for each stream using a one-sided rank sum test (Helsel and Hirsch 1993, p. 118). Under this test, the values of $T_{Qmean}$ for the two periods are ranked together. The sums of the rankings for Period A and B are compared. The critical values for the rank sum of $T_{Qmean}$ for period A, listed in Table 2.4 for each stream, indicate when the probabiltiy that annual values of $T_{Qmean}$ during period A were equal to or less than the annual values of $T_{Qmean}$ during period B is less than 5%.

Annual values of $T_{Qmean}$ were significantly lower ($p < 0.05$) in period B than in period A for Juanita, Leach, Mercer, and Newaukum creeks while $T_{Qmean}$ did not vary significantly over time in the other basins. Juanita, Leach, and Mercer Creeks are in urban basins where extensive development occurred between periods A and B. In contrast, Newaukum Creek has relatively low levels of urban development, though silvicultural and agricultural activities may have contributed to hydrologic change.

### 2.4.2. Coefficient of variation of the annual maximum flood distribution

Increased flood magnitude is a characteristic effect of urban development on stream flow. Differences in flood patterns between urban and suburban streams were evaluated by comparing the annual maximum flood distributions for 25 Puget Lowland streams spanning the range of urban development in the region. The maximum instantaneous (or 15-minute mean) discharge rate for each water year was selected for the period WY 1989 to 1998. The geometric mean annual maximum peak rate and the area-normalized value of the mean were calculated for each stream. Flood distributions for an earlier period (WY 1960 to 1969) were constructed for 12 of the streams to assess changes over time in the flood distributions of individual streams.

Simulated and observed effects of urban development show a differential increase in the magnitude of smaller, frequent floods relative to larger floods (James 1965, Hollis 1975). As a result, the coefficient of variation of annual maximum floods ($CV_{AMF}$) in a stream should decrease in response to urban development. $CV_{AMF}$ was tested as a discriminant of urban stream flow patterns from suburban stream flow patterns. The $CV_{AMF}$ was calculated assuming that annual maximum flood peaks follow a two parameter log-normal distribution:

$$CV_{AMF} = \sqrt{\exp(\sigma_Y{}^2) - 1} \qquad (2.1)$$

where $\sigma_Y{}^2$ is the variance of the natural logarithms of the annual maximum flood discharge rates (Stedinger et al 1993). The two parameter log normal distribution is an appropriate assumption given the generalized skew coefficient for annual maximum floods in the Puget Lowland region is 0.02 (Interagency Advisory Committee on Water Data, 1982).

The area-normalized peak discharge rates of mean annual floods are higher in urban streams than suburban streams (Table 2.2). The mean peak discharge of urban streams was 0.34 $m^3/s/km^2$ while the mean peak discharge of suburban streams was 0.2 $m^3/s/km^2$. Physiographic conditions other than drainage area have a strong influence on peak discharge rates from a stream during a flood. As a result, streams with mountain headwaters (e.g., May, Issaquah, and Newaukum Creeks) or on the Kitsap Pennisula (e.g., Huge Creek) have relatively high area-normalized flood peaks.

The variation in the flood distribution for urban streams is lower than the variation in the flood distribution for suburban streams (Figure 2.13). The urban streams had a mean $CV_{AMF}$ of 0.5 while the suburban stream had a mean $CV_{AMF}$ of 1.0 (Table 2.2). The lower values of $CV_{AMF}$ in urban streams may be due, in part, to a basin-scale effect.

Smith (1992) showed that drainage area influences flood distribution properties including a peak in the variability of the $CV_{AMF}$ for stream basins between 26 and 260 $km^2$. As a result, the largest values of $CV_{AMF}$ for small basins (drainage area <26 $km^2$) are expected to be lower than the largest values of $CV_{AMF}$ for a stream of intermediate size. The peak in the variability of $CV_{AMF}$ for intermediate-sized basins (drainage area > 20 $km^2$) is evident for Puget Lowland streams (Figure 2.14).

The geometric mean annual flood in streams throughout the Puget Lowland, increased from the earlier period (WY 1960 to WY 1969) to the later period (WY 1989 to WY 1998) (Table 2.4). All of the urban streams in Table 2.4, except Swamp Creek, had an increase in the geometric mean annual flood of more than 72% where as only one suburban stream, Rock Creek, had an increase greater than 72%.

The $CV_{AMF}$ for all of the streams except Clover Creek are larger during period B than period A. $CV_{AMF}$ is not stationary from decade to decade even in basins with only marginal changes in land use over time. Consequently, $CV_{AMF}$ may not be useful for characterizing hydrologic change due to land use in a stream basin, but it may discriminate between urban and suburban streams for a common period of time.

### 2.4.3. Fraction of time that stream flow exceeds the magnitude of a frequent flood

The fraction of time, $T_{X\ yr}$, that stream flow exceeds the magnitude of a flood with an average frequency of 1/X times per year compares the frequency of a discharge to its cumulative duration. $T_{X\ yr}$ is influenced by the frequency of storms, the duration of storm flow, and recession rates after storms. Streams with short-duration storm flow and rapid storm flow recession have low values of $T_{X\ yr}$.

Values of $T_{X\ yr}$ were estimated for 11 Puget Lowland streams from flow duration curves for a series of frequent floods. Both flow duration and flood frequency were based on 15-minute stream flow data, with the exception of Mercer Creek where daily mean discharge data were used to construct a flow duration curve. The period of record varies among the streams from 4 to 10 years. The annual flood frequency for each stream was calculated from a partial duration series (Langbein, 1949). The partial duration series used here comprises stream flow peaks (i.e., local maxima in a hydrograph) that exceeded a stream-specific threshold discharge rate and were separated by at least 10 days. Where multiple peaks occurred within 10 days of each other, the highest value was used. The threshold discharge rate for each stream was selected so that each series had 30 to 50 floods.

The cumulative duration of time that stream flow exceeds the magnitude of a flood with a given frequency is shorter in urban streams than suburban streams (Figure 2.15). In particular, the cumulative duration of time that discharge exceeds the magnitude of a flood occurring twice a year on average ($T_{0.5\ yr}$) is less than 0.01 for all of the urban streams and more than 0.01 for all of the suburban streams. $T_{0.5\ yr}$ distinguishes between streams with moderate levels of urban development (Cedar tributary 0308; Swamp, North, and May Creeks) from those with lower levels (Rock, Jenkins, and Covington Creeks). Moreover, $T_{0.5\ yr}$ varies with road density within these groups. The relationship does not appear to vary with drainage area, as demonstrated by the relatively small variation between the curves for North Creek (100 $km^2$) and Cedar Tributary 0308 (2.7 $km^2$) which have similar, moderate levels of urban development in their basins.

## 2.5. Evaluating hydrologic influences on benthic biological conditions

The three hydrologic measures ($T_{Qmean}$, $CV_{AMF}$ and $T_{0.5\ yr}$) of annual and inter-annual stream flow patterns are compared to the biological conditions of Puget Lowland streams using the benthic index of biological integrity (B-IBI) developed for gravel-bed streams in the Puget Lowland (Karr and Chu, 1999). B-IBI is a multimetric index that includes measures of the taxonomic diversity, trophic and age structures, and life histories of benthic macroinvertebrate assemblages. B-IBI ranges over a scale from 10 to 50 with higher scores indicating greater diversity particularly among three orders of insects (Ephemeroptera, Plecoptera, and Trichoptera), clingers, more taxa generally intolerant of poor water quality, and long-lived genus; higher proportion of predators; lower proportions of taxa generally tolerant of degraded water quality, and greater population eveness among the dominant taxa. Morley (2000) provides a complete description of the B-IBI and scores for Big Bear, Swamp, Miller, North, Jenkins, May, and Rock creeks. Kliendl (1995) calculated the B-IBI scores for Big Beef, Juanita, Mercer (Kelsey), Covington, Des Moines and Hylebos Creeks. Where B-IBI was calculated for multiple sites on a stream, the site closest to the stream gage is presented here.

The biological condition of a stream generally varies with the level of urban development in the stream's basin. Figure 2.16 shows the relationship between B-IBI and road density for the 13 Puget Lowland streams spanning the range of urban development in the region. Urban streams have B-IBI values less than 28 while suburban streams have B-IBI values greater than 24. Big Beef Creek with a low road density (2.1 $km/km^2$) and a low B-IBI (26) lies furthest from an inverse relationship between urban development and biological conditions. While no single factor is likely to explain the variation in B-IBI over the region, the influence of annual and inter-annual stream flow patterns on biological conditions in streams is examined using the three hydrologic measures described above.

The biological conditions of stream are likely to vary with any of the hydrologic measures, because each measure varies with urban development. The relationship between B-IBI and stream flow patterns is less variable for $T_{Qmean}$ (Figures 2.17) than for $CV_{AMF}$ (Figure 2.18). B-IBI ranges by 10 points in the neighborhood of any given value of $T_{Qmean}$ but by over 20 points for any value of $CV_{AMF}$. The relationship between B-IBI and $T_{0.5\ yr}$ is the least variable (Figure 2.19), but only 8 streams were analyzed.

## 2.6. The hydrologic effects of urban development on annual and inter-annual stream flow patterns

The three hydrologic measures characterize stream flow patterns of progressively shorter durations and higher magnitude variability. Mean annual discharge is exceeded approximately 30% of the time in Puget Lowland streams. The discharge rate of a flood occurring, on average, twice a year is exceeded less than 10% of the time. The discharge rate of an annual maximum floods is exceeded less than 2% of the time.

For Puget Lowland streams, the mean annual discharge rate ($Q_{mean}$) is representative of recessional flow after storms and wet-season (winter and spring) base flow. Urban development increases storm flow recession rates and decreases wet-season base flow in streams, so that there are more days when stream flow is less than $Q_{mean}$. $T_{Qmean}$ has low inter-annual variability so it can be used to analyze the hydrologic effects of urban development over time for a stream and to compare flow patterns in urban and suburban streams even when those streams have different periods of recorded discharge. Physiographic factors, such as lakes, surficial geology, topography, and basin area, influence the value of $T_{Qmean.}$ .For basins less than 20 $km^2$, the value of $T_{Qmean}$ may be relatively low (<30%) even in suburban basins. As a result, $T_{Qmean}$ may have limited application for assessing stream flow patterns in such small basins. For basins larger than 20 $km^2$, $T_{Qmean}$ was greater than 30% in suburban streams and less than 30% in urban streams

$T_{0.5\ yr}$ indicates the duration over a multiple year period that discharge exceeds a nominal level equal to the magnitude of a common flood (i.e., one occurring on average twice a year). $T_{0.5\ yr}$ varies inversely with road density such that urban streams have low values (i.e., short durations) and suburban streams have higher values. While a continuous time-series of high resolution (e.g., 15-minute) discharge data is needed to assess streams using $T_{X\ yr}$, it does not appear to be sensitive to basin-scale.

$CV_{AMF}$ is an indicator of the variation over a multiple year period of the largest annual flood in a stream. Urban streams flow patterns can be distinguished from suburban stream flow patterns, even for smaller streams, using $CV_{AMF}$. $CV_{AMF}$ is generally lower for urban streams than suburban streams but it varies considerably over periods of decades even when land use is fairly stable. Consequently, $CV_{AMF}$ is not useful for assessing land use change over time or differences between streams with gages covering different periods of record. While $CV_{AMF}$ appears less sensitive to basin-scale (i.e., drainage area) than does $T_{Qmean}$, a larger sample, particularly of small (< 20 $km^2$) suburban streams and large (>20 $km^2$) urban streams is necessary to confirm this preliminary observation.

The hydrologic measures not only discriminate urban from suburban stream flow patterns but also indicate two potential mechanisms of geomorphic instability in urban streams. For the first mechanism, inter-annual patterns of large floods control the geomorphic stability and instability in streams. Frequent floods will be more geomorphically effective (i.e., produce extensive bed disturbance) as they approach the magnitude of infrequent floods. In this case, $CV_{AMF}$ will be low as was observed for urban streams in the Puget Lowland. Wolman and Miller (1960) articulated this mechanism as the principle that when the variability of the magnitude of geomorphically effective events is low, frequent events will have a dominant influence shaping a landscape. In a stream with a low $CV_{AMF}$, frequent floods can be expected to be effective in transporting sediment. In a stream with high $CV_{AMF}$, frequent floods can be expected to be less effective in transporting sediment because larger, infrequent floods have expanded the channel and armored the bed (i.e., formed a coarse layer at the bed surface).

For the second mechanism, storm-scale stream flow patterns control geomorphic stability and instability of stream beds. The low values of $T_{Qmean}$ and $T_{X/yr}$ indicate rapid recession rates in storm flow and a relatively short duration of intermediate to high flows. Short duration flows may be insufficient to exhaust the local supply of small and unconstrained particles from a gravel stream bed (Reid and Laronne, 1995; ) and, thus, will fail to form a stable armor layer. Frequent but short-duration periods of high flows will continue to entrain and transport bed material. These two potential mechanisms of geomorphic instability are evaluated in Chapter 5.

The biological conditions of macroinvertebrate assemblages in streams are most closely related to annual stream flow patterns, as represented by $T_{Qmean}$ and $T_{2/yr}$ rather than inter-annual variation in the distribution of large floods as represented by $CV_{AMF}$. B-IBI generally varies with $T_{Qmean}$ indicating higher taxonomic diversity, more complex trophic structure, and less dominance (in terms of relative number of organisms) of the most common taxa in the macroinvertebrate assemblages inhabiting streams where stream flow exceeds the annual mean discharge rate during a greater fraction of the year. Annual stream flow patterns are likely to influence the biological conditions of streams by means other than flood disturbance. For example, $T_{Qmean}$ varies with wet-season base flow indicating the area or depth of aquatic habitat available in a stream or the flux of nutrient through a reach in the spring. As a result, annual stream flow measures may be ecologically-relevant beyond their indication of physical disturbance patterns.

## 2.7. Conclusions

Urban development in the Puget Lowland modifies stream flow patterns including increased peak discharge during storms, more rapid recession of stream flow after storms, and lower wet-season base flow. These hydrologic changes manifest as changes in annual and inter-annual stream flow patterns during periods of urban development as well as differences in stream flow patterns between urban and suburban streams. In urban streams, the annual mean discharge rate is exceeded during fewer days of the year, the coefficient of variation in annual maximum floods is lower, and the cumulative duration that the discharge rate exceeds the magnitude of frequent floods is lower than in suburban streams. The biological conditions of macroinvertebrate assemblages in streams are most closely related to annual stream flow patterns, as represented by $T_{Qmean}$ and $T_{0.5\ yr}$ but not the inter-annual variation in the distribution of large floods as represented by $CV_{AMF}$.

The fraction of a year that stream flow is greater than the annual mean discharge rate ($T_{Qmean}$) varies with drainage area and other physiographic conditions, so this measure can only be used to compare similar stream basins. $T_{Qmean}$ does, however, indicate hydrologic change over time in a stream basin.

In contrast to $T_{Qmean}$, the coefficient of variation of annual maximum floods ($CV_{AMF}$) is highly variable over time so it is not a reliable indicator of the hydrologic consequences of land use changes in a stream basin. $CV_{AMF}$ serves as a basis for comparing the high flow regime of a group of streams during a common time period. The relationship between B-IBI and $CV_{AMF}$ is relatively weak compared to other hydrologic measures.

The cumulative duration that the discharge rate exceeds the magnitude of frequent floods ($T_{X/yr}$) is a robust indicator of urban development in the Puget Lowland showing little sensitivity to drainage area. Since $T_{X/yr}$ is the cumulative time that discharge is greater than a flood's magnitude, it indicates the potential stability of gravel-bed stream channels under that flood with greater stability likely for flows that occur for longer durations. $T_{X\ yr}$, however, must be estimated using a time series of high temporal resolution (e.g., 15 minute or hourly) discharge data from a period of multiple years.

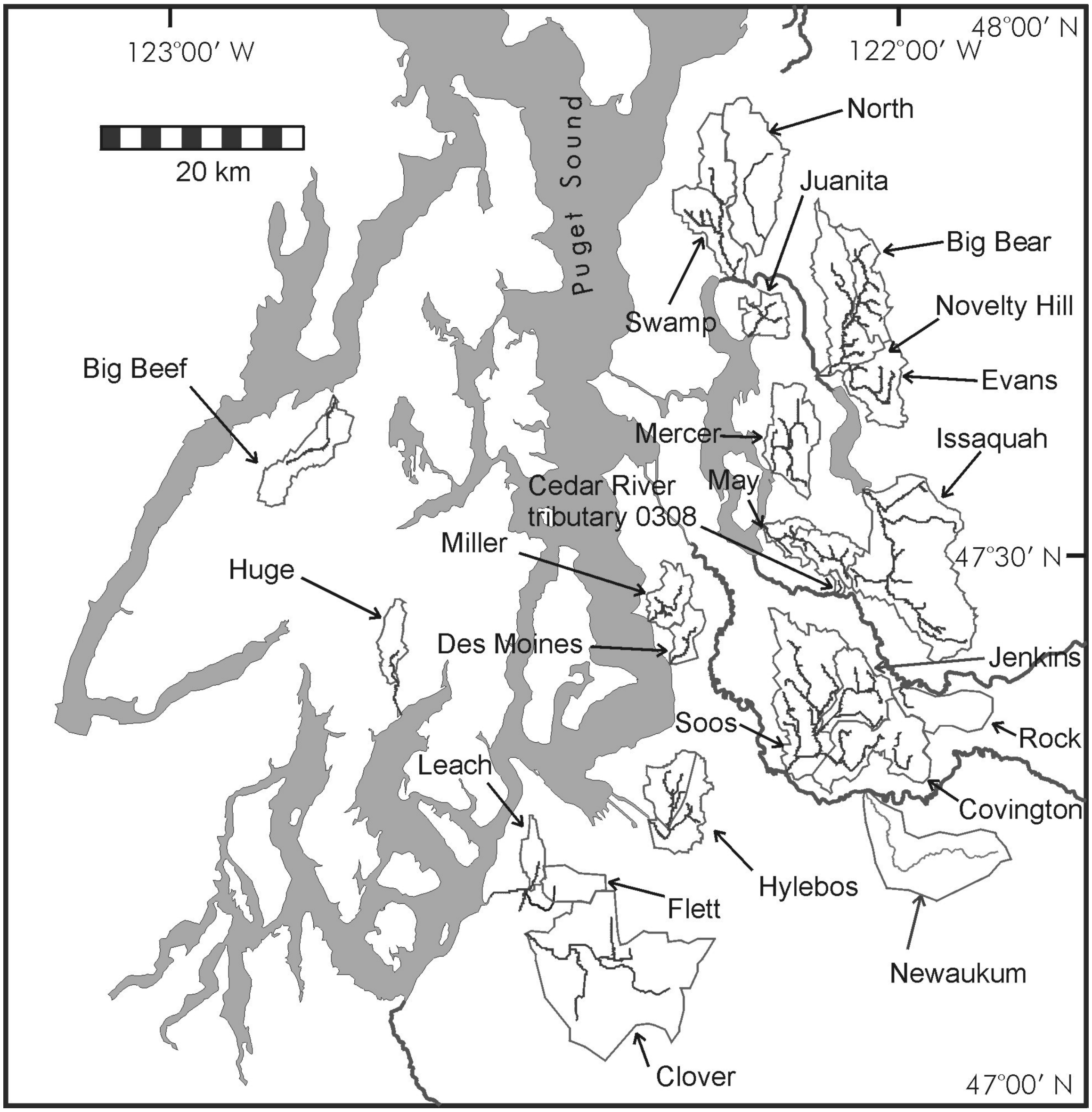
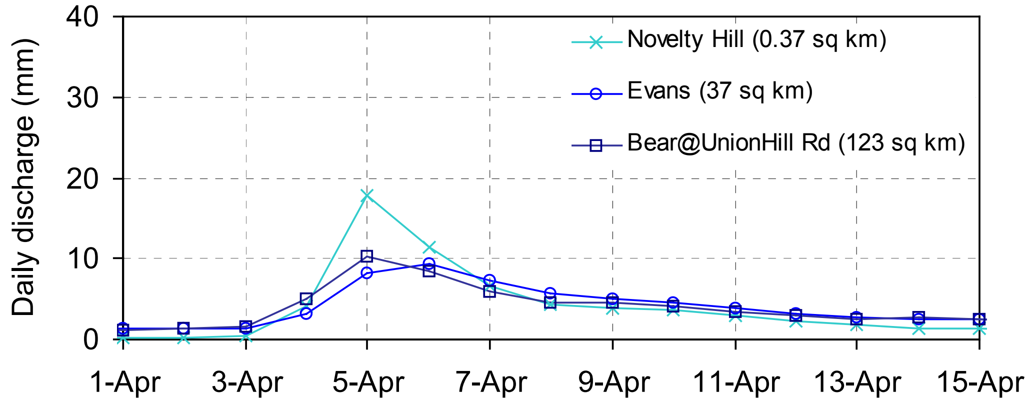
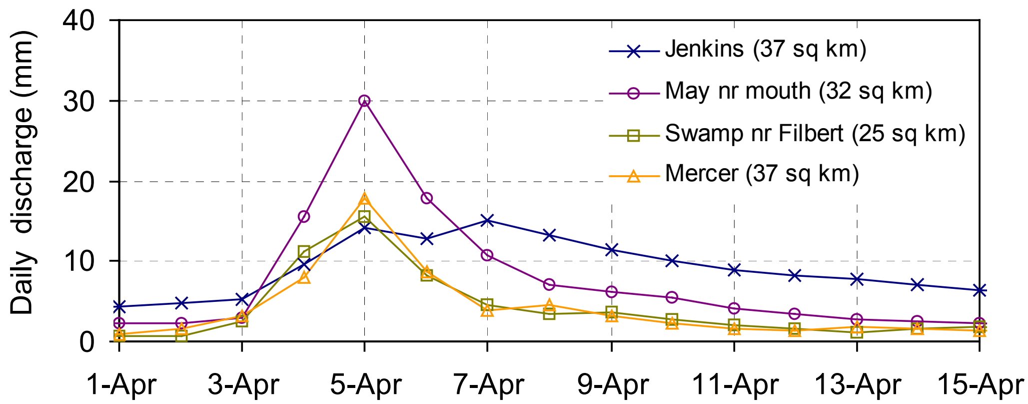



Figure 2.1: Map of streams in the Puget Lowland included in the hydrologic analysis.

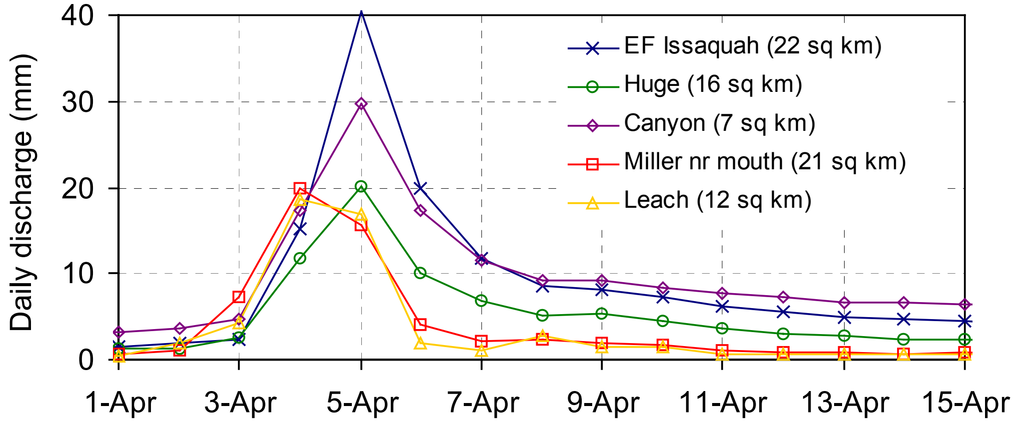



(a) netsted suburban streams

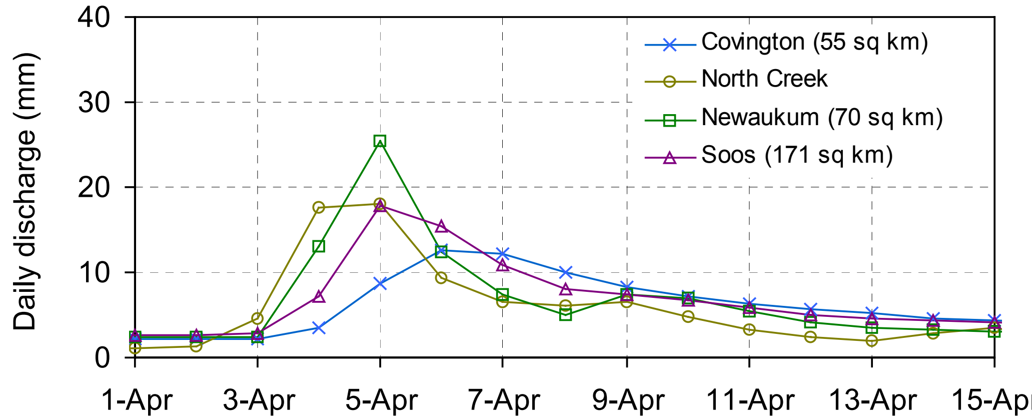



(b) intermediate-size urban and suburban streams

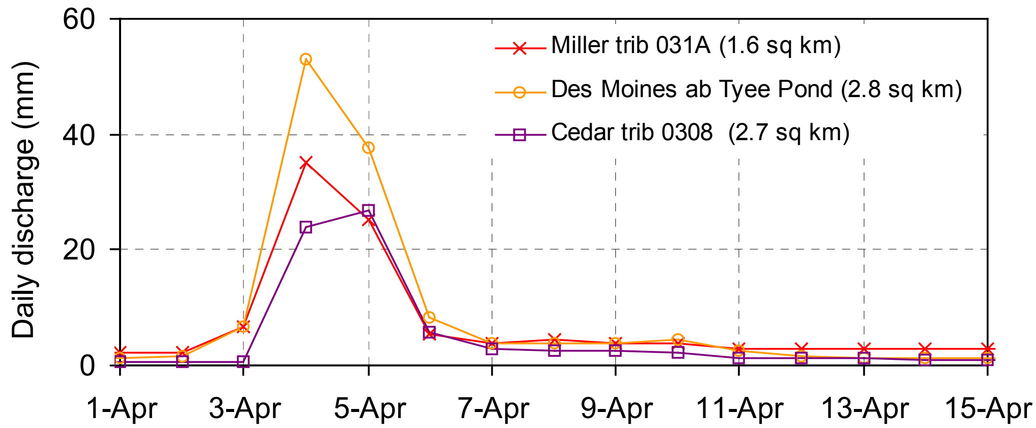
Figure 2.2: Area-normalized daily discharge for Puget Lowland streams from 1 to 15 April 1991 with drainage areas listed in parenthesis: (a) nested suburban streams, (b) intermediate-size urban and suburban streams; (c) smaller urban and suburban streams, (d) large suburban streams, (e) small urban streams. Note scale for runoff changes in (e).




(c) smaller urban and suburban streams




(d) large streams




(e) small urban streams

Figure 2.2 (continued).

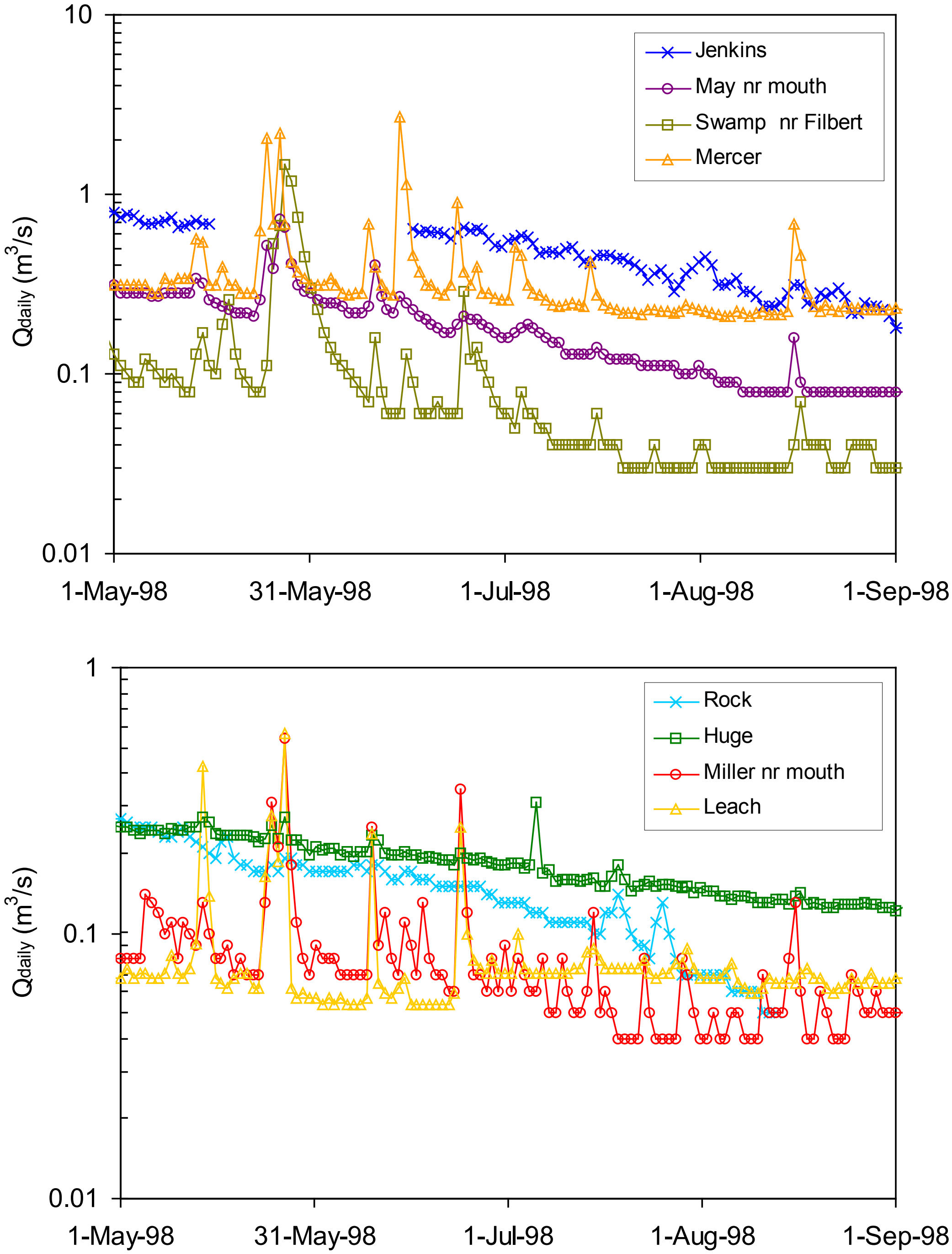



Figure 2.3: Daily mean discharge rate ($Q_{daily}$) for 8 Puget Lowland streams from 1 May to 1 September 1998.

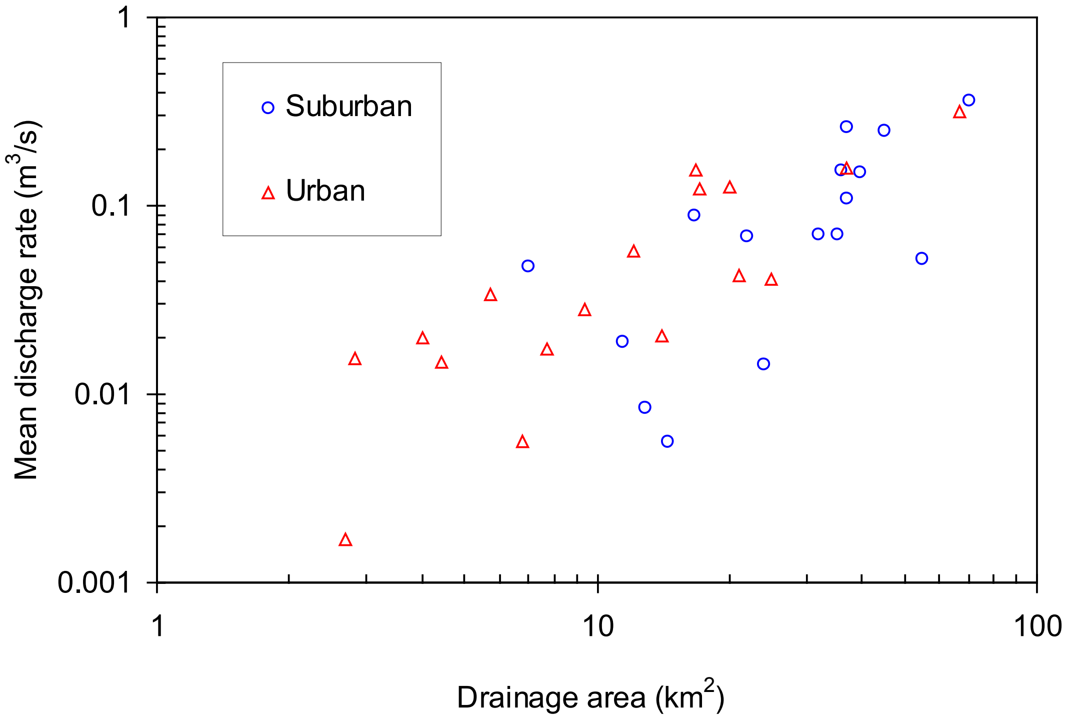
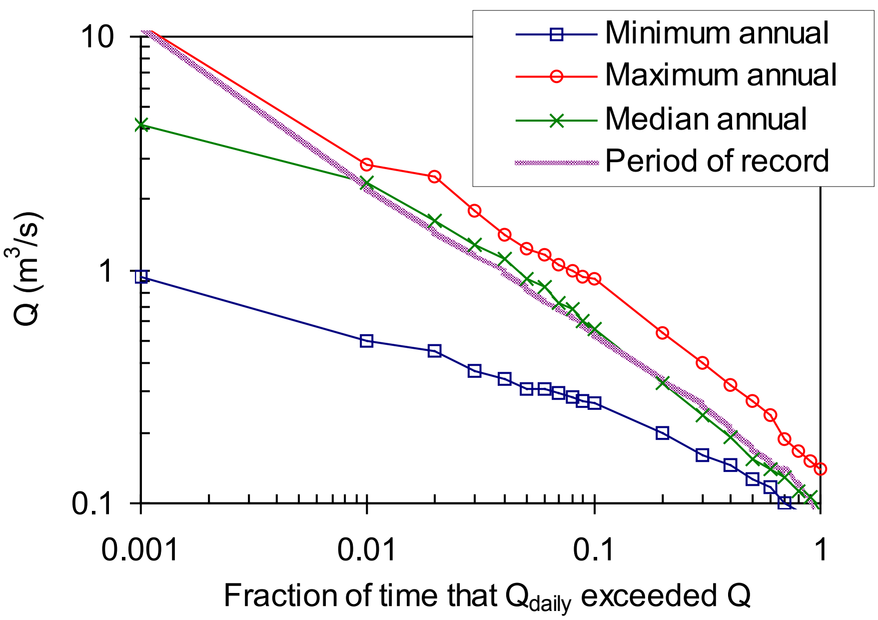
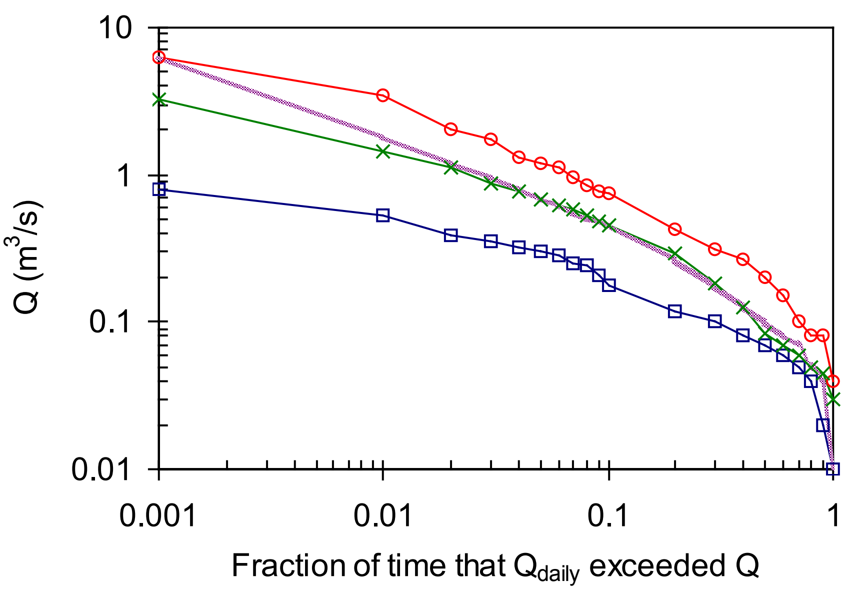



Figure 2.4: Mean discharge rate for August 1994 in 38 Puget Lowland streams as a function of drainage area.

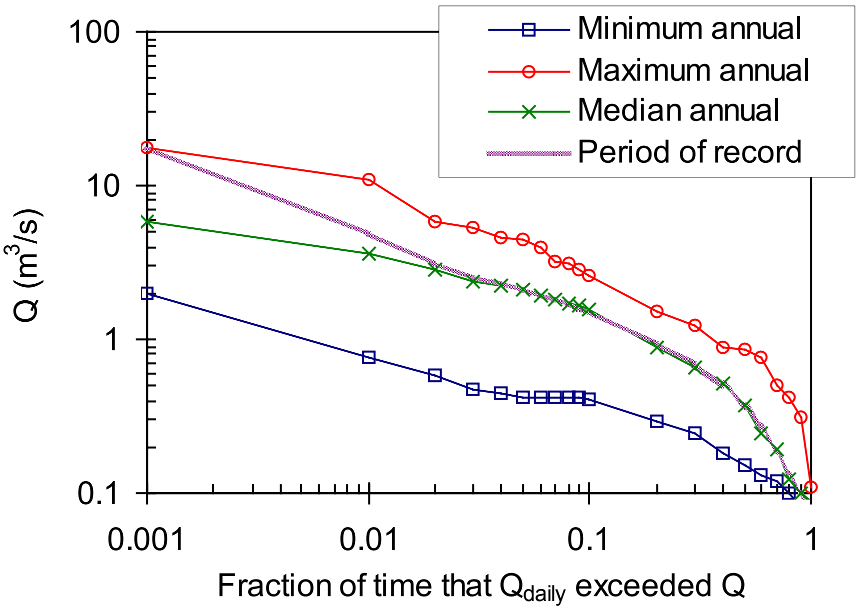



(a) Huge Creek

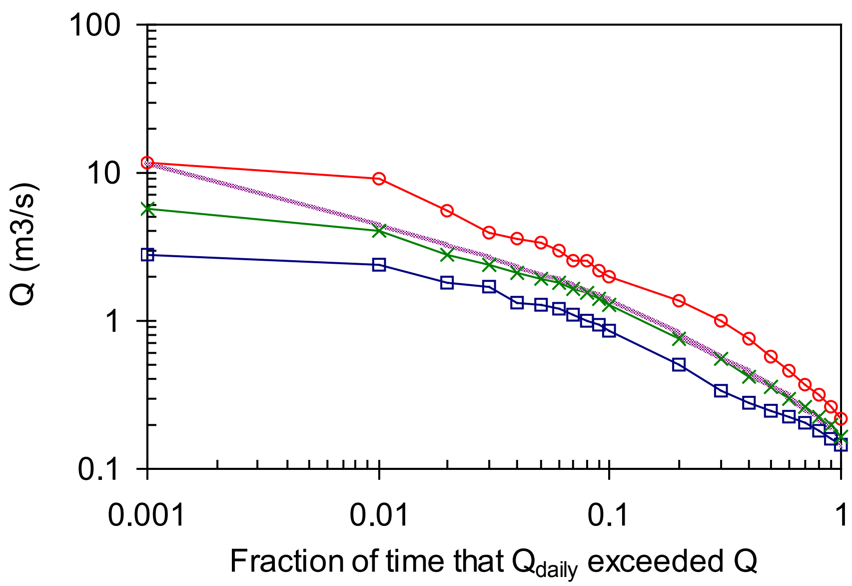



(b) Miller Creek

Figure 2.5: Minimum, median, and maximum annual flow-duration curves showing the fraction of time that mean daily discharge rate ($Q_{daily}$) exceeded the discharge rate (Q) for Huge, May, Mercer, and Miller Creeks during WY1989-1998.




(c) May Creek




(d) Mercer Creek

Figure 2.5 (continued).

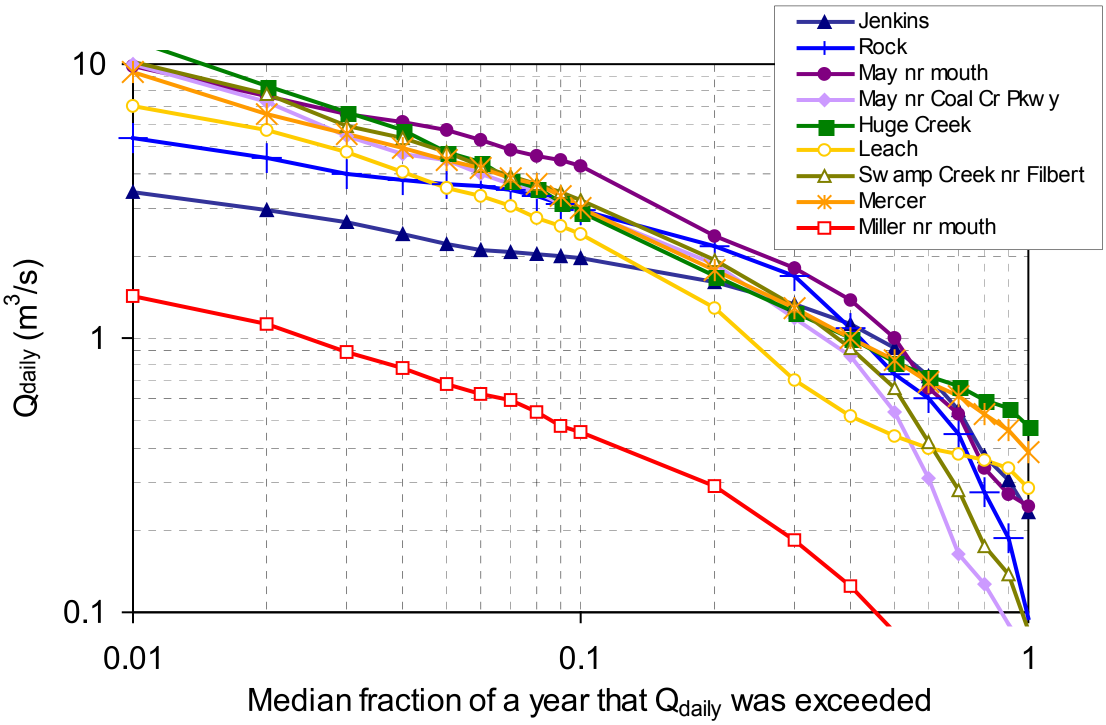



Figure 2.6: Median annual flow duration curves for Puget Lowland streams.

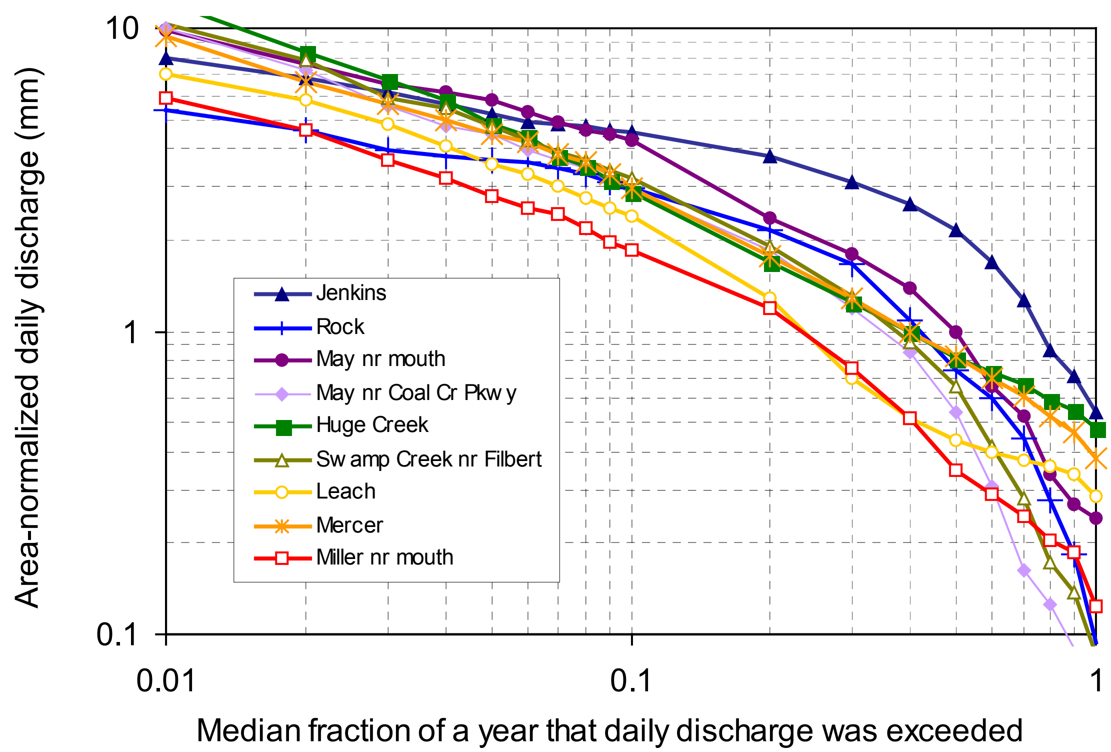



Figure 2.7: Area-normalized, median annual flow duration curves.

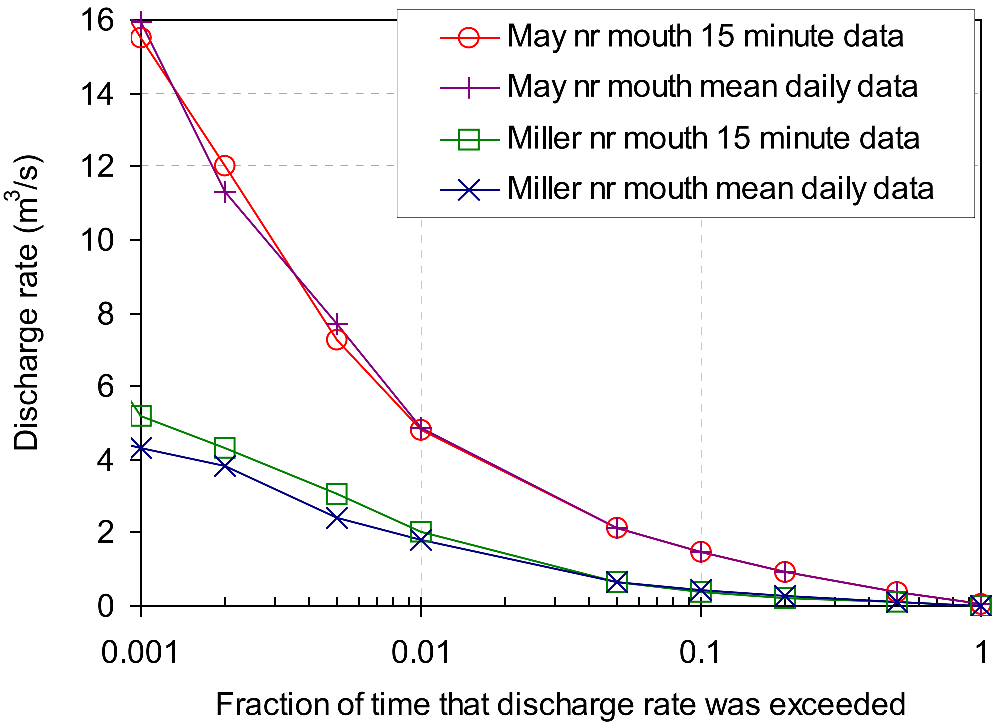



Figure 2.8 Comparison of flow duration curves based on 15 minute and daily mean discharge rate for May and Miller Creeks during WY 1989-1998.

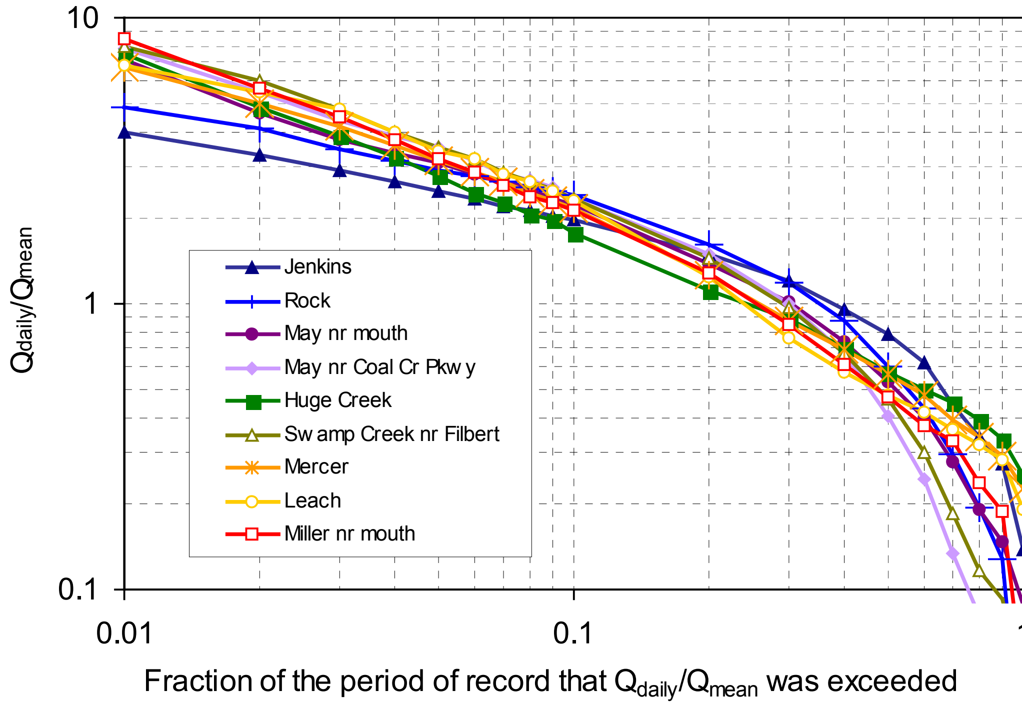



Figure 2.9a: Daily flow duration curves normalized by mean discharge rate.

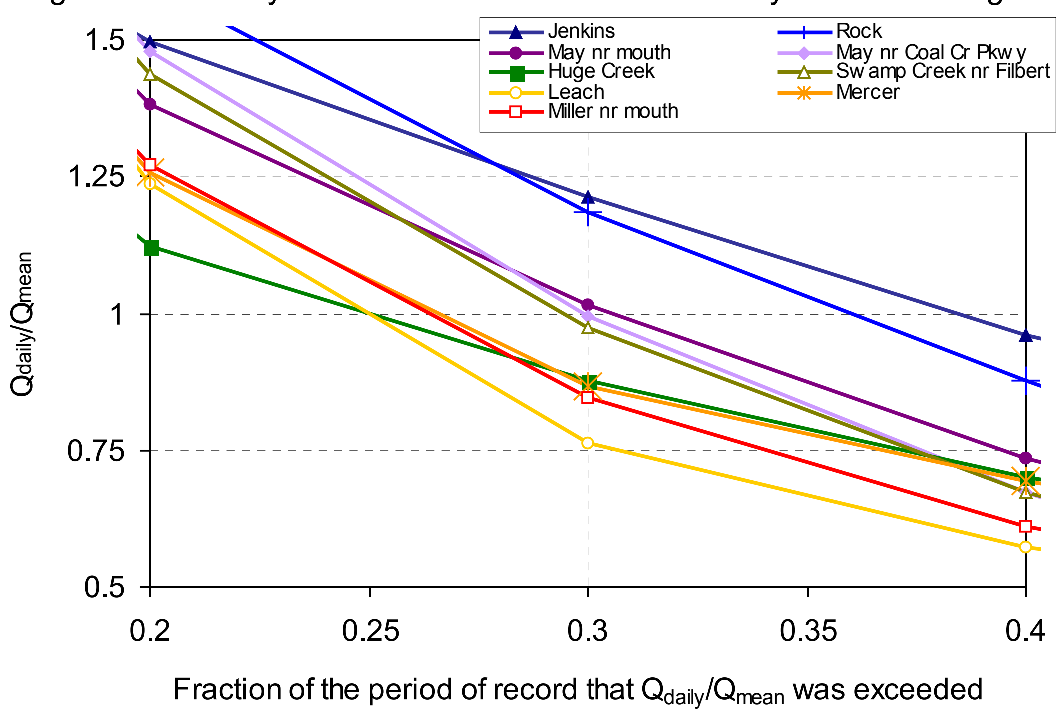



Figure 2.9b: Intermediate quantiles of the daily flow duration curves normalized by mean discharge rate.

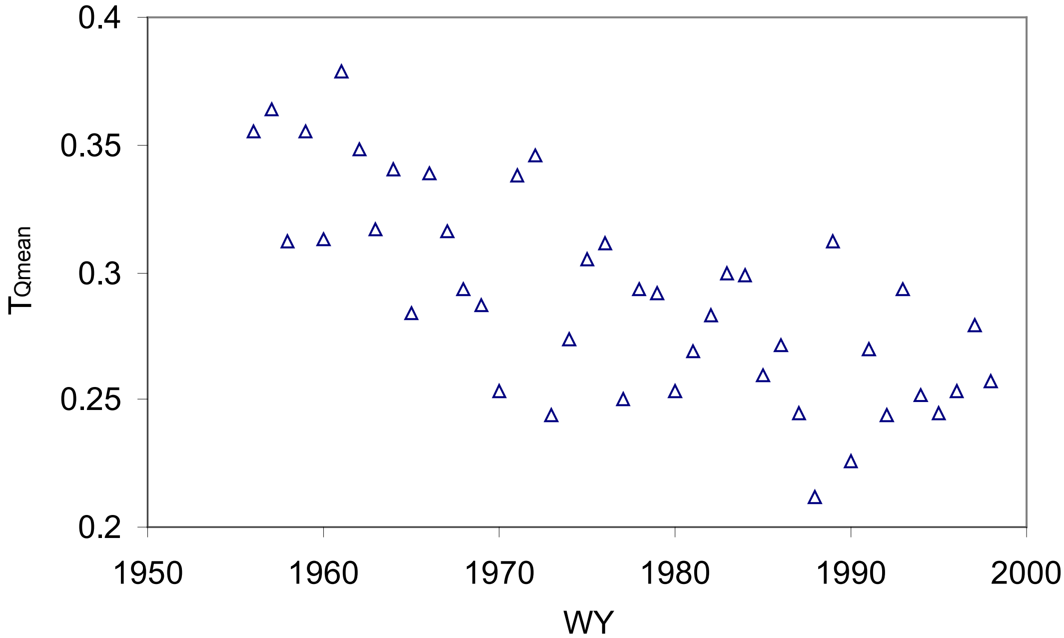



Figure 2.10: Fraction of the year that daily mean discharge exceeded annual mean discharge rate ($T_{Qmean}$) for Mercer Creek for WY 1954 to 1998.

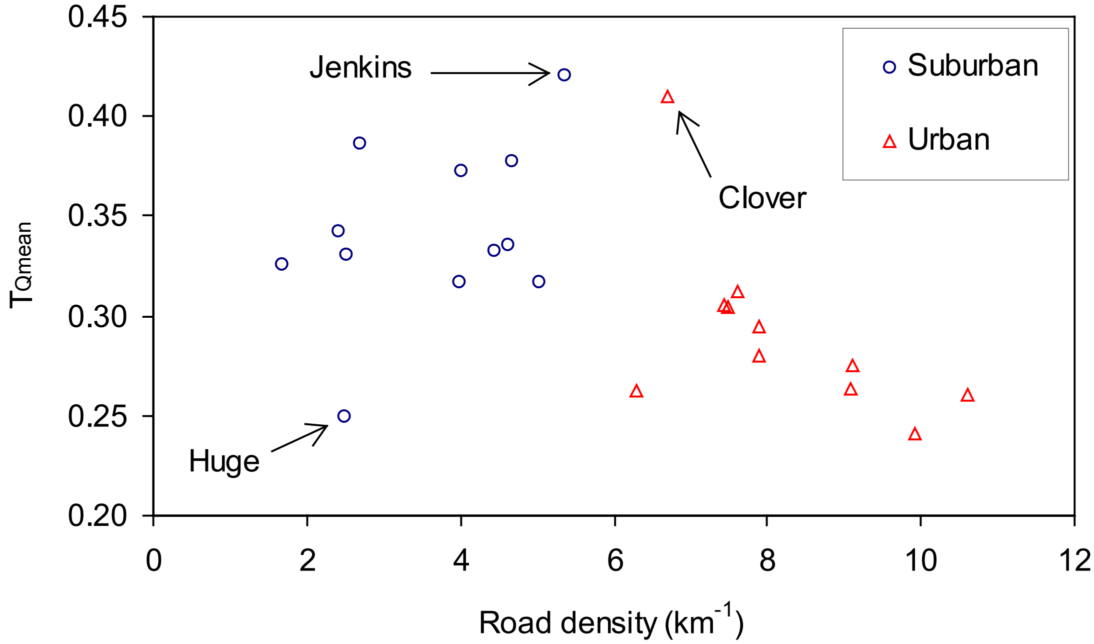



Figure 2.11: Fraction of year that daily mean discharge rate exceeded annual mean discharge rate ($T_{Qmean}$) as a function of road density.

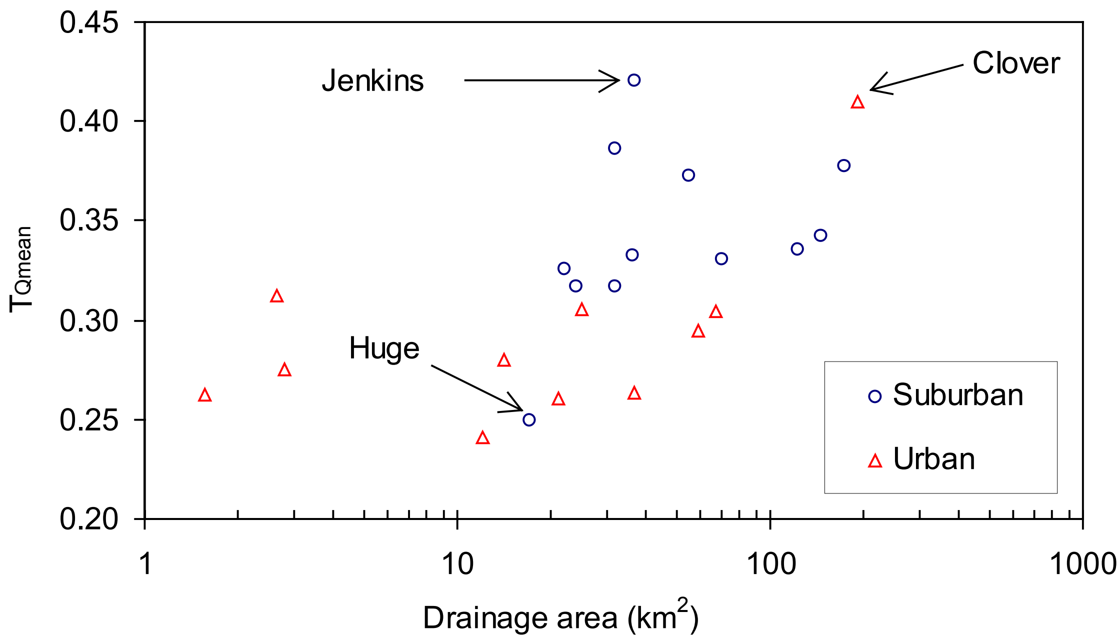



Figure 2.12: Fraction of year that daily mean discharge rate exceeded annual mean discharge rate ($T_{Qmean}$) as a function of drainage area for 18 streams during WY1989-1998.

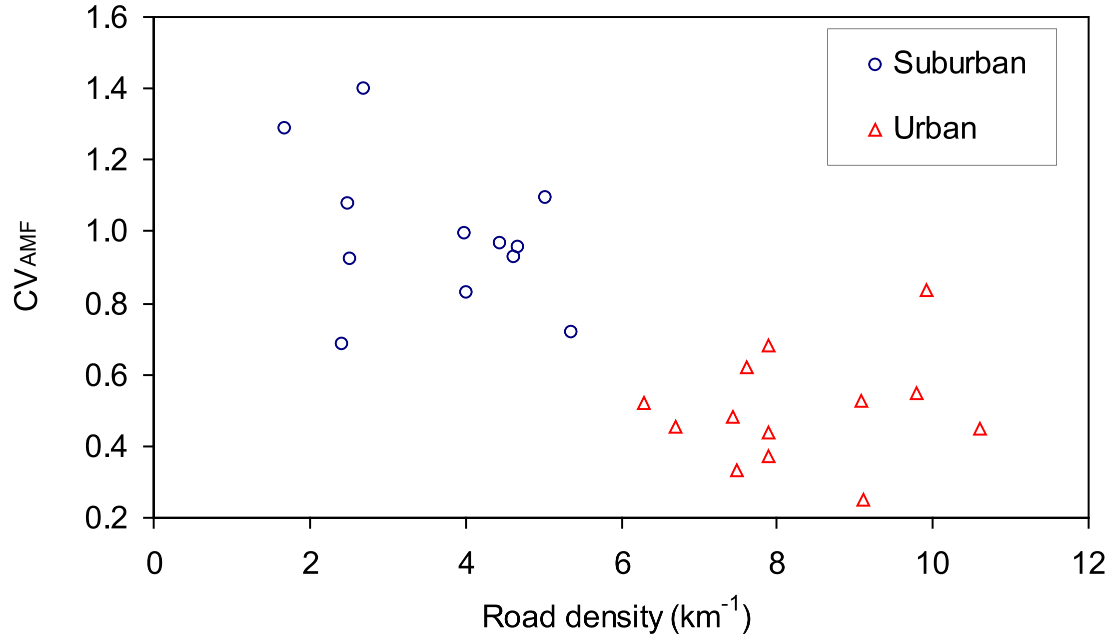



Figure 2.13: Coefficient of variation of annual maximum flood ($CV_{AMF}$)as a function of road density during WY 1989 to 1998.

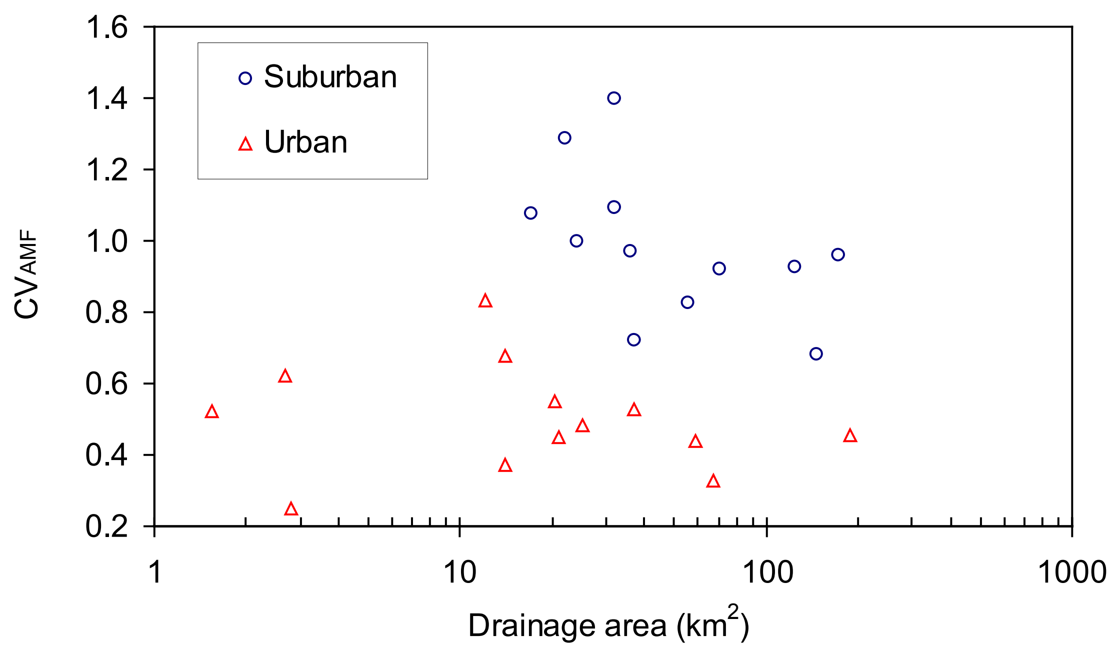



Figure 2.14: Coefficient of variation of annual maximum flood ($CV_{AMF}$) as a function of drainage area during WY 1989 to 1998.

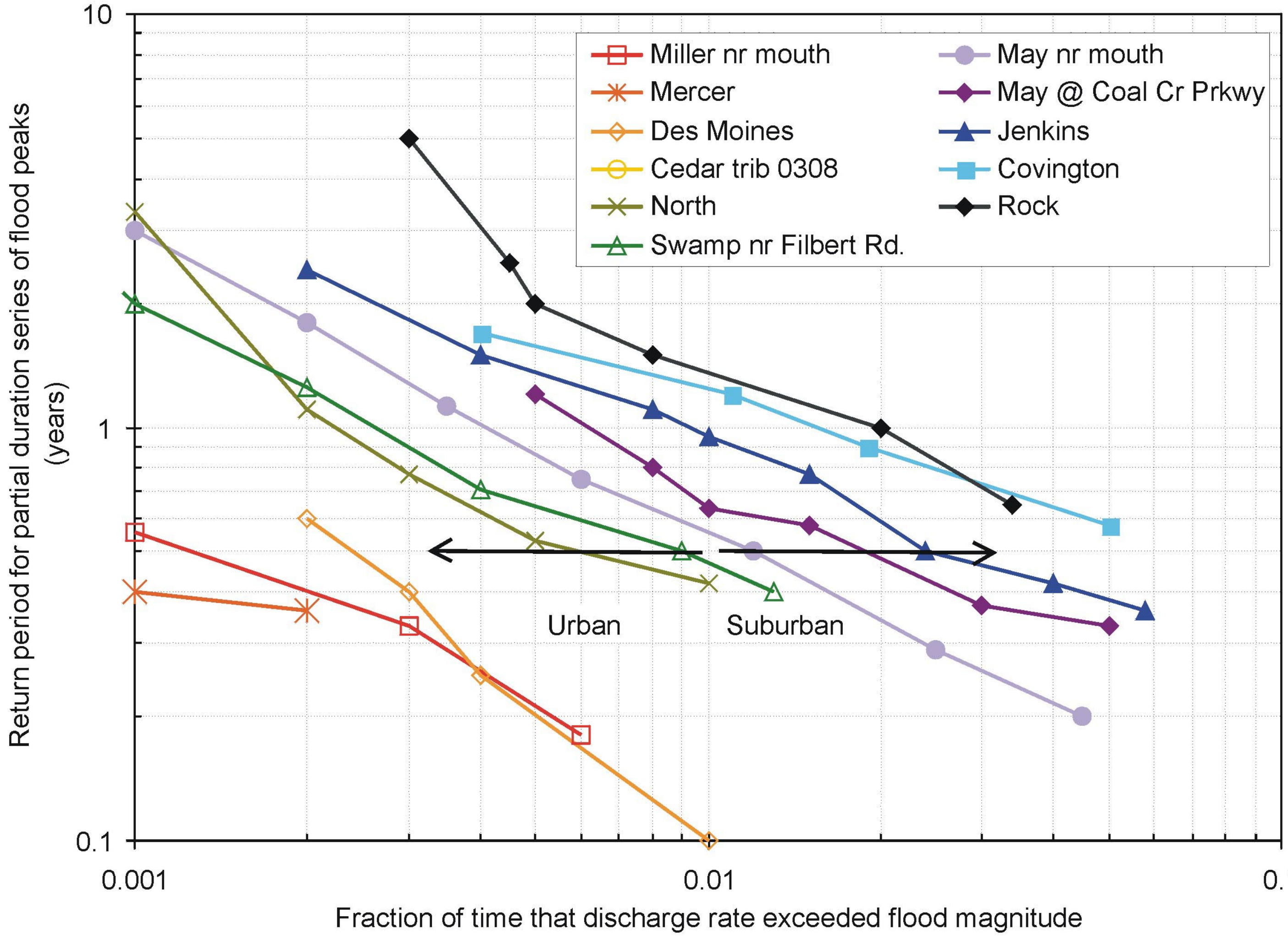



Figure 2.15: Return period and cumulative flow duration for high flows in Puget Lowland streams.

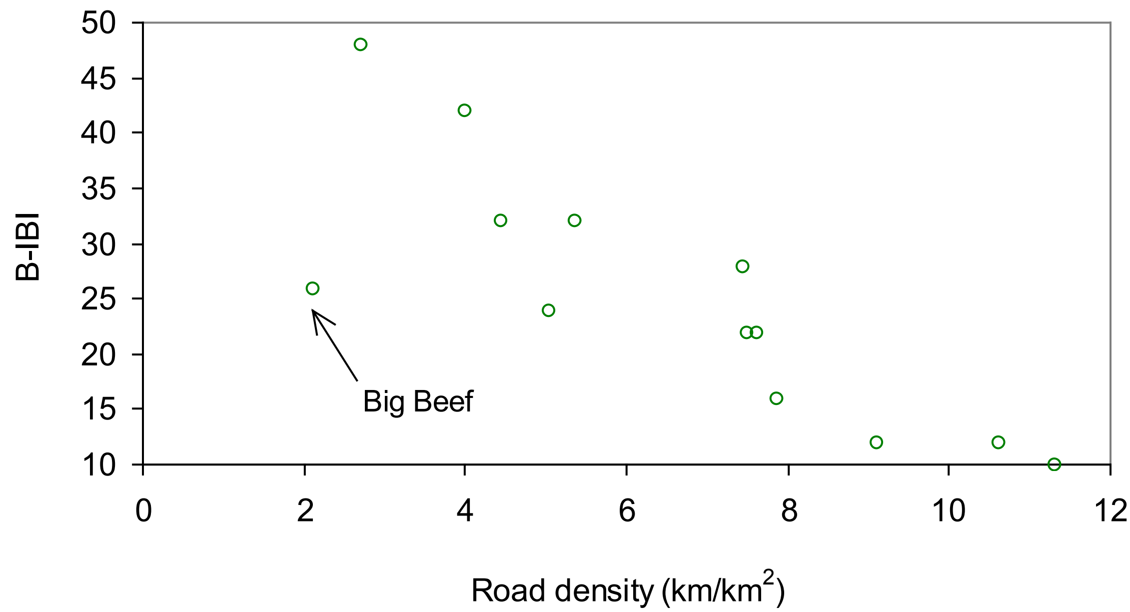



Figure 2:16: Benthic index of biological integrity (B-IBI) plotted against road density for 13 Puget Lowland streams.

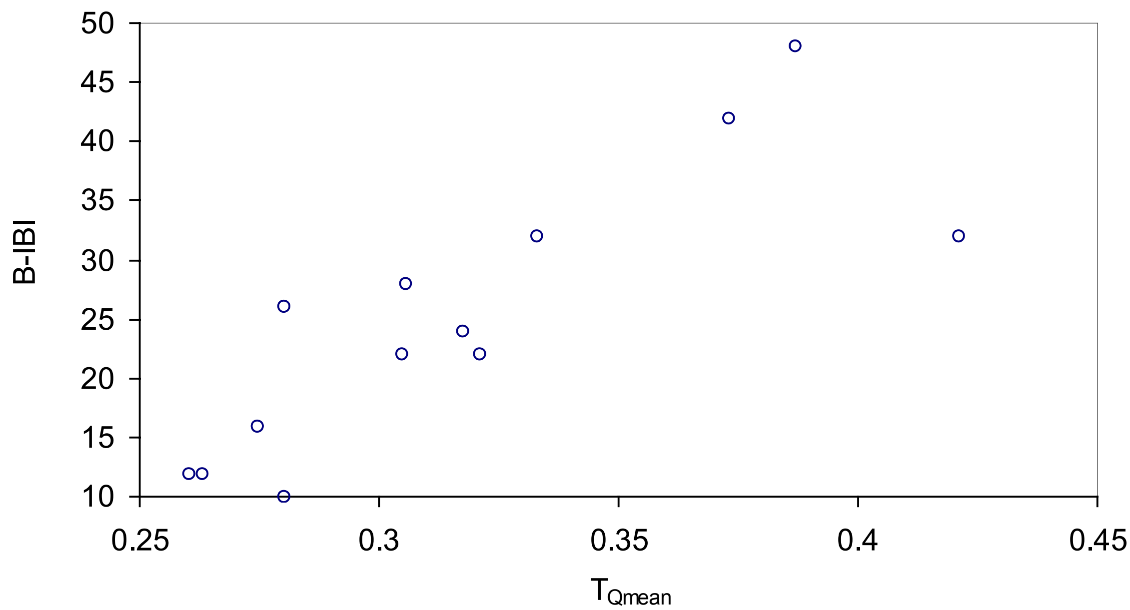



Figure 2:17: Benthic index of biological integrity (B-IBI) plotted against fraction of time that daily mean discharge rate exceeds annual mean discharge rate ($T_{Qmean}$) for 13 Puget Lowland streams.

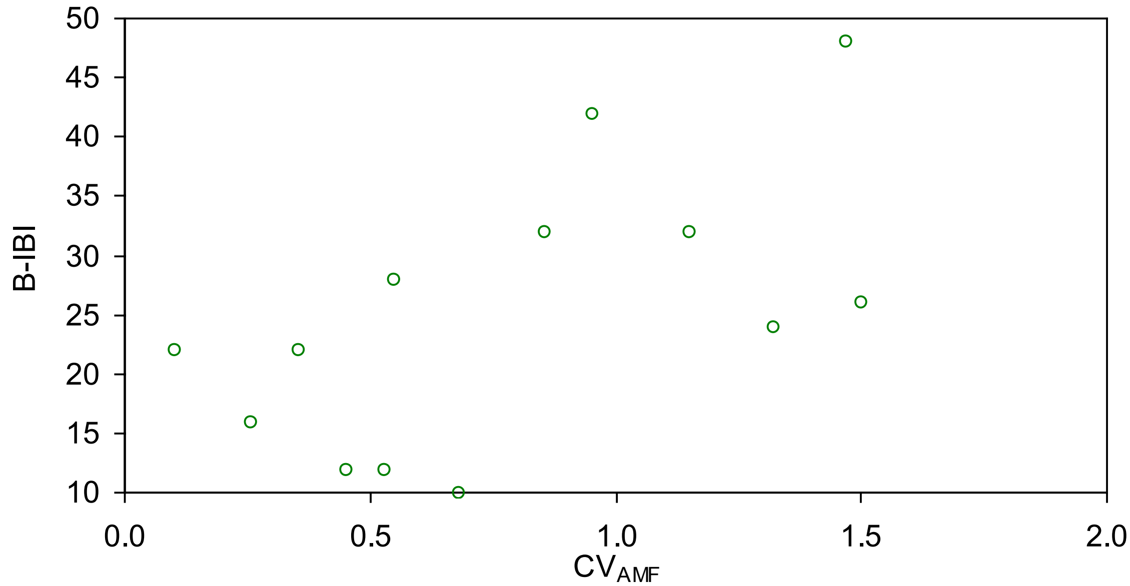



Figure 2.18: Benthic index of biological integrity (B-IBI) plotted against the coefficient of variation of the annual maximum flood ($CV_{AMF}$) for 13 Puget Lowland streams.

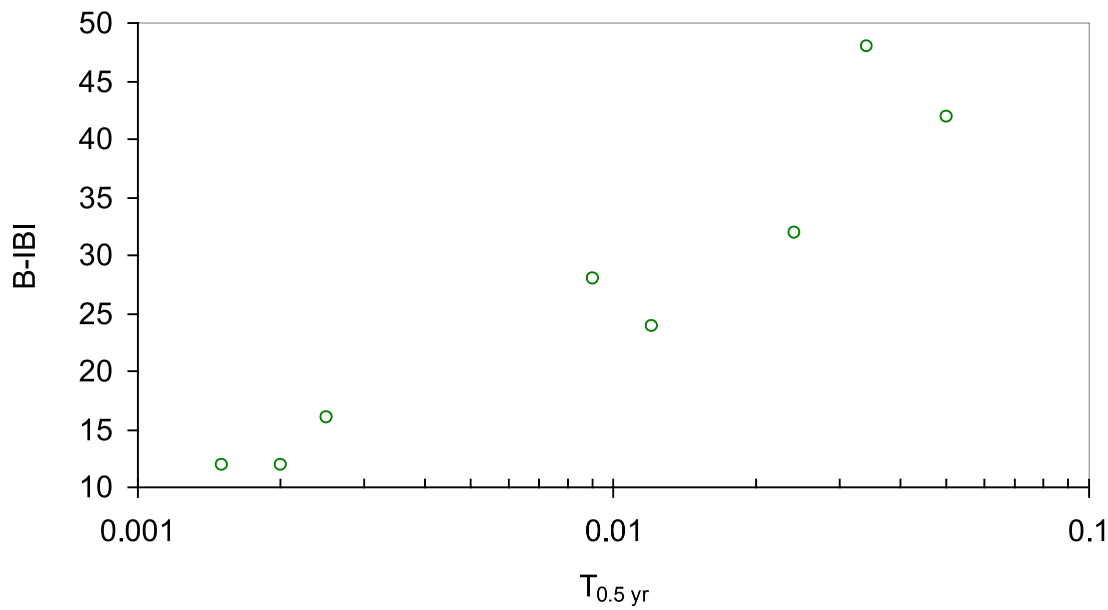



Figure 2.19: Benthic index of biological integrity (B-IBI) plotted against the fraction of the period of record that discharge exceeded the magnitude of a "1/2 yr" flood ($T_{0.5.yr}$) for 8 Puget Lowland streams.

Table 2.1: Drainage area, road density, and operators of stream gages used in hydrologic analysis.

| Suburban streams | Drainage area (km$^2$) | Road density (km/km$^2$) | Stream gage operators |
|---|---|---|---|
| Novelty Hill | 0.37 | 0.0 | King Co. |
| Canyon Creek | 7.0 | 1.3 | King Co. |
| Huge Creek | 17 | 2.5 | USGS |
| East Fork Issaquah Creek | 22 | 1.7 | King Co. |
| May Creek @ Coal Cr. Pkwy. | 24 | 4.0 | King Co. |
| Rock Creek | 32 | 2.7 | USGS/King Co. |
| May Creek near mouth | 32 | 5.0 | USGS/King Co. |
| Big Beef Creek | 35 | 2.1 | USGS |
| Bear Creek @ 133rd Ave. N.E. | 36 | 4.4 | USGS/King Co. |
| Evans Creek | 37 | 3.6 | King Co. |
| Jenkins Creek | 37 | 5.4 | King Co. |
| Covington Creek | 55 | 4.0 | King Co. |
| Newaukum Creek | 70 | 2.6 | USGS |
| Bear Creek @ Union Hill Rd. | 123 | 4.6 | King Co. |
| Issaquah Creek | 145 | 2.4 | USGS |
| Soos Creek | 171 | 4.7 | USGS |

| Urban streams | | | |
|---|---|---|---|
| Miller tributary 031A | 1.6 | 6.3 | King Co. |
| Cedar River tributary 0308 | 2.7 | 7.6 | King Co. |
| Des Moines Creek above Tyee pond | 2.8 | 9.1 | King Co. |
| Leach Creek | 12 | 9.9 | USGS |
| Des Moines Creek near mouth | 14 | 7.9 | King Co. |
| Juanita Creek | 17 | 11.3 | USGS |
| Flett Creek | 20 | 9.8 | USGS |
| Miller Creek near mouth | 21 | 10.6 | King Co. |
| Swamp Creek near Filbert Rd. | 25 | 7.4 | Snohomish Co. |
| Mercer Creek | 37 | 9.1 | USGS |
| Swamp Creek near mouth | 59 | 7.9 | USGS |
| North Creek | 67 | 7.5 | Snohomish Co. |
| Clover Creek | 189 | 6.7 | USGS |

Table 2.2: Comparison of discharge statistiscs for urban and suburban streams during WY1989 to 1998.

| | Years in record | Area-normalized daily mean discharge rate (mm/day) | Fraction of a year that daily mean discharge rate exceeds annual mean discharge rate ($T_{Qmean}$) | |
|---|---|---|---|---|
| Suburban streams | | | Mean of annual $T_{Qmean}$ | Coefficient of variation of annual $T_{Qmean}$ |
| Huge Creek | 10 | 1.5 | 0.25 | 0.16 |
| East Fork Issaquah Creek | 6 | 2.2 | 0.33 | 0.12 |
| May Creek @ Coal Cr. Pkwy. | 6 | 1.3 | 0.32 | 0.09 |
| Rock Creek | 4 | 1.3 | 0.39 | 0.08 |
| May Creek near mouth | 6 | 1.9 | 0.32 | 0.11 |
| Bear Creek @ 133rd Ave. N.E. | 7 | 2.0 | 0.33 | 0.03 |
| Jenkins Creek | 9 | 2.5 | 0.42 | 0.08 |
| Covington Creek | 6 | 1.3 | 0.37 | 0.11 |
| Newaukum Creek | 10 | 2.0 | 0.33 | 0.11 |
| Bear Creek @ Union Hill Rd. | 8 | 1.7 | 0.34 | 0.15 |
| Issaquah Creek | 10 | 2.1 | 0.34 | 0.08 |
| Soos Creek | 10 | 1.8 | 0.38 | 0.06 |
| Average of suburban streams | | 1.8 | 0.34 | |

Table 2.2 (continued)

| Urban streams | Years in record | Area-normalized daily mean discharge rate (mm/day) | Fraction of a year that daily mean discharge rate exceeds annual mean discharge rate ($T_{Qmean}$): Mean of annual $T_{Qmean}$ | Coefficient of variation of annual $T_{Qmean}$ |
|---|---|---|---|---|
| Miller tributary 031A | 5 | 2.3 | 0.26 | 0.47 |
| Cedar River tributary 0308 | 6 | 0.4 | 0.31 | 0.22 |
| Des Moines Creek above Tyee pond | 7 | 1.6 | 0.27 | 0.17 |
| Leach Creek | 10 | 1.1 | 0.24 | 0.11 |
| Juanita Creek | 10[a] | 1.2 | 0.28 | 0.11 |
| Miller Creek near mouth | 7 | 0.8 | 0.26 | 0.15 |
| Swamp Creek @ Filbert Rd. | 10 | 1.4 | 0.31 | 0.07 |
| Mercer Creek | 10 | 1.5 | 0.26 | 0.09 |
| Swamp Creek near mouth | 10[a] | 1.4 | 0.30 | 0.10 |
| North Creek | 9 | 1.3 | 0.30 | 0.19 |
| Clover Creek | 6 | 0.6 | 0.41 | 0.12 |
| Average of urban streams | | 1.2 | 0.29 | |

[a] WY 1980-89

Table 2.2 (continued)

| | | Statistics of annnual maximum flood distributions (instantaneous or 15-minute mean discharge rate) | | |
|---|---|---|---|---|
| Suburban streams | Years in record | Geometric mean discharge ($m^3/s$) | Area-normalized geometric mean discharge ($m^3/s/km^2$) | Coefficient of variation - (2 parameter lognormal distribution) |
| Huge Creek | 10 | 4.5 | 0.27 | 1.08 |
| East Fork Issaquah Creek | 6 | 7.8 | 0.35 | 1.29 |
| May Creek @ Coal Cr. Pkwy. | 7 | 6.0 | 0.25 | 1.00 |
| Rock Creek | 4 | 5.2 | 0.16 | 1.40 |
| May Creek near mouth | 10 | 7.7 | 0.24 | 1.09 |
| Bear Creek @ 133rd Ave. N.E. | 9 | 5.0 | 0.14 | 0.97 |
| Jenkins Creek | 11 | 5.0 | 0.13 | 0.72 |
| Covington Creek | 7 | 4.9 | 0.09 | 0.83 |
| Newaukum Creek | 10 | 19.8 | 0.28 | 0.92 |
| Bear Creek @ Union Hill Rd. | 10 | 15.2 | 0.12 | 0.93 |
| Issaquah Creek | 10 | 38.6 | 0.27 | 0.69 |
| Soos Creek | 10 | 24.5 | 0.14 | 0.96 |
| Average of suburban streams | | | 0.20 | 0.99 |

Table 2.2 (continued)

| Urban streams | Years in record | Statistics of annnual maximum flood distributions (instantaneous or 15-minute mean discharge rate) Geometric mean discharge ($m^3/s$) | Area-normalized geometric mean discharge ($m^3/s/km^2$) | Coefficient of variation - (2 parameter lognormal distribution) |
|---|---|---|---|---|
| Miller tributary 031A | 4 | 0.5 | 0.35 | 0.52 |
| Cedar River tributary 0308 | 7 | 0.7 | 0.25 | 0.62 |
| Des Moines Creek above Tyee pond | 7 | 3.7 | 1.33 | 0.25 |
| Leach Creek | 10 | 2.9 | 0.24 | 0.83 |
| Juanita Creek | 10[a] | 6.1 | 0.44 | 0.68 |
| Des Moines Creek near mouth | 7 | 4.4 | 0.31 | 0.37 |
| Flett Creek | 10 | 2.2 | 0.11 | 0.55 |
| Miller Creek near mouth | 9 | 7.5 | 0.35 | 0.45 |
| Swamp Creek @ Filbert Rd. | 10 | 6.0 | 0.24 | 0.48 |
| Mercer Creek | 10 | 10.3 | 0.28 | 0.53 |
| Swamp Creek near mouth | 10[a] | 14.5 | 0.25 | 0.44 |
| North Creek | 10 | 17.4 | 0.17 | 0.33 |
| Clover Creek | 10 | 7.9 | 0.04 | 0.45 |
| Average of urban streams | | | 0.34 | 0.50 |

[a] WY 1980-89

TABLE 2.3: Comparison across time of fraction of time that daily mean discharge exceeds annual mean discharge ($T_{Qmean}$).

| | Period A: Water Years 1960-69 | | | Period B: Water Years 1989-1998 | | | Rank sum test of annual values in each period | | |
|---|---|---|---|---|---|---|---|---|---|
| | Number of years | Average of annual values of $T_{Qmean}$ for period | Coefficient of variation | Number of years | Average of annual values of $T_{Qmean}$ for period | Coefficient of variation | Critical value for rank sum for Period A ($p$ =0.05) | Rank sum of annual values of $T_{Qmean}$ for Period A | Rank sum of annual values of $T_{Qmean}$ for Period B |
| Urban streams | | | | | | | | | |
| **Juanita** | **6** | **0.34** | **0.13** | **10**[a] | **0.28** | **0.11** | **>67** | **70** | **64** |
| **Leach** | **8** | **0.25** | **0.11** | **10** | **0.24** | **0.11** | **>95** | **97** | **74** |
| **Mercer** | **10** | **0.32** | **0.09** | **10** | **0.26** | **0.09** | **>127** | **149** | **60** |
| Swamp near mouth | 6 | 0.31 | 0.10 | 10[a] | 0.30 | 0.10 | >61 | 56 | 80 |
| Clover | 5[b] | 0.41 | 0.04 | 6 | 0.41 | 0.12 | >39 | 27 | 38 |
| Suburban streams | | | | | | | | | |
| May near mouth | 7 | 0.33 | 0.10 | 6 | 0.32 | 0.11 | >71 | 55 | 36 |
| Soos | 9 | 0.38 | 0.11 | 10 | 0.38 | 0.06 | >110 | 89 | 98 |
| **Newaukum** | **10** | **0.36** | **0.10** | **10** | **0.33** | **0.11** | **>127** | **129** | **80** |
| Huge | 10 | 0.27 | 0.14 | 10 | 0.25 | 0.16 | >127 | 117 | 91 |
| Issaquah | 6 | 0.36 | 0.10 | 10 | 0.34 | 0.08 | >66 | 59 | 75 |

Bold type indicates streams where annual values of $T_{Qmean}$ were likely to have been higher during Period A than Period B

[a] WY 1980-89

[b] WY 1950-54

TABLE 2.4: Comparison of annual maximum flood distributions over time in urban and suburban streams.

| | Period A: Water Years 1960 - 1969 | | | | Period B: Water Year 1989 - 1998 | | | |
|---|---|---|---|---|---|---|---|---|
| | Number of years in record | Geometric mean annual flood ($m^3/s$) | Area normalized geometric mean annual flood ($m^3/s/km^2$) | Coefficient of variation | Number of years in record | Geometric mean annual flood ($m^3/s$) | Area normalized geometric mean annual flood ($m^3/s$) | Coefficient of variation |
| **Urban streams** | | | | | | | | |
| Leach | 10 | 1.5 | 0.12 | 0.19 | 10 | 2.9 | 0.24 | 0.83 |
| Flett | 10 | 1.1 | 0.05 | 0.39 | 10 | 2.2 | 0.11 | 0.55 |
| Juanita | 6 | 2.7 | 0.16 | 0.13 | 10[a] | 6.1 | 0.35 | 0.68 |
| Mercer | 10 | 5.6 | 0.15 | 0.17 | 10 | 10.3 | 0.28 | 0.53 |
| Swamp | 6 | 11.7 | 0.20 | 0.21 | 10[a] | 14.5 | 0.25 | 0.44 |
| Clover | 10 | 4.5 | 0.02 | 0.46 | 4 | 7.9 | 0.04 | 0.45 |
| **Suburban streams** | | | | | | | | |
| Huge | 10 | 3.6 | 0.21 | 0.83 | 10 | 4.5 | 0.27 | 1.08 |
| Rock | 10 | 2.0 | 0.06 | 0.82 | 4 | 5.2 | 0.16 | 1.40 |
| May | 5 | 6.8 | 0.21 | 0.28 | 10 | 7.7 | 0.24 | 1.09 |
| Newaukum | 10 | 19.3 | 0.28 | 0.30 | 10 | 19.8 | 0.28 | 0.92 |
| Issaquah | 6 | 40.6 | 0.28 | 0.33 | 10 | 38.6 | 0.27 | 0.69 |
| Soos | 9 | 18.3 | 0.11 | 0.51 | 10 | 24.5 | 0.14 | 0.96 |

[a] WY 1980 - 1989

# Chapter 3: Spatial extent of stream networks during summer base flow conditions

The availability of aquatic habitat in a stream basin at any time depends on the spatial extent of surface flow in the channel network. In the Puget Lowland, the area and length of streams expand during the winter and contract during the summer. The extent of surface flow in a channel network during the dry season defines the transition between the downstream limit of ephemeral reaches, which have no surface flow for a portion of the year, and the upstream limit of perennial reaches, which flow all year.

The seasonal desiccation of ephemeral streams represents a periodic form of disturbance that does not occur regularly in perennially flowing streams. As a consequence of the annual disturbance, lotic communities in ephemeral streams are less diverse and complex than those inhabiting perennial streams (Stehr and Branson, 1938; Rabeni and Wallace, 1998) and comprise organisms adapted to the occasional lack of surface flow (McAuliffe, 1984), for example, by migrating vertically into the hyporheic zone (Clinton et al., 1996).

Hydrologic changes that reduce groundwater elevations or subsurface flow near headwater reaches could potentially reduce the spatial extent of stream flow in a channel network. In this chapter, I examine whether the spatial extent of perennial streams is greater in suburban areas than in urban areas of the Puget Lowland using two measures: the drainage area of streams at the transitions from ephemeral to perennial flow and the density of perennial streams in a basin. These measures reflect the assumption that drainage area is the primary factor controlling perennial flow in stream channels.

Structural changes in channel networks associated with urban development (e.g., filling small channels, extending a drainage network with ditches, replacing a stream channel with underground pipes) change the extent of a drainage network in a stream basin. While the total length of channel, pipes, and other drainage features typically increases (Graf, 1977), the extent of natural stream channel decreases (e.g., Figure 18-3 in Dunne and Leopold, 1978). Small, ephemeral streams are particularly affected by structural changes in urban areas. Nonetheless, larger, perennial streams are confined by long culverts in urban areas of the Puget Lowland such as Thornton Creek at the Northgate Shopping Center and Longfellow Creek at both its headwaters and mouth in West Seattle.

## 3.1. The dynamics of stream networks

The extent of surface flow in a stream network is dynamic, expanding up hillslopes during storms and contracting down the channel network during dry seasons (Gregory and Walling, 1968; Dunne and Black, 1970; Day, 1978; Gardiner, 1995; Montgomery and Dietrich, 1995). Gregory and Walling (1968) related the discharge rate (Q) at a point in the channel network to an exponential function of the total length of stream channel with surface flow (L) upstream of that point:

$$Q = a\, L^{b} \tag{3.1}$$

where a and b are parameters that must be calibrated for the physiographic (e.g., topography, lithology, vegetation) conditions in the catchment. In the two catchments examined by Gregory and Walling, a varied from 8.3 to 9.7 and b varied from 2.1 to 2.9 where Q was measured in cubic feet per second and L was measured in feet.

Equation 3.1 provides an analytical framework for assessing how the hydrologic effects of urban development may influence the spatial extent of surface flow in stream network. An urban stream is likely to have a higher discharge rate during a small storm than it would have if its basin were covered by forest. According to Equation 3.1, the higher discharge rate will be associated with a greater stream length in the basin. This effect is plausible especially since ditches and other drainage structures are likely to be flowing and their combined length is appreciable relative to that of a natural stream network (Graf, 1977). After a storm, the length of stream channel with surface flow is likely to contract rapidly in an urban basin as stream flow declines rapidly and smaller channels (i.e., ditches) stop flowing. During dry-season base flow conditions, however, no general difference in summer base flow was discernable between urban and suburban streams (Figures 2.3 and 2.4). If there is no change in the dry-season base flow in a stream as a result of urban development, then the extent of perennial streams would not change according to Equation 3.1.

These assessments of the effects of urban development on perennial stream lengths, however, are limited by the assumption that the parameters a and b in Equation 3.1 are unaffected by urban development. However, Gregory and Walling (1968) suggest that land use influences the values of these parameters. Furthermore, differences in the values of a and b between basins allow for variation in stream length between basins with similar discharge rates.

## 3.2. Methods for analyzing the spatial extent of stream networks

The investigation of base flow stream networks has two components: field observation and spatial analysis using a geographic information system. I observed the status of stream flow (flowing, intermittently flowing, not flowing) at 136 points along streams in the Puget Lowland, Washington. I include the results of an additional 43 point observations from the stream temperature surveys coordinated by the Center for Urban Water Resources Management on 19 August 1998 and 4 August 1999.

All observations were made after an extended period without rain when stream flow was dominated by steady ground water discharge rather than storm flow. 163 (91%) of the observations were made between 16 August and 18 September 1998. During this period, discharge was steady in most streams and the extent of surface flow in a stream network was unlikely to vary much from day to day. As 1998 was a relatively dry year, the reaches observed with surface flow are likely to flow perennially.

16 (9%) of the observations were made on 4 August 1999. Water Year 1999 was much wetter in the Puget Lowland region in contrast to 1998. As a result, stream flow was still receding during early August 1999 and there were points along channel networks where flowing water was observed during August 1999 that had been dry in August 1998. Surface flow was observed at 10 of the 16 points observed in 1999. These observations are likely to biased relative to the 1998 observations indicating surface flow higher (upstream) in the channel network. The 6 points observed to be dry in 1999 are likely to have been dry during August 1998.

Field observations and other geographic data were analyzed in the geographic information system ArcView© (ESRI, 1999). The points of observation, streams, and watersheds were digitized from USGS 7.5 minute quadrangles. Each stream was delineated from its mouth to the highest point along the mapped (on the USGS quadrangle) channel where surface flow was observed. Surface flow may have been present in stream channels some (unknown) distance upstream of the furthest upstream point of observation. As a result, the delineated streams may under-represent the extent of the perennial stream network in a basin.

Watersheds were delineated for the 179 observation points by digitizing a polygon around the area upstream of each observation point with its boundary orthogonal to the mapped contour lines (5 m interval). Generally, there is only a single, unbifurcated channel within each of these watersheds, so the drainage areas associated with these observation points are hence forward referred to as first-order stream basins. The locations of the 179 first-order basins are shown in Figure 3.1.

The watersheds for 52 larger, perennial streams were delineated in the same manner and are also shown in Figure 3.1. These watersheds comprise second and third order streams (Strahler, 1952) and are referred to as higher order stream basins. Table 3.1 provides the name of each higher order basin and the number of its first-order basins included in the analysis. Stream lengths in the higher order stream basins and drainage areas for all stream basins were calculated using "ReturnLength" and "ReturnArea" algorthims in ArcView.

Drainage area is posited as the primary independent variable for both the locations of the transition between perennial and ephemeral flow in first-order basins and the total length of perennial streams in higher order basins. The drainage areas of first-order urban basins are compared with drainage areas of first-order suburban basins where the basins have been stratified by flow status (i.e., flow observed, no flow observed). The likelihood that surface flow will be observed in a stream channel increases in the downstream direction.

The conditional probability of observing surface flow at a point along a channel with a drainage area of A was calculated according to Bayes theorem (Hoel, 1971):

$$P\langle Q>0 \mid A\rangle = \frac{P\langle Q>0\rangle\, P\langle A \mid Q>0\rangle}{P\langle Q>0\rangle\, P\langle A \mid Q>0\rangle + P\langle Q=0\rangle\, P\langle A \mid Q=0\rangle} \qquad (3.2)$$

where $Q > 0$ represents surface flow, $Q = 0$ represents no flow, $P(A \mid Q>0)$ is the cumulative distribution function (CDF) of drainage areas for observation points where there was flow, $P(A \mid Q=0)$ is the complement of the CDF of drainage areas for observation points where there was no surface flow. Reaches observed to have intermittent flow were excluded from this analysis.

A general relationship between perennial stream length and drainage area is developed for the higher order streams basins. Perennial stream length ($L_{stream}$) is the total length of stream channel with surface flow during the summer low flow period in a stream basin. Stream densities ($D_{stream}$) are calculated for the higher order stream basins as:

$$D_{stream} = \frac{L_{stream}}{A} \tag{3.3}$$

This usage of stream density is distinct from the "drainage density" *in sensu* Horton (1945) and Carlston (1963), where channel length is the numerator. It is, however, comparable to Gregory and Wallings (1968) usage of drainage density. Deviations from the general stream density relationship are compared with topographic characteristics and level of development in the higher order basins.

The effects of land use on the extent of stream networks are analyzed in 29 higher order stream basins. The density of roads serves as an indicator of the level of urban development in a stream basin. Georeferenced line drawings of roads ca. 1990 were obtained from King, Kitsap, Pierce, and Snohomish Counties and incorporated into the GIS. The road data do not include sidewalks, parking lots, driveways, or service roads. Road densities ($D_{road}$) were calculated as the total length of roads in each basin ($L_{road}$) divided by the basin's drainage area:

$$D_{road} = \frac{L_{road}}{A} \tag{3.4}$$

Higher order stream basins with road densities greater than 6 km/km$^2$, and all of their first-order basins, are classified as urban. Those higher order stream basins with road densities less than 6 km/km$^2$, and all of their first-order basins, are classified as suburban. The division of streams as urban and suburban may obscure any incremental effects of urban development.

Other physiographic conditions besides drainage area are expected to influence the spatial extent of stream flow at the scales of both first-order and higher order stream basins. The influence of basin width, basin length, valley relief, and valley slope are analyzed using a group of 40 first-order stream basins and the 29 higher order stream basins where road densities were calculated. Basin length ($L_{basin}$) was measured along a stream basin's longitudinal axis that extends from its outlet along the valley bottom to the watershed. An example using Shinglemill Creek is shown in Figure 3.2. Basin width was calculated as:

$$w_{basin} = \frac{A}{L_{basin}} \tag{3.5}$$

Valley relief and slope were calculated using USGS digital elevation models (DEMs) with 10 m horizontal resolution. The DEMs were generated from 7.5 minute quadrangles with elevation contour lines at 12 m (40 ft) intervals. The intersection of a stream basin's longitudinal axis and the watershed defines the point of the maximum valley elevation (see Figure 3.2). Valley relief is defined as the difference between the elevation of the basin outlet ($z_{outlet}$) and maximum valley elevation ($z_{max}$). The valley slope of each basin was calculated as:

$$S_{valley} = \frac{z_{max} - z_{outlet}}{L_{basin}} \tag{3.6}$$

## 3.3. Results

### 3.3.1. Drainage area of first-order streams

The flow status of 151 points in first-order streams were observed under summer base flow conditions: 96 points had surface flow (i.e., $P(Q>0) = 0.64$) and 55 points had no flow (i.e., $P(Q=0) = 0.36$). The conditional (i.e., based on drainage area) probability of observing surface flow was calculated using Bayes theorem (Equation 3.2) and is shown for all streams in Figure 3.3. There was a 50% chance of observing flow at a point in a channel network where the drainage area is 1.2 $km^2$. The probability of observing surface flow for a point where the drainage area was 1.2 $km^2$ was slightly higher (63%) for urban streams and slightly lower (40%) for suburban streams.

The observations of flow and no flow were widely distributed with respect to drainage areas for both urban and suburban streams. The smallest drainage area at a location of perennial flow was 0.02 $km^2$ (Fauntleroy) for the urban streams and 0.07 $km^2$ (Fisher) for suburban streams. The largest drainage area at a location of no flow was 11.8 $km^2$ (Swamp) for urban streams and 8.3 $km^2$ (Soos) for suburban streams.

The physiographic conditions of stream basins analyzed here do not account for the wide and overlapping ranges of drainage areas for ephemeral and perennial streams. Figure 3.4 shows the values of four physiographic characteristics plotted against drainage area for 40 first-order streams: (a) valley slope; (b) basin width; (c) basin length; (d) valley relief. The values for first-order basins with surface flow during summer are plotted as triangles while the values for those basins that were dry are plotted as circles. In Figure 3.4, symbols for first-order urban streams are filled while symbols for suburban streams are unfilled. In general, the topographic factors, in combination with drainage area, do not resolve differences between flowing (triangles) and dry (circles) streams.

### 3.3.2. Higher order basins

The length of perennial streams ($L_{stream}$) was analyzed in 52 higher order stream basins (Figure 3.5). Perennial streams length, measured in units of km, can be approximated as a linear function of drainage area, measured in units of $km^2$:

$$L_{stream} = 0.4A + 0.8 \tag{3.7}$$

Equation 3.7 over-estimated perennial stream lengths on average and had a root mean square percentage error of 9.8% relative to observed stream lengths.

The average value of $D_{stream}$ is 0.53 $km^{-1}$ with a standard deviation of 0.32 $km^{-1}$ for all basins. The root mean square error of estimates of $D_{stream}$ based on Equation 3.7 is 0.056 $km^{-1}$. No relationship is evident between drainage density, $D_{stream}$, and road density, $D_{road}$, for these higher order stream basins (Figure 3.6) even with the possible spurious correlation between $D_{stream}$ and $D_{road}$, which share a common denominator, drainage area (Betson 1965). The difference between $D_{stream}$ of urban basins (0.56 $km^{-1}$) and suburban basins (0.46 $km^{-1}$) is not significant (Student's t-test of sample with unequal variance). The average deviation between the predicted (by Equation 3.7) and observed values of $D_{stream}$ for each stream is not related to the road density of the stream (Figure 3.7a).

Stream densities in higher order stream basins do not vary with any of the three physiographic factors considered here: basin shape, valley relief, and valley slope (Figures 3.7b, c, and d). The only evident pattern is that $D_{stream}$ is higher than expected in basins with low valley slopes (< 0.006).

## 3.4. The spatial extent of stream networks during summer base flow conditions

There is not a well-defined area-based threshold for perennial surface flow among the first-order Puget Lowland stream basins analyzed here. Stream basins less than 1 $km^2$ can generate summer base flow. Likewise, streams with drainage areas as large as 10 $km^2$ may be ephemeral. Urban development and the physiographic conditions evaluated here appear to have little influence on distribution of perennially flowing streams in the basins examined here. This conclusion parallels the lack of a difference in area-normalized summer base flow between urban and suburban streams shown in Chapter 2.

The extent of perennial flow in stream networks may be related to other physiographic conditions of the stream basins, such as their surficial geology and, in particular, the location of aquifers relative to the land surface. For example, perennial flow is likely to occur where the land surface converges on a confined aquifer, such as at springs along a valley wall. While the discharge from such springs may vary seasonally, their location will be stable. Alternatively, the extent of surface flow in a channel may vary over time with groundwater elevations in an unconfined alluvial aquifer. In this case, the local depth and width of alluvium in combination with recharge rates of the aquifer may determine the location of surface flow in stream channel. In such cases of local control on the location of the surface flow in a stream channel, the influence of urban development may be limited.

## 3.5. The hydroperiod of ephemeral streams

While there was no observed effect of urban development on the extent of perennial streams in a basin, the hydroperiod (i.e., the period of time during a year when surface water is present) of an ephemeral stream may change as consequence of urban development. However, the change may not be simply an increase or decrease in the duration of surface flow. Urban development is likely to reduce flow in stream networks during dry periods in the spring and early summer when urban hillslopes produce less runoff than forest hillslopes (e.g., runoff in the form of shallow subsurface flow). Conversely, urban hillslope may produce more runoff to streams during storms in summer and autumn than forested hillslopes.

The differences in hydroperiods between urban and suburban ephemeral streams are illustrated by hydrographs from Novelty Hill and Klahanie, two small ephemeral streams that drain catchments with distinctly different land covers in the Puget Lowland (Burges et al., 1998). Novelty Hill drains a 0.37-km$^2$, second-growth forest. Klahanie drains a 0.14-km$^2$, residential development.

During WY 1992, the total duration of flow was longer at Klahanie, despite its smaller drainage area, than Novelty Hill (Figure 3.8). Klahanie had measurable discharge at its outlet on 223 days (61% of the year). In comparison, Novelty Hill had measurable discharge 139 days (38% of the year). The longest continuous period of flow, however was 74 days at Klahanie and 106 days at Novelty Hill.

The longer cumulative duration of runoff from Klahanie was a result of more frequent periods of storm flow during summer and autumn, when the forested hillslopes at Novelty Hill produced little runoff. The periods of continuous flow were shorter at Klahanie than Novelty Hill during winter and spring reflecting the lower water storage capacity of Klahanie's hillslopes and the more rapid delivery of runoff to the stream. The short duration of continuous flow and frequent periods of no flow represent forms of disturbance in ephemeral streams that are likely to increase in response to urban development.

## 3.6. Conclusions

In the Puget Lowland, the total length of perennial streams ($L_{stream}$) in a basin increases with the drainage area of basin (A) and is described generally by $L_{stream} = 0.4A + 0.8$, where $L_{stream}$ has units of km and A has units of $km^2$, with a root mean square percentage error of 9.8% relative to observed stream lengths. Streams draining more than 1.2 $km^2$ had a 50% probability of being perennial while those draining less than 1.2 $km^2$ are likely to be dry at some point during the summer. The extent of perennial streams in a channel network during does not systematically differ between urban and suburban stream basins in the Puget Lowland, nor does it vary with basin shape, valley relief, or valley slope. Urban development may, however, reduce the period of continuous flow in ephemeral streams such that the duration and frequency of droughts increases.

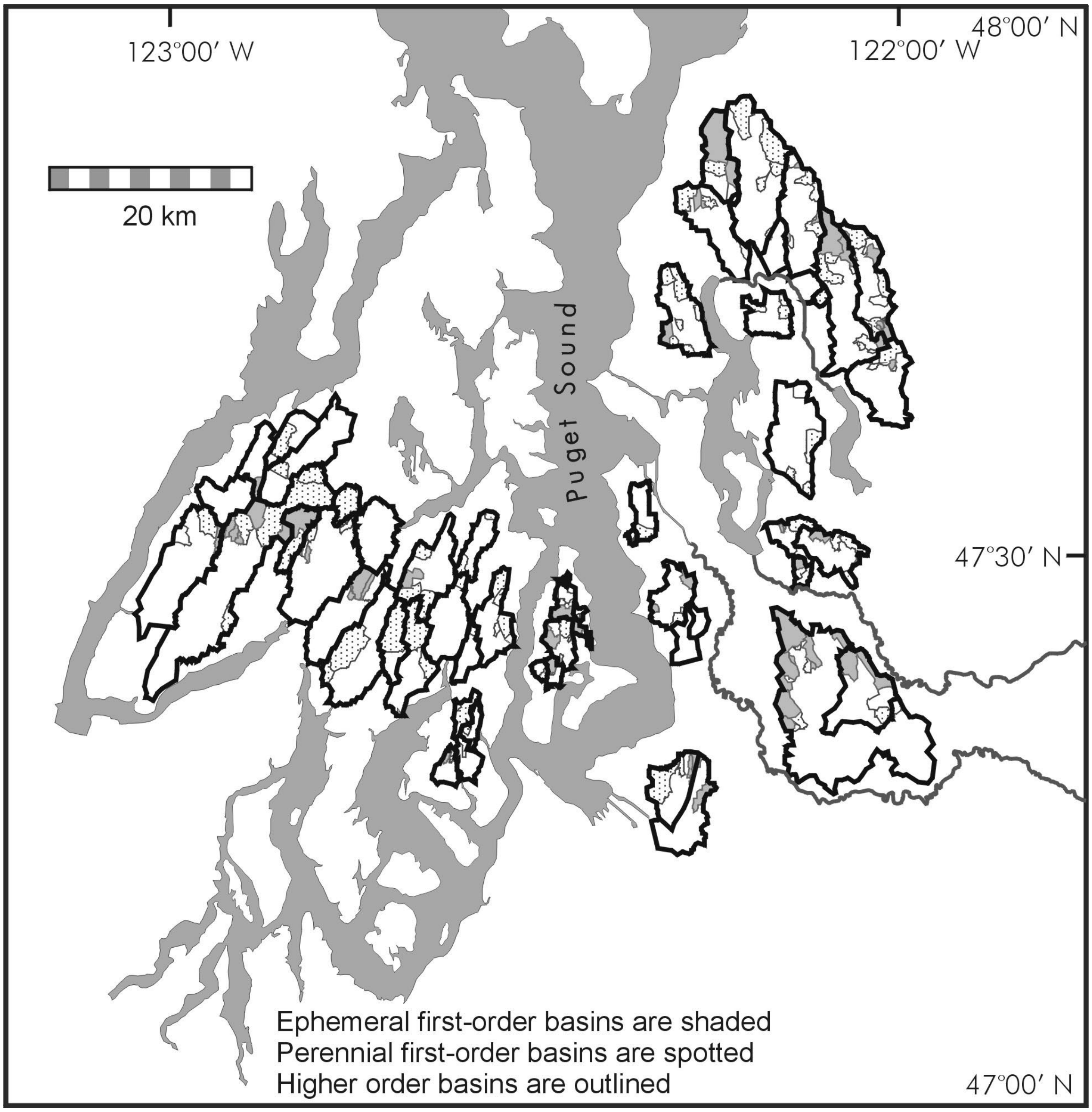



Figure 3.1: Map of Puget Lowland with 179 first-order stream basins included in the low flow analysis.

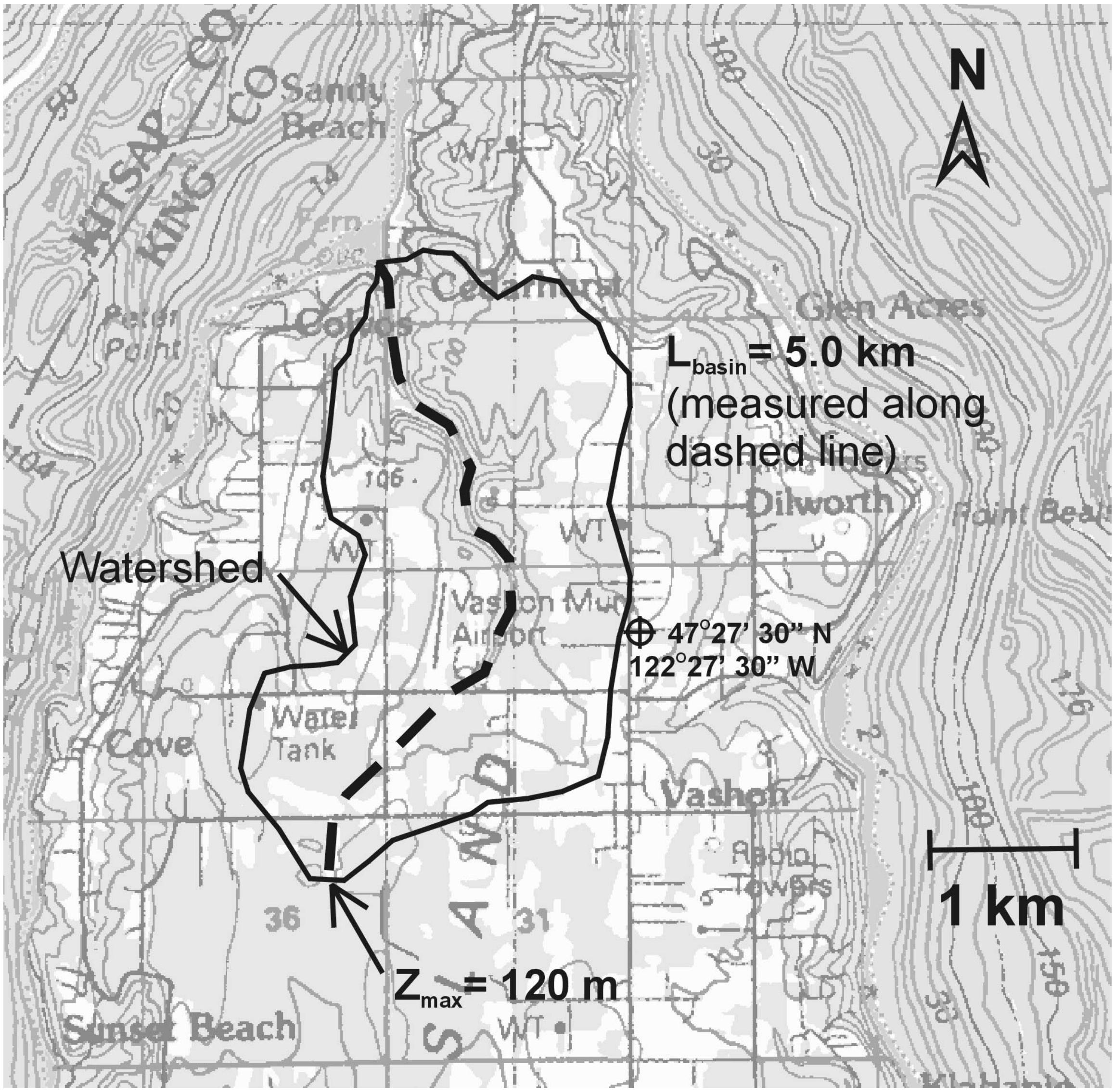



Figure 3.2: Shinglemill Creek basin showing the watershed, basin length ($L_{basin}$), and point of maximum valley elevation ($Z_{max}$).
Source: USGS Vashon 7.5 minute quadrangle, 1: 24,000.

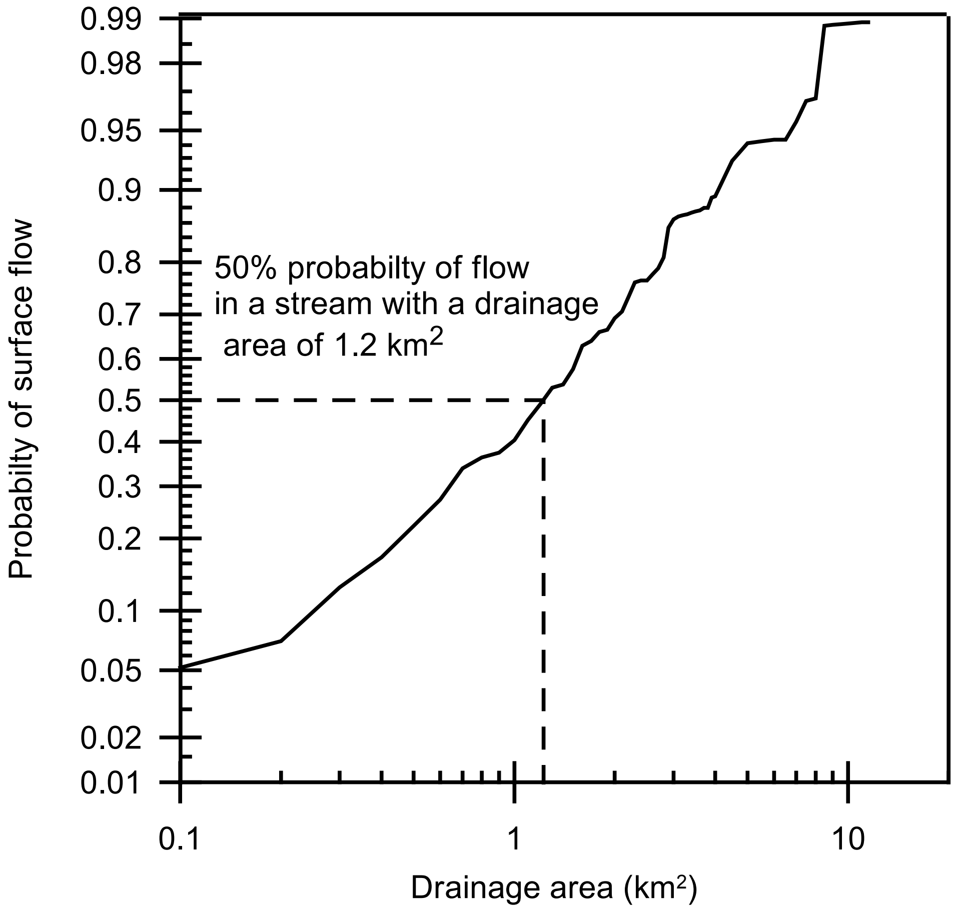
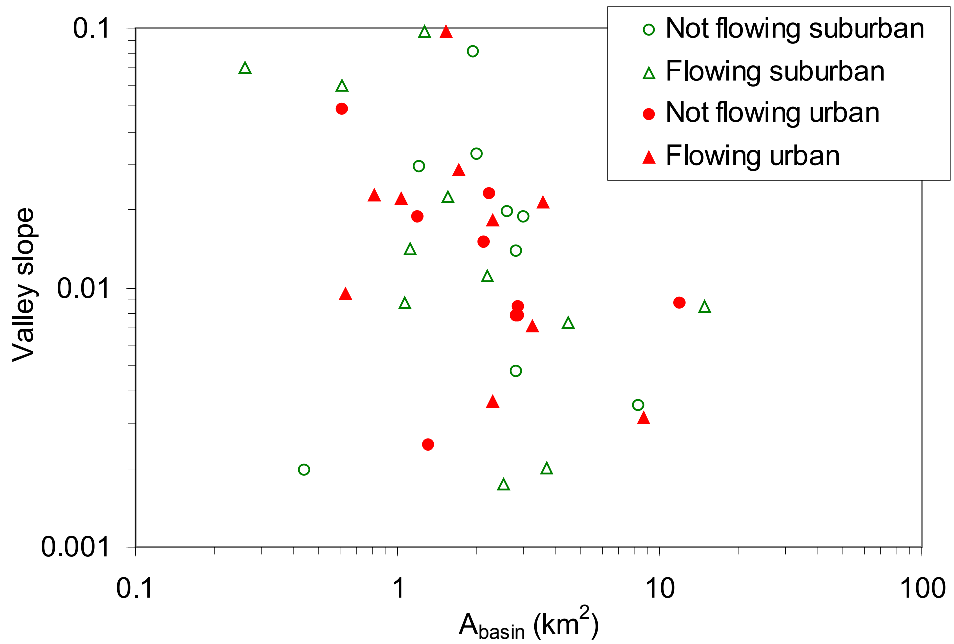
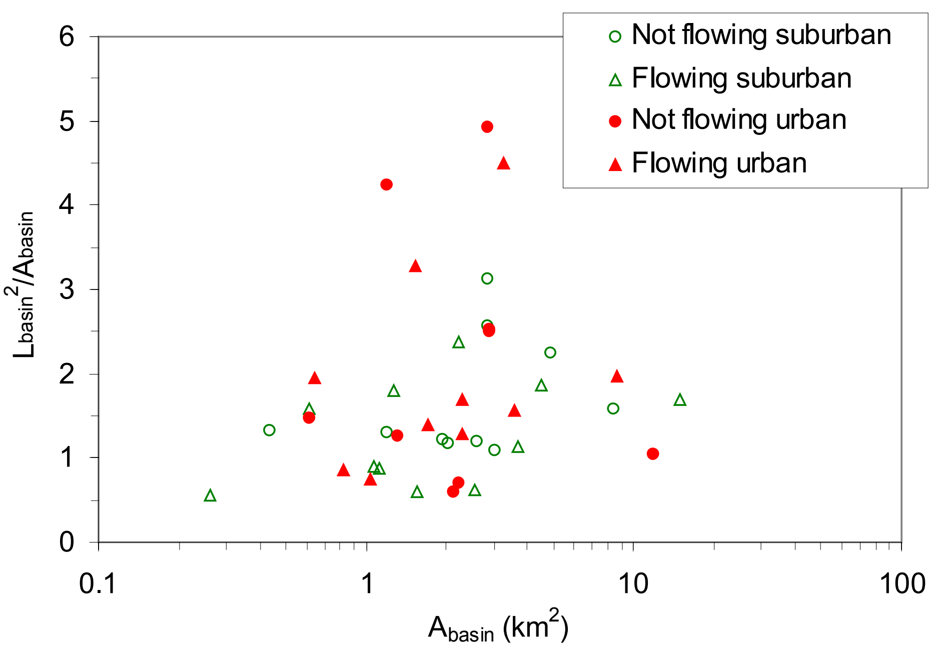



Figure 3.3: Probability of perennial flow for first-order streams as a function of drainage area.

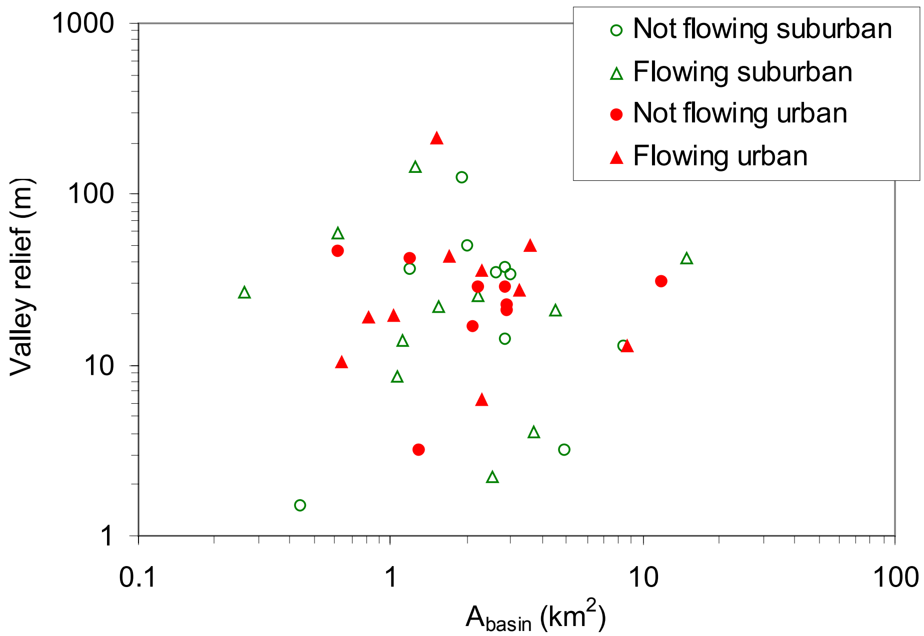



(a)




(b)

Figure 3.4: Drainage area for first-order streams plotted with (a) valley slope, (b) the ratio of the square of basin length to drainage area and, (c) valley relief.




(c)

Figure 3.4 (continued).

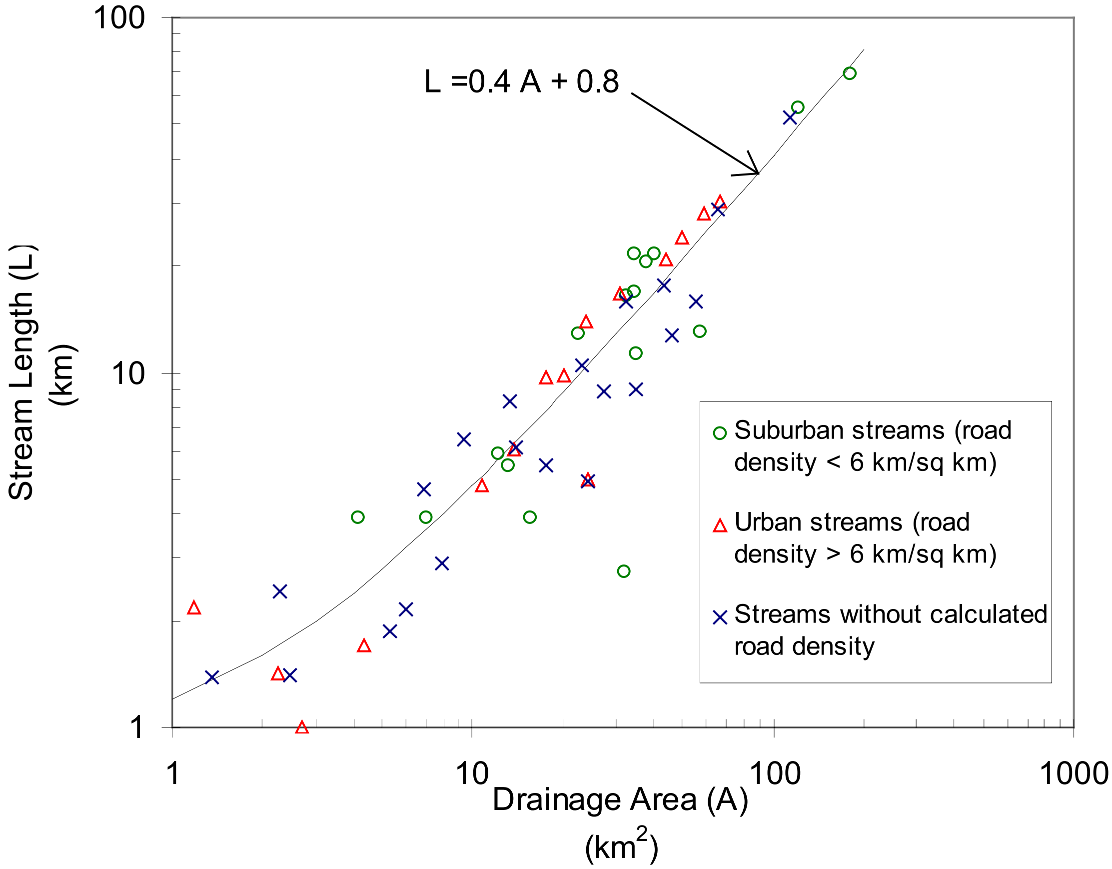



Figure 3.5: Comparison of base flow length (L) and drainage area (A) for 52 stream basins in the Puget Lowland, Washington.

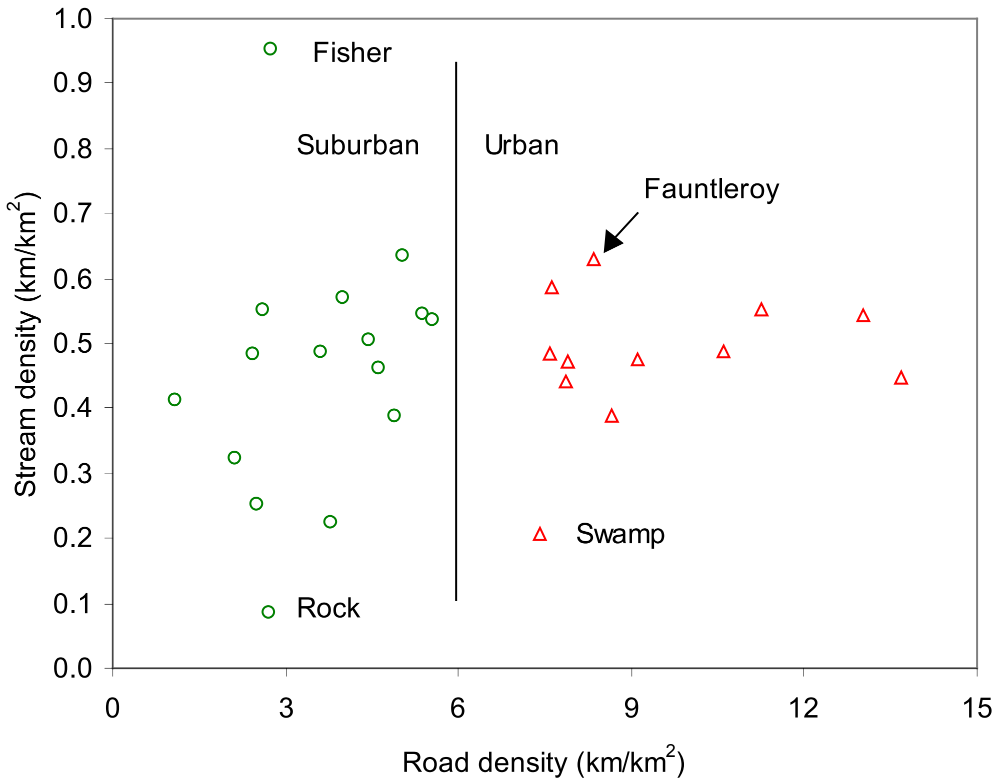
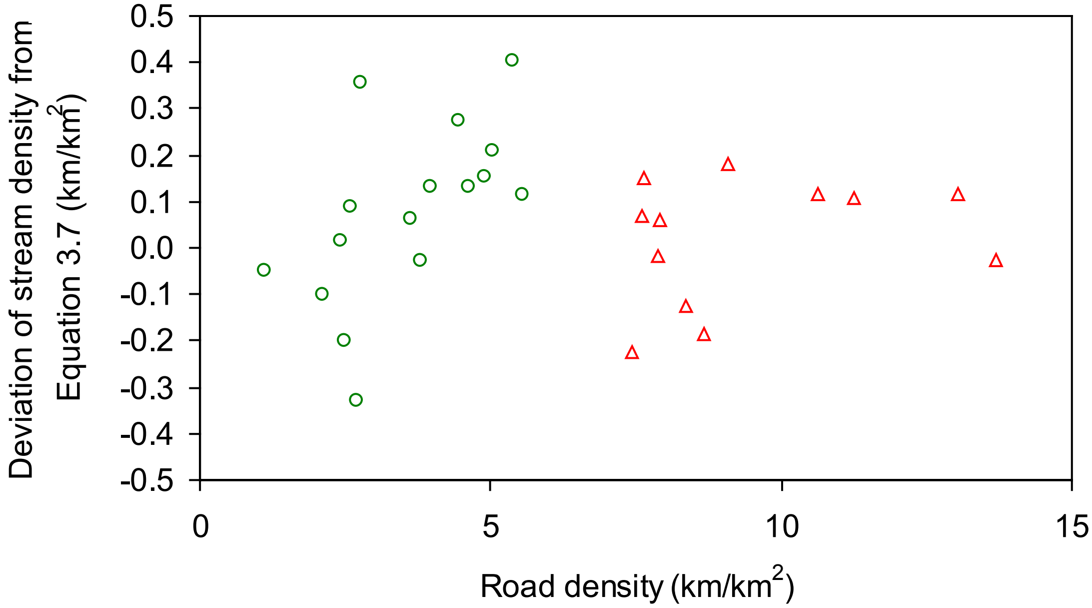
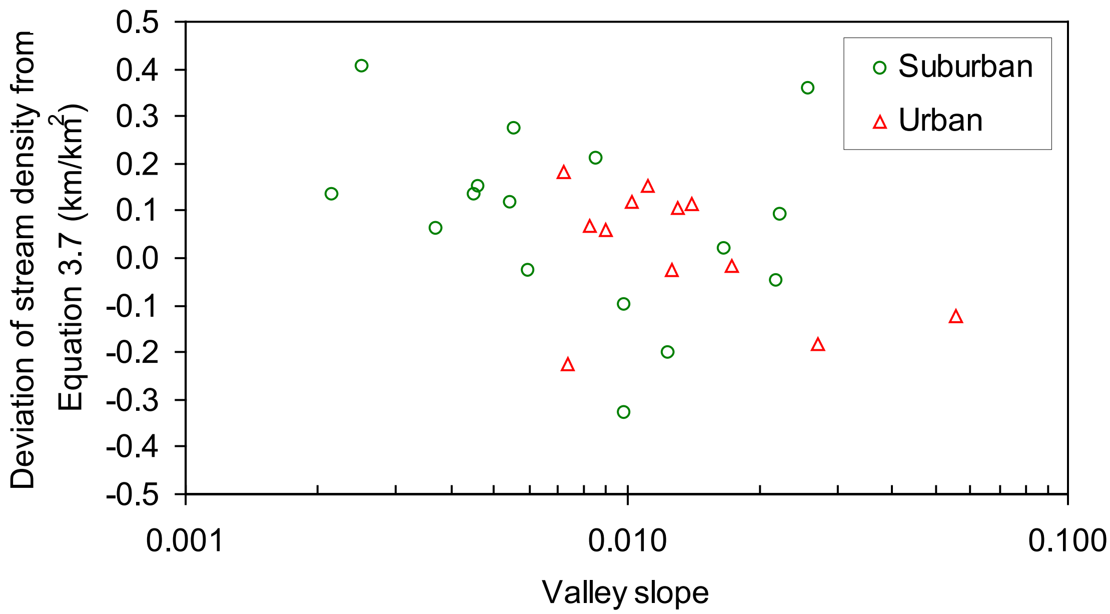



Figure 3.6: Comparison of road and stream densities for 29 stream basins in the Puget Lowland, Washington.

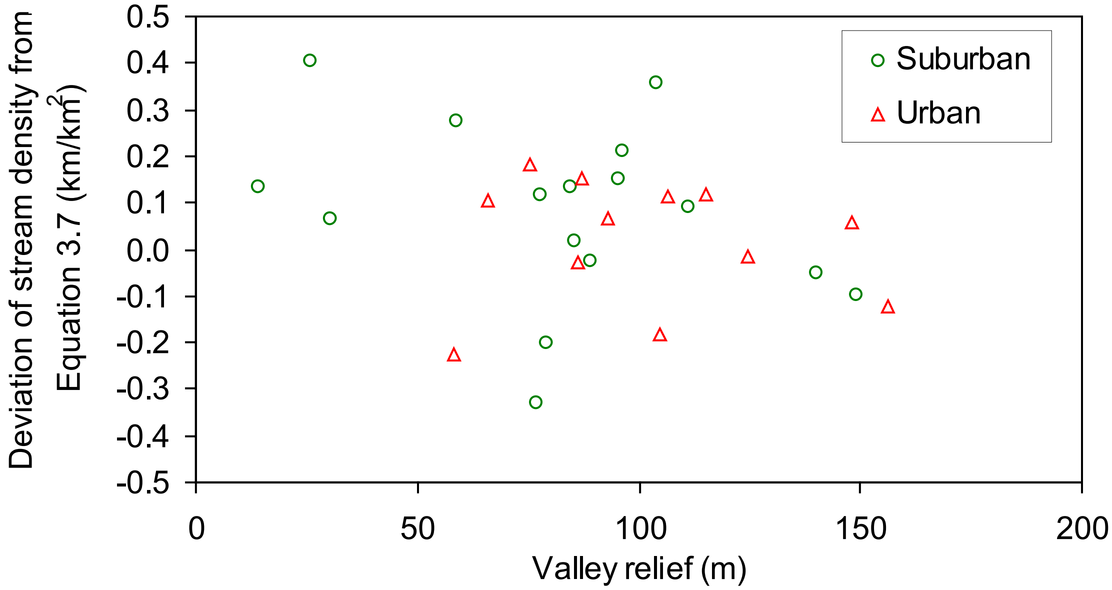



(a)

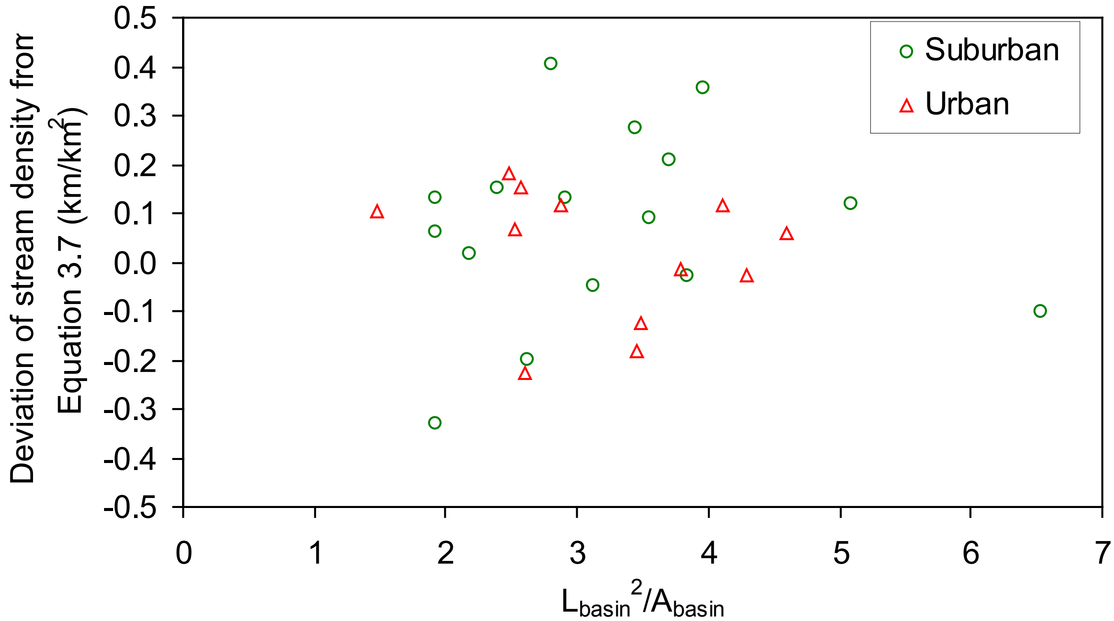



(b)

Figure 3.7: Deviation of mapped and calculated drainage densities plotted with (a) road density, (b) valley slope, (c) valley relief, (d) valley length$^2$/area.




(c)




(d)

Figure 3.7 (continued).

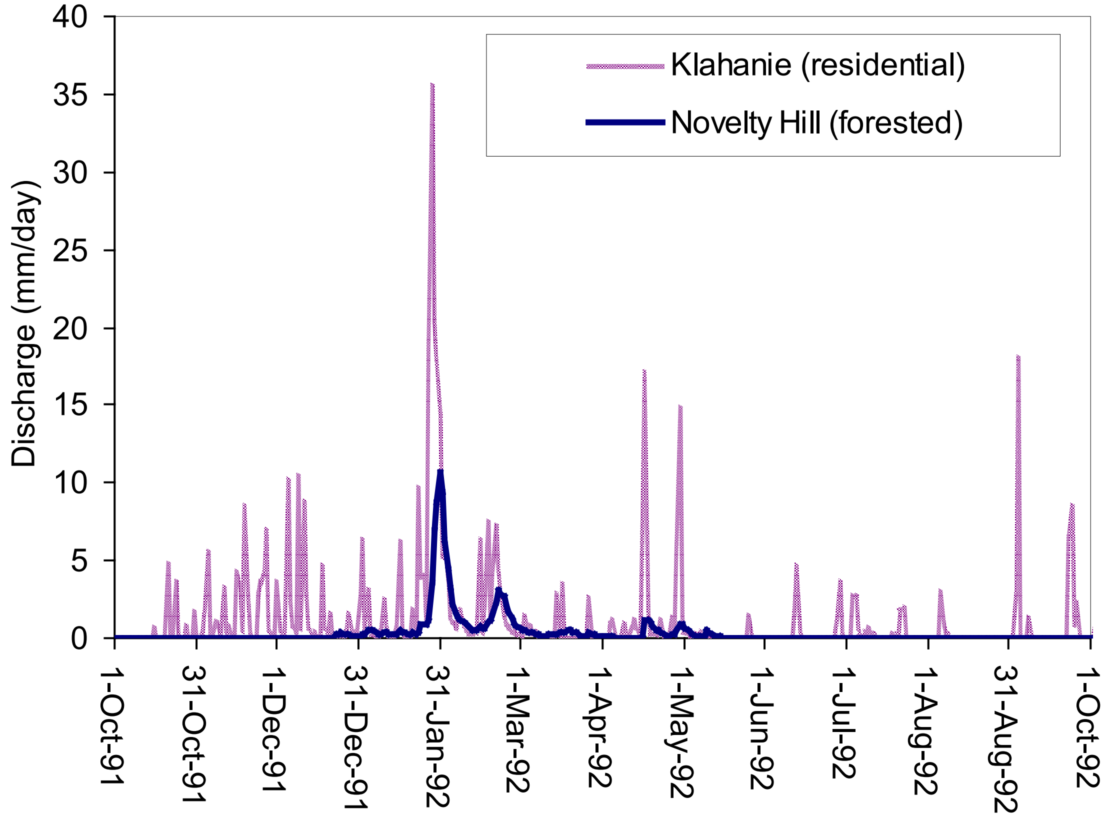



Figure 3.8: Hydrographs of daily discharge for two ephemeral streams.
Source: Burges et al. (1998)

Table 3.1: Streams in the low flow analysis.

| Suburban Streams | Drainage area ($km^2$) | Perennial stream density ($km/km^2$) | Number of first-order streams in analysis | Road density ($km/km^2$) |
|---|---|---|---|---|
| Fisher | 4 | 0.95 | 4 | 2.7 |
| Shinglemill | 7 | 0.55 | 5 | 2.6 |
| Judd | 12 | 0.48 | 5 | 2.4 |
| Stavis | 13 | 0.41 | 1 | 1.1 |
| Huge | 16 | 0.25 | 1 | 2.5 |
| May @ Coal Cr. Prkwy. | 22 | 0.57 | 4 | 4.0 |
| Rock | 32 | 0.09 | 1 | 2.7 |
| Bear @ NE 133rd St. | 32 | 0.51 | 16 | 4.4 |
| May | 34 | 0.64 | 9 | 5.0 |
| Evans | 35 | 0.49 | 4 | 3.6 |
| Big Beef | 35 | 0.32 | 2 | 2.1 |
| Jenkins | 38 | 0.55 | 6 | 5.4 |
| Little Bear | 40 | 0.54 | 8 | 5.5 |
| Covington | 57 | 0.23 | 0 | 3.8 |
| Bear @ Union Hill Rd. | 121 | 0.46 | 24 | 4.6 |
| Soos | 179 | 0.39 | 17 | 4.9 |
| Average for suburban streams | | 0.46 | | |
| Urban Streams | | | | |
| Fauntleroy | 1 | 1.84 | 3 | 14.4 |
| Hollywood | 2 | 0.63 | 2 | 8.4 |
| Cedar Trib 0308 | 3 | 0.37 | 1 | 7.6 |
| Maplewood | 4 | 0.39 | 2 | 8.7 |
| Longfellow | 11 | 0.45 | 1 | 13.7 |
| Des Moines | 14 | 0.44 | 1 | 7.9 |
| Juanita | 17 | 0.55 | 3 | 11.3 |
| Miller | 20 | 0.49 | 4 | 10.6 |
| West Fork Hylebos | 24 | 0.59 | 4 | 7.6 |
| Swamp @ Filbert Road | 24 | 0.21 | 2 | 7.4 |
| Thorton | 31 | 0.54 | 7 | 13.0 |
| Mercer | 44 | 0.47 | 4 | 9.1 |
| Hylebos | 50 | 0.48 | 5 | 7.6 |
| Swamp | 59 | 0.47 | 11 | 7.9 |
| North | 67 | 0.46 | 5 | 7.5 |
| Average for urban streams | | 0.56 | | |

Table 3.1 (continued)

| Road density not calculated for these streams | Drainage area ($km^2$) | Perennial stream density ($km/km^2$) | Number of first-order stream basins in analysis |
|---|---|---|---|
| Fern | 0.2 | 1.9 | 0 |
| Needle | 0.6 | 0.7 | 0 |
| Vason | 1.4 | 1.0 | 2 |
| Ellisport | 2.3 | 1.1 | 1 |
| Christiansen | 2.5 | 0.6 | 1 |
| Donkey | 5.3 | 0.4 | 1 |
| McCormick | 6.0 | 0.4 | 2 |
| Artondale | 6.9 | 0.7 | 2 |
| Wollochet | 7.9 | 0.4 | 1 |
| Purdy | 9.3 | 0.7 | 1 |
| Salmonberry | 13.3 | 0.6 | 2 |
| Seabeck | 14.0 | 0.4 | 2 |
| Anderson | 17.6 | 0.3 | 1 |
| Olalla | 23.2 | 0.5 | 4 |
| Gorst | 24.2 | 0.2 | 0 |
| Burley | 27.2 | 0.3 | 1 |
| Blackjack | 32.2 | 0.5 | 4 |
| Coulter | 34.8 | 0.3 | 4 |
| Minter | 43.2 | 0.4 | 3 |
| Rocky | 45.8 | 0.3 | 2 |
| Dewatto | 55.5 | 0.3 | 3 |
| Union | 65.7 | 0.4 | 6 |
| Tahuya | 112.8 | 0.5 | 14 |

# Chapter 4: Partial entrainment of the surface material from gravel bars during floods

Sand, gravel, cobbles, bedrock, logs, and roots form stream beds in the Puget Lowland, Washington. Stream bed material provides habitats for algae, protozoa, mollusks, insects, amphibians, and other benthic organisms in lotic communities. Floods, defined here as transient periods of increased stream discharge typically resulting from quick-response runoff (i.e., overland flow and shallow subsurface flow) during rainstorms and snowmelt, entrain material from the surface of a stream bed and transport it downstream. The mechanical disturbance of the stream bed surface changes it biological conditions (e.g., the mass, age, and health of benthic organisms, the types and distribution of benthic species, and the sizes of their populations). The biological effects of floods depend, in part, on the frequency that the bed is disturbed and the spatial extent of the disturbance, which can range from a few grains to all of the material forming the stream bed surface.

Small, frequent floods entrain only a portion of the material comprising the surface of gravel stream beds. Thus, partial entrainment of the stream bed surface represents a common mode of disturbance in gravel-bed stream ecosystems. The objective of this investigation is to develop a method for the quantitative assessment of the spatial extent of bed disturbance during floods in gravel-bed streams. The approach is to relate the fraction of a gravel bar’s surface disturbed during a flood, which referred to here as partial entrainment, to the peak shear stress generated by the flood and the particle-size distribution of the bed surface material.

## 4.1. Ecological effects of floods

Stream flow patterns influence a wide range of biological conditions in streams (Shelford and Eddy, 1929; Odum, 1956; Horwitz, 1978; Fisher et al., 1982; Schlosser 1985; Resh et al., 1988; ASCE Task Committee, 1992; Closs and Lake, 1994; Death and Winterbourne, 1995; Poff and Allan, 1995). Among the many different stream flow patterns with biologic influences (see Richter et al., 1996; Poff et al., 1997), the disturbance regime dictated by flood patterns has a broad influence on the type and relative abundance of organisms in lotic communities. Anthropogenic changes in flood disturbance regimes have been identified as a principal cause for degradation of stream ecosystems where flood frequency and magnitude have been either reduced by reservoirs (Ward and Stanford ,1979; Wooton et al., 1996) and or increased by urban development (Orser and Shure, 1972).

Floods disturb lotic communities when high flows transport sediment and organic material downstream. The biologic effects of flood disturbances include decreased periphyton biomass (Douglas, 1958; Fisher et al., 1982; Power and Stewart, 1987; McCormick and Stevenson, 1991), decreased densities and populations of benthic invertebrates and vertebrates (Stehr and Branson, 1938, Anderson and Lehmkuhl, 1968; Orser and Shure, 1972), decreased taxonomic diversity of fish (Gorman and Karr, 1978), increased taxonomic evenness (i.e., the relative abundance of different species) (McAuliffe, 1984), decreased predator abundance (Closs and Lake, 1994), and increased dominance of diatoms rather than detritus as trophic base (Fisher et al., 1982). While these effects have been widely observed, the disturbance response of a specific lotic community to a flood is mediated by many biotic and abiotic factors, ranging from local to basin scales, including habitat diversity (Gorman and Karr, 1978; Gurtz and Wallace; 1984) and biotic interactions (McAuliffe, 1984; Feminella and Resh, 1990; McCormick and Stevenson, 1991; Wootton et al., 1996).

Many biological conditions in streams recover within months after a flood (Fisher et al., 1982; DeBray and Lockwood, 1990; Boulton et al., 1992; Jones et al.,1995) though the rate of recovery varies, for example, among species (Power and Stewart, 1987). Lotic communities can recovery rapidly after floods because of the availability of refugia, the ability of organisms to disperse and recolonize recently disturbed habitat, and the high growth rates of their populations (Patrick, 1975; Schlosser, 1985; Allan 1996; Matthaei et al., 1999).

Over many cycles of disturbance and succession, the community-scale effects of disturbance regimes depend on the areal extent, patchiness, and frequency of disturbance (White and Pickett, 1985). Connell (1978) proposed that the highest levels of diversity and production in a biological community are associated with a disequilibrium maintained by moderate levels of disturbance in both space and time. According to this "intermediate disturbance hypothesis," disturbance promotes production and diversity in communities when (1) organisms with high growth rates replace those with lower growth rates (because of differences between the organisms both in age and taxonomy), and (2) populations of superior competitors do not have sufficient time between disturbances to exclude other species. In contrast, biological communities in ecosystems with either frequent and extensive disturbance or no disturbance are likely to have less diverse taxonomic, trophic, and life cycle history characteristics.

The intermediate disturbance hypothesis has been applied to lotic communities particularly with the development of the patch dynamics model in which benthic habitat represents a mosaic of patches, each with an assemblage of organisms in some stage of post-disturbance succession (Townsend, 1989). Whereas the conditions within a patch depend on the frequency and extent of disturbance in that patch, the structure and composition of the larger community that comprises many patches is influenced by the frequency, extent, size, and number of patches of benthic habitat disturbed, for example, during floods. Thus, the ecological effects of a flood depend on the spatial extent of disturbance both within and among habitat patches forming a stream bed.

Benthic habitat in streams comprises many different patches including channel forms such as pools and bars (riffles), floodplains, and smaller elements such as large rocks, patches of sand, organic debris, periphyton, and roots. Unique among these different channel forms, gravel bars have a diversity of hydraulic conditions (e.g., the juxtaposition of high velocity currents and low velocity zones of flow separation that occur over hydraulically rough surfaces (Morris, 1955).

Gravel bars are defined here as local in-channel deposits of gravel that rise above the mean profile of the stream bed surface (e.g., Church and Jones, 1982). In pool-riffle channels, bars span from pool to pool forming a riffle downstream of their crest. Low-relief bars may also form in otherwise plane-bed reaches. The surface material of a gravel bar is typically coarser than the bed as a whole, though the texture of bed material may vary along a bar (Allen, 1965).

The hydraulic and sedimentologic characteristics of gravel bars appeal to many different types of benthic organisms (Statzner et al., 1988) allowing bars to support high densities of those organisms (McAuliffee, 1984). Given their stability relative to finer grained alluvial deposits, gravel bars may provide refugia for benthic organisms during floods. Furthermore, gravel bars comprise a large fraction of the stream bed in many types of channels providing much of the habitat available for benthic organisms in a stream.

In spite of its biological importance, the spatial extent of bed material entrainment has not been quantified in ecological investigations of flood disturbance (e.g., Anderson and Lehmkuhl, 1968; Fisher et al., 1982; Biggs and Close, 1989; Boulton, 1992). Some investigations have employed a threshold discharge value for a “disturbing” flood corresponding to a threshold shear stress for entraining bed material (e.g., McElravy et al., 1989; Feminella and Resh, 1990). While a threshold force is necessary for moving an individual particle on the surface of a stream bed, there is a wide distribution of particle sizes and hydraulic forces over a gravel bar so that any single-valued threshold is either a local measure or an index for some extent of disturbance of a natural gravel stream bed.

As a alternative, I develop a method for estimating the spatial extent of disturbance in terms of the fraction of a gravel bar's surface entrained during a flood.

## 4.2. Threshold condition for entrainment of stream bed material

Stream flow entrains sediment and organic debris from the stream bed when the force applied by water flowing over the bed exceeds the resisting force of the bed material. The initial motion of particles on a stream bed can be analyzed as a threshold-force problem. Newton's second law of motion holds that a particle will begin moving when water imparts a force in excess of the resisting force of the particle. The resisting force of a single particle depends on both particle properties (e.g., mass, density, shape) and bed properties (e.g., slope, particle-size distribution, packing, protrusion, and pivot angle between particles). The applied force includes drag and lift components which depend, respectively, on the near-bed velocity of stream flow and its vertical velocity gradient around the particle. The applied force for a given velocity distribution will also depend on the shape of a particle, its exposed surface area, and the form and magnitude of turbulence in stream flow (Rouse 1978).

Sediment transport has been analyzed as a threshold problem at spatial scales ranging from individual grains (White, 1940; Fenton and Abbot, 1977) to the bed of a flume or stream (e.g., DuBoy, 1879 and Shields, 1936 cited in Leliavsky, 1959; Lane and Kalinske, 1940; Einstein, 1942; Meyer-Peter and Muller, 1948; Parker et al., 1982; Reid and Frostick, 1986; Kuhnle, 1993; Wilcock, 1993). Under these approaches, sediment entrainment is analyzed in terms of the applied and resisting shear stresses, which are tangential forces per unit surface area of a stream bed.

White (1940) introduced a grain-scale mechanical analysis of incipient motion that accounts for many of the sources of variation in threshold of motion among individual particles including packing and protrusion, angle of repose, velocity distribution, form and surface drag. White (1940) formulated the threshold condition in terms of the critical shear stress ($\tau_{cr}$) required to entrain a spherical particle from a stream bed:

$$\tau_{cr} = \alpha\eta\frac{\pi}{6}(\gamma_{grain} - \gamma_{water})D_{grain}\tan\varphi \qquad (4.1)$$

where $\alpha$ is a coefficient that accounts for the moment acting on the particle if the applied force of the stream flow doesn't act through the particle's center of gravity, $D_{grain}$ is the grain diameter, $\eta$ is a packing coefficient defined as the product of $D_{grain}^2$ and the number of grains per unit bed area, $\gamma$ is the specific weight, and $\phi$ is the pivot angle or angle of repose of the grain. The packing coefficient is low when grains are generally exposed and protruding up into the current. According to Equation 4.1, the critical shear stress of a spherical particle is proportion to its diameter. For a non-spherical particle, $D_{grain}$ is approximated by the length of the intermediate axis of the particle.

Lane and Kalinske (1940) recognized that both the critical shear stress ($\tau_{cr}$) to move bed material and the shear stress applied by stream flow have distributed values in space and time. The distribution of $\tau_{cr}$ for a stream bed ranges widely where the bed material is well graded (i.e., has a large range of particle sizes) or comprises many distinct textural patches (i.e., contiguous regions of a stream bed with similar particle-size distributions) (Paola and Seal, 1995). Furthermore, the resisting force of particles of the same size class will vary as a result of differences in pivot angles, particle shapes, packing, clustering and other bed structures (Miller and Byrne, 1966; Brayshaw et al., 1983; Carling, 1983; Li and Komar, 1986; Kirchner et al., 1990; Gomez, 1994). Despite the influence of these factors, the median of a particle-size distribution ($D_{50}$) forming the surface of a sediment indicates the prevailing hydraulic conditions controlling sediment transport and deposition (Inman, 1949) and, for unimodal distributions, provides a reliable basis for estimating its critical shear stress (Wilcock, 1993).

The distribution of $\tau_{cr}$ over a stream bed may change as material is entrained and deposited. For example, $\tau_{cr}$ increases at places where stream flow erodes small particles from the bed surface leaving coarser particles on the bed surface (armoring); conversely, $\tau_{cr}$ decreases when high flows entrains material indiscriminately with respect for particle size from an armored bed (Little and Mayer, 1976; Garde et al., 1977; Gomez ,1983, 1994; Shen and Lu, 1983; Kuhnle and Southard, 1988; Chin et al., 1992). Changes in $\tau_{cr}$ may be evident as variation over time in the particle-size distribution of the bed surface or variation over time in the relationship between bed load transport rates and stream power (Gomez, 1983a, b). At low sediment transport rates, $\tau_{cr}$ may continue to increase as particles move into more stable positions even if the particle-size distribution of the surface is unchanged (Church et al., 1998).

The magnitude of the force applied by water also has distributed values in time and space rather than a single, uniform value. The force applied by stream flow varies over time as a stream rises and fall, while it varies from section to section where the channel's form changes (e.g., width, direction). The applied force typically has low values near its banks where the current is slow and high near the center of the channel where the current is swift. Furthermore, the distribution, protrusion, and shape of bed material influence the magnitude of the forces applied locally by stream flow on the bed (Rouse, 1965).

At time scales of seconds or less, the force of water applied locally to particles on the stream bed varies with the sweeps and bursts of turbulent flow (Grass, 1971; Jackson, 1976; Nelson et al., 1995). These turbulent fluctuations in velocity generate apparent or "Reynolds" stresses in addition to the sustained shear stress generated by the steady component of the stream flow (Schlichting, 1979).

Taylor (1935) observed that the correlation between fluid velocities in space and time for turbulent flow can be represented as a diffusion process. In this case, the additional Reynolds stress of turbulent flow form a Gaussian distribution around the "mean" stress generated by steady flow. Einstein (1942) and Gessler (1970) used a

Gaussian distribution to represent shear stress for turbulent flow, although Grass (1970) found that the applied shear stress over a hydraulically-smooth bed had a positively skewed distribution at any given instant.

The total boundary shear stress ($\tau_0$) is used in this analysis as a central measure of the distribution of the applied shear stress over a stream bed. In turbulent flow, the magnitude of local shear stresses varies in space and time about $\tau_0$. The total boundary shear stress along a reach with uniform flow is calculated as:

$$\tau_0 = \gamma_{water} RS \tag{4.2}$$

where $\gamma_{water}$ is the specific weight of water, R is the hydraulic radius, and S is the downstream energy gradient of the stream flow. For uniform flow, the total boundary shear stress is balanced on a unit bed-area basis by all the forces that resist the gravitational acceleration water as it flows through a channel.

Only a portion of the total applied shear stress, represented by $\tau_0$, is effective at entraining and transporting bed material. The portion of the total shear stress that acts on the bed surface is referred to as the grain shear stress ($\tau_g$), or "skin friction" while another portion of the total boundary shear stress, the form drag component, acts in regions of flow separation but is dissipated as turbulence without contributing to sediment transport (Einstein and Barbarossa, 1952; Smith and McLean, 1977; McLean et al., 1999). The Prandtl-Von Karman logarithmic velocity distribution can be used to estimated $\tau_g$ (Nece and Smith, 1970; Grass, 1971) for a fully developed turbulent boundary layer where the vertical velocity profile does not change in the streamwise direction (Schlichting, 1979).

The streams in this analysis, however, are relatively shallow with many individual particles casting wakes that extend in some cases to the water surface. As a result, the vertical velocity profiles are not strictly logarithmic (Wiberg and Smith, 1991). $\tau_{grain}$ can only be estimated from velocity measurement made at small distances from the bed surface (Wilcock, 1996; Biron et al 1998). Because the turbulent boundary layer is not fully developed in these streams (i.e., the velocity profile varies downstream of individual roughness elements), any single estimate of $\tau_g$ would be a local value and not necessarily indicative of the bar-scale distribution of $\tau_g$. Given these constraints, this analysis uses $\tau_0$ but is limited to regions of approximately uniform flow along the tops of long, low relief bars in reaches where form drag is relatively small. At these sites, $\tau_0$ should provide a consistent index for the distributed values of $\tau_g$ but with a bias to higher values.

The balance between the applied and critical shear stresses in a reach can be represented by a dimensionless shear stress ($\tau_0$*), which is the ratio of total boundary shear stress to the unit-area buoyant weight of the median of the particle-size distribution:

$$\tau_0{}^* = \frac{\tau_0}{(\gamma_{\text{sediment}} - \gamma_{\text{water}})D_{50}} \tag{4.3}$$

$\tau_0$* was introduced by Shields in 1936 as a parameter to define the threshold of motion for a uniform sediment, which varies as a function of the ratio of the particle diameter to the height of the laminar boundary layer (Lelivsky, 1959).

The reported values of $\tau_0$* at the threshold of motion for particles (i.e., the critical dimensionless shear stress) in a turbulent boundary layers span a wide range from 0.02 to 0.08 likely reflecting different criteria used to define initial motion, from movement of a single particle to wide-spread entrainment of a bed's surface (ASCE Task Committee, 1966; Buffington and Montgomery, 1997). Gilbert (1914) recognized that some particles are entrained at much lower flow rates than others and reported the portion of particles moving in his flume-experiment results. Neill and Yalin (1969) proposed that a constant bed load-transport rate should be the basis for a quantitative criterion of initial motion. They show how the number of particles entrained under such a standard varies depending on the size of particles (i.e., "initial motion" for a fine-grained sediment corresponds to more moving particles than initial motion of a coarser sediment). Rather than adopting a single criterion for initial motion of a sediment, I analyze how the partial entrainment of a sediment varies with dimensionless shear stress.

## 4.3. Probabilistic models of bed material entrainment

Over scales larger than a single particle, the partial entrainment of a bed surface is equal to the fraction of its area where $\tau_g$ exceeds $\tau_{cr}$ of the bed material. If magnitude of $\tau_g$ and $\tau_{cr}$ are known at specific locations on the bed surface, then partial entrainment of the bed surface can be calculated. If the magnitudes, but not the locations, of $\tau_g$ and $\tau_{cr}$ are known, then it is uncertain whether the bed material at a point will be entrained but the probability of entrainment can be estimated. For spatially-homogeneous distributions of $\tau_g$ and $\tau_{cr}$, the probability of entrainment at a point is equal to the product of the marginal distribution of $\tau_g$ and the fraction of the cumulative distribution of $\tau_{cr}$ less than $\tau_g$ integrated for the range of $\tau_g$ applied over the bed surface. In this case, the probability of entrainment at a point is equal to the expected value of partial entrainment (Einstein, 1942).

Probabilistic approaches have been used in sediment transport models to calculate the supply of material to bed load transport (Einstein 1950, Bridge and Bennett 1992), and to estimate the extent of bed material entrainment (Grass, 1970). Three processes, in particular, have been represented with probability distributions in sediment transport models: fluctuations in the applied shear stress resulting from velocity fluctuations in turbulent flow (Einstein, 1942); size-selective transport of a mixed sediment, where $\tau_{cr}$ of particles varies with their size (Lane and Kalinsky, 1940); and partial transport of a mixed sediment, where only some particles in a given size class of material are mobile (Wilcock, 1997).

Models developed by Einstein (1942, 1950), Gessler (1970), and Wilcock (1997) are described below and applied to predict the extent of partial entrainment of a gravel bar's surface during a flood. Each model provides an estimate of the mobility of individual size classes of material in a mixed-size sediment which must be summed over all size classes to estimate the total extent of bed material entrainment. Einstein's bed load function (1950) is a steady-state model of sediment transport where particles are entrained randomly by turbulent fluctuations in stream flow. The probability of entraining a grain is equal for all grains in a size class. Since the entrainment probability is spatially independent, the total surface area of a gravel bar moved during a flood depends on the both the exchange probability and the duration of the flood.

Gessler and Wilcock recognized that the probability of entrainment varies among grains within a size class. Their models do not assume spatial independence of $\tau_{cr}$, so the probability of entrainment will decline as a bed becomes armored (Gessler, 1970) or will be higher in regions covered by a mobile population of grains (Wilcock, 1997). Thus over time, the cumulative area of stream bed entrained at a given flow will approach a limit that may be less than the whole bed.

### 4.3.1. Exchange probability

Einstein (1942) introduced the concept of the "exchange probability" when he formulated a bed load transport equation for a uniform sediment in turbulent flow. He later applied this probabilistic approach to the transport of a mixed-size sediment (Einstein, 1950). The exchange probability for a given size class of material represents the fraction of time during which particles in that size class are entrained and replaced with particles of the same size. The bed load transport rate per unit width of channel ($q_i$) is related to the exchange probability ($p_i$) for the $i^{th}$ size class of bed material:

$$q_i = \frac{f_i}{A_1 D_i^2} \frac{p_i}{T_{ei}} A_2 D_i^3 \rho_s A_L D_i \qquad (4.4)$$

where $A_1D_i^2$ is the area of bed area occupied by a single grain, $T_{ei}$ is the time required for an eroded particle to be exchanged or replaced by another particle of the same size, $f_i$ is the fraction of the bed material in a size class, $A_2D_i^3\rho_s$ is the mass of a grain, and $A_LD_i$ is the mean displacement length each time a particle moves. The coefficients $A_1$, $A_2$ and $A_L$ are constant for all size classes. The frequency of exchanges is represented by $p_i/T_{ei}$.

The exchange probability is equal to the fraction of the stream bed from which material is entrained at any instant for all locations on the bed that have the same distributions of $\tau_0$ and $\tau_{cr}$. Einstein (1950, p. 34) describes the areas satisfying this condition as "statistically-equivalent". The exchange probability for each size class of material is:

$$p_i = \frac{A_*\Phi_{*i}}{1 + A_*\Phi_{*i}} \tag{4.5}$$

where $A_* = 43.5$ and $\Phi_{*i}$ is a dimensionless measure of the intensity of bed load transport. Equation 4.5 is the corrected form of Einstein's (1950) Equation 57 for the exchange probability (Raudkivi 1990).

For each size class of material, the intensity of bed load transport ($\Phi_{*i}$) is related to the flow intensity ($\Psi_{*i}$) using a curve fit to observed transport rates and flow intensities (Figure 10 in Einstein, 1950). The parameter $\Psi_{*i}$, an inverse form of $\tau_0$*, accounts for a variety of factors that modify the applied shear stress (e.g., bar resistance, local velocity around the grain, hiding of smaller grains behind larger ones, and the pressure force generated by flow separation at the grain's boundary). The relationship between $\Phi_{*i}$ and $\Psi_{*i}$ allows the exchange probability for a given size class to be calculated from the dimensionless shear stress.

### 4.3.2. Probability for a particle to remain in a surface layer

Gessler (1970) used a clear-water flume to form an armored bed surface representing a static equilibrium where the population of mobile grains was depleted over the duration of an experimental run. Gessler developed an analytical function to estimate the cumulative probability that particles will remain as part of an armored surface layer as a bed degrades. The cumulative probability of a particle of size i to remain on the bed ($PS_i$) is:

$$PS_i = \frac{m_{i_{surface}}}{(m_{i_{surface}} + m_{i_{entrained}})} \frac{f_{i_{surface}}}{f_i} \tag{4.6}$$

where $m_{i\ surface}$ is the weight of surface material per unit area of bed, $m_{i\ entrained}$ is the weight of material entrained from the bed per unit area of bed, $f_{i\ surface}$ is the particle-size distribution of the surface layer, and $f_i$ is the particle-size distribution of the parent material. Gessler found that for any size class of material, the probability of remaining on the bed surface was normally distributed with respect to $\tau_{50i}/\tau_0$ where $\tau_{50i}$ is the shear stress that entrains 50% of the surface particles of size i.

### 4.3.3. Mobile proportion of individual size classes

Wilcock and McArdell (1993; 1997) observed the mobile and immobile populations of particles in a sediment-recirculating flume. Based on these observations, Wilcock (1997) proposed a conceptual model for the mean bed load transport rate of an individual size fraction under steady-state conditions that expands the entrainment probability to include both the portion of a size-class that is mobile ($Y_i$) and the average entrainment frequency for a particle in that size class:

$$q_i = m_i \frac{f_i Y_i}{D_i^2} \frac{N_i}{T} L_i \qquad (4.7)$$

where $q_i$ is transport rate for a size class, $m_i$ is the average mass of a particle in a size class, $f_i$ is the fraction of a size class in the bed material distribution, $Y_i$ is the mobile portion of a size class, $N_i$ is the mean number of times a grain is entrained during T, and $L_i$ is the mean displacement length each time a particle moves.

Wilcock (1997) observed that the equilibrium mobile fraction of any size class $Y_i$ has a log normal distribution with respect to $\tau_0/\tau_{50i}$ where $\tau_{50i}$ is the shear stress that entrains 50% of the surface grains in size class i. The value of $\tau_{50i}$ depends on the ratio of $D_i$ to $D_{50}$ and varies with the particle-size distribution of the bed material, so that it must be determined empirically for each size class in a mixture.

In contrast to Einstein's assumption of statistical equivalence within a size class (i.e., all particles of a given size have equal probability of entrainment), Wilcock (1997) observed that a fraction of the particles in any size class were mobile during flume experiments whereas others remained immobile at low transport rates. Einstein's exchange probability ($p_i$) can be equated to the product of the mobile proportion ($Y_i$) of a size class and the mean frequency of entrainment for that proportion:

$$\frac{p_i}{T_{ei}} \frac{1}{f_{bi}} = Y_i \frac{N_i}{T} \qquad (4.8)$$

Wilcock and McArdell (1997) observed that the populations of mobile and immobile grains approach steady values at a specific time scale and noted that this time scale for gravel bed streams may be longer (e.g., a few days) than the duration of high flow that are competent to move bed material. As a result, the relationship between $\tau_0$

and steady-state values of $Y_i$ observed in flume experiments may over-estimate the extent of bed material entrainment during a flood in a gravel bed stream.

### 4.3.4. Application of probabilistic models to predict the extent of bed disturbance

The three probabilistic models provide approaches for estimating the partial entrainment of a gravel bar, $PE_{bar}$, which is the fraction of the area of a gravel bar's surface entrained during a flood. Since each model is applied to individual particle size classes of a sediment, the mobile fraction of each size class must be summed over all particle size classes of a sediment to estimate the total fraction of the stream bed surface entrained during a flood. Each model was modified and applied to a gravel bar in May Creek, King County, WA. The modifications are described below.

Einstein (1950) equated the exchange probability of each size class of bed material ($p_i$) to the fraction of the area covered by grains of size i where $\tau_0 > \tau_{cr}$ at any time. The sum of the exchange probabilities for all size classes provides an "instantaneous" estimate of $PE_{bar}$. All grains within a size class have an equal probability of entrainment, so that entrainment is represented as a process of independent, random selection of particles from the bed surface. In this case, the expected value for $PE_{bar}$ is:

$$E[PE_{bar}] = 1 - \sum_i \int_o^T f_i (1 - \frac{p_i}{t_1}) e^{-(1-p_1)t} dt \qquad (4.9)$$

where $t_1$ is a unit of time, t, that is taken here as 1 second. Equation 4.9 must be integrated over the period of active bed load transport (T) during which $p_i$ varies over time with the shear stress applied by a flood. Two nominal values of T ($10^3$ and $10^4$ seconds) are used to assess the sensitivity of predicted values of $PE_{bar}$ to the duration of an applied dimensionless shear stresses.

Wilcock (1997) expressed $Y_i$ in terms of the cumulative distribution of $\tau_g/\tau_{50i}$ where $\log(\tau_g/\tau_{50i})$ is normally distributed with a mean of 0 and standard deviation of 0.2. $Y_i$ serves as the basis of calculating partial entrainment as a function of $\tau_0$*:

$$PE_{bar} = \sum_i f_i Y_i \tag{4.10}$$

The application of Equation 4.10 to May Creek deviates from Wilcock's partial entrainment model. $\tau_g$ is represented by $\tau_0$. Furthermore, the values of $\tau_{50i}$ could not be estimated for smaller size classes of bed material at the May Creek site according to the method suggested by Wilcock (1997), because these size classes are expected to be completely mobile at the reference transport condition ($\tau_{ref}$* = 0.08). Instead, an analytical function for size-selective entrainment had to be employed to estimate $\tau_{50i}$ over the range of particle sizes at the May Creek site.

For size selective transport, $\tau_{50i}$ is a function of $D_i$. Wilcock (1997) observed that $\tau_{50i} \propto D_i^{0.38}$ for bed load transport data collected at Oak Creek, OR. The exponent relating $D_i$ to $\tau_{50i}$ may vary with the particle-size distribution of a stream bed, but the value of 0.38 is adopted here to estimate $\tau_{50i}$ at the May Creek site according to:

$$\tau_{50i} = \tau_{ref}^{*}\left(\frac{D_i}{D_{50}}\right)^{0.38}(\gamma_{sediment} - \gamma_{water})D_{50} \tag{4.11}$$

Gessler (1970) estimated the cumulative probability that a particle in a given size class would remain on the bed surface ($PS_i$) during the development of a surface armor is equal to the cumulative distribution function of $\tau_0/\tau_{50i}$ assuming $\tau_0/\tau_{50i}$ has a normal distribution with a mean value of 1 and standard deviation of 0.6. Partial entrainment of gravel bar can be estimated using $PS_i$ as:

$$PE_{bar} = 1 - \sum_{i} PS_i \tag{4.12}$$

where the values of $\tau_{50i}$, used to estimate $PS_i$, are calculated with Equation 4.11.

There are a number of sources of uncertainty regarding the prediction of the spatial extent of bed material entrainment using these probabilistic models. The estimate of partial entrainment of a stream bed will reflect the cumulative error of estimates of entrainment for individual size classes as well as assumptions about whether all of the particles in a size class have the same critical shear stress (e.g., Einstein, 1950) or distributed values (e.g., Gessler, 1970 and Wilcock, 1997). The general conditions for applying these models (i.e., steady state transport for Einstein and Wilcock or a degrading bed for Gessler) may not be typically of most floods in gravel bed streams. Furthermore, the particular particle-size distribution of a sediment may influence the relative transport rates of individual size fractions as well as the partial transport function for each size fraction. Results from bed tag experiments are used to evaluate the application of the models to the prediction of the partial entrainment of a stream bed.

## 4.4. Bed tag experiments

The partial entrainment of seven gravel bars was observed in field experiments conducted in three streams using bed tags. Bed tags are metal washers place into the stream bed between particles. They are dislodged when the surface particles move, thus providing a record of the location of bed material entrainment. The results are used to test whether bed material entrainment can be represented as a process of independent, random selection of grains from the bed where the probability of entrainment is uniformly distributed at the scale of gravel bars. A relationship was developed between partial entrainment of the gravel bars and the peak dimensionless shear stress during floods. Changes over time in the relationship between partial entrainment and the applied shear stress are examined to assess the influence of a flood's magnitude on the strength of the bed surface after the flood.

### 4.4.1. Experimental sites

The field experiments were conducted at seven sites in three streams in the Puget Lowland, Washington: Jenkins, May, and Swamp creeks. The locations of the streams are shown in Figures 4.1a, b, and c. The basins draining to these creeks have intermediate levels of urban development but contrasting physiographic conditions.

Jenkins Creek drains a 37-$km^2$ basin formed by a plateau with wetlands and lakes. It has a total relief of 110 m. The surficial geology of the basin is predominately permeable, glacial outwash deposits. Jenkins Creek had a mean discharge rate of 1.1 $m^3/s$ during water years (WY) 1989 to 1998 at the King County gage (26A) near its mouth.

May Creek drains a 32-$km^2$ basin that includes mountain headwaters with a total relief of 490 m. The surficial geology is largely glacial till and bedrock. May Creek had a mean discharge rate of 0.7 $m^3/s$ during WY 1989 to 1998 at the King County gage (37A) near its mouth.

Swamp Creek drains a 59-$km^2$ basin. The topography of the Swamp Creek basin is similar to Jenkins Creek with a total relief 195 m and many lakes and wetlands. The Swamp Creek basin, however, has more glacial till than Jenkins Creek. The only active stream gage in Swamp Creek is operated by Snohomish County at a location where the drainage area is 25 $km^2$. The mean discharge rate for Swamp Creek at this gage was 0.4 $m^3/s$ during WY 1989 to 1998. Downstream at the USGS gage (12127100) near its mouth, the mean discharge rate for Swamp Creek was 1.0 $m^3/s$ during WY 1980 to 1989.

All of the sites are in straight sections of pool-riffle or plane-bed reaches (Montgomery and Buffington, 1997) with mid-channel or transverse gravel bars (Church and Jones, 1982) where the particle-size distributions of bed material are relatively homogeneous and hydraulic conditions are relatively uniform. Maps of the reaches comprising each bar and longitudinal profiles along the channel's thalweg are shown in Figures 4.2 to 4.4. Physical characteristics of each reach are listed in Table 4.1. The bar length was measured from deepest point of the pools upstream and downstream of the bar. Bar amplitude was measured as the maximum height of the bar relative to a line drawn between these points.

Two gravel bars were monitored in Jenkins Creek. The bars are 1 km apart without any major intervening tributaries. Jenkins Creek A is a mid-channel bar in a straight reach that has a consistently low gradient of ~0.004 at low and intermediate stages (Figure 4.2a). The water surface slope in the reach declined to ~ 0.001 during the largest observed floods. The bar is located the middle of the channel with lateral pools at the base of the channel banks. Small logs lying in the channel, along with brush at high stages, contribute to flow resistance in the reach. There is a narrow (~ 2m wide) floodplain along the right bank.

Jenkins Creek B is a transverse bar in a steeper reach. The water surface slope varied from 0.016 at low flow to 0.008 at higher stages (Figure 4.2b). The bar has coarser gravel and lower amplitude than Jenkins A. It extends across the width of the channel. Brush along the right bank contributes to flow resistance at higher stages. There is a long-radius meander bend upstream of this riffle and a concrete box culvert downstream. The meander bend does not generate a strong secondary (cross-stream) flow pattern. The culvert exerts a hydraulic control at higher stages, creating a backwater that extends over the downstream end of the bar.

Three gravel bars were monitored in May Creek. May Creek Z is a transverse bar in an incised reach (Figure 4.3a). The water surface slope varied from 0.016 at lower stages to 0.011 at higher stages. No over-bank flow was observed during this investigation at this site. Mid-reach, where the channel banks expand, a narrow (1 m) floodplain has been deposited along the left side of the channel. Channel-spanning logs are located at the upstream and downstream ends of the bar, but they are suspended over the channel and above the maximum observed stage during this investigation.

May A is a transverse gravel bar with a step, foreset slope leading to a pool formed at a meander bend (Figure 4.3b). The upstream end of the bar is ill defined in a plane-bed reach. The water surface slope in May A varies from 0.011 at low stages to 0.006 at high stages due to the backwater created by the downstream bend. A large cottonwood log spans the channel downstream of the bar at the apex of the bend. There are two logs that project no more than 1 m into the channel from the bank in this reach.

May B is a transverse bar (Figure 4.3c) located 200 m downstream of May A. The water surface slope in this reach varies from 0.008 at low stages to 0.016 at high stages. There is one log extending from the right bank in this reach. The bar is forced by a downstream constriction in the channel where a large cottonwood tree reinforces the left bank.

Both May Creek A and B are also located in an incised channel so that the stream rarely flows over its banks. The channel meanders through its valley in this reach, but flow across the bars is approximately uniform and parallel to the channel banks with little cross-stream current.

Two gravel bars were monitored in Swamp Creek. Swamp Creek A is a mid-channel bar located 100 m upstream of the active stream gage. The channel is straight with uniform width, a nearly plane bed, and low amplitude bars (Figure 4.4a). The water surface flow varies from 0.003 at lower stages to 0.012 at higher stages. Channel banks are nearly vertical with narrow floodplains extending no more than 1 m from the banks before meeting steep valley walls. The reach runs along side an interstate highway. There is rip rap in the channel and along the banks in places. The bar rises only approximately 10 cm above the reach-average bed elevation and is formed of poorly-sorted material with very large cobbles (> 0.3 m diameter) lying beneath the gravel surface layer.

Swamp Creek B is a mid-channel bar located near the mouth of the stream in a reach that was re-constructed ca. 1996 (Figure 4.4b). The water surface slope varies from 0.002 at lower stages to 0.007 at higher stages. There is an expansive floodplain with side channels. Features of the reconstructed channel include a low levee along its right bank and logs projecting from the stream banks to act as flow deflectors. The bar is located 20 m downstream of a bend with a log projecting from its outer bank. The levee forms the right bank along the bar; its top is approximately 1.2 m above the thalweg. There is a floodplain with dense brush and trees (cottonwood and red alder) along the left bank of the channel.

The particle-size distributions of the bar surfaces were estimated using Wolman (1954) pebble counts. Each count included 100 particles plucked from the channel surface from an area approximately 5 m long and extending from bank-to-bank. Two to five pebble counts were conducted on each bar where bed tags were located. Pebble counts were conducted during summer. Table 4.1 lists the 10$^{th}$ ($D_{10}$), 50$^{th}$ ($D_{50}$), and 90$^{th}$ ($D_{90}$) percentiles of the particle-size distribution for the surface material of each gravel bar. A nonparametric 95% confidence interval around the $D_{50}$ was estimated using the 40$^{th}$ and 60$^{th}$ percentiles of the particle size distribution (Helsel and Hirsch, 1993, p. 70).

### 4.4.2. Bed tag experiments

Patterns of bed material entrainment were observed using arrays of bed tags placed at each field site. Bed tags are steel washers (38 mm diameter, 2 mm thick), with a short length (<10 cm ) of plastic flagging, inserted vertically between the particles forming the surface layer of the gravel bars. Each washer is pushed down between particles until its top is flush with the point of contact of the particles as shown in Figure 4.5. Placed in this manner, tags did not induce local scour and remained immobile unless the particles forming the surface of the bed were entrained.

At each site, the tags were placed at 0.5-m intervals across the stream channel in a series of rows across the bars. The locations of the rows are shown in maps of each reach (Figures 4.2 - 4.4). The bars had between 45 and 103 tags placed in 5 to 8 rows (Table 4.2). A 30-cm spike was driven into the left stream bank at each row and the first tag was located 1 m from the spike, assuring consistent tag locations over time. The rows spanned the stream bed from bank to bank except at the right end of Jenkins A, Row 5, where the bed material was silt and fine organic debris rather than gravel.

In addition to the five rows of bed tags on the May A gravel bar (Rows 5, 6, 7, 8, 10), seven additions rows of bed tags (Rows 1, 2, 3, 4, 9, 11, and 12) were placed in the May A reach during the first field season (Dec 1997-March 1998) to identify patterns of bed material entrainment over a range of spatial scales and a variety of channel forms. The observations of tags in these seven rows are excluded from the bar-scale results; they are described in a separate section on intra-reach patterns of bed material entrainment and stability.

To test of the reliability of the bed tag design, U-shaped wires with plastic flagging were inserted upside-down into the stream bed next to each bed tag in Jenkins A and May B. The two types of tags produced identical results (i.e., present or missing) at every location during two trial periods. The comparison demonstrated that the weight of the tag and drag on the plastic flagging are unlikely to influence the results as long as the tags are flush with the bed surface and do not extend below the largest particles forming the bed surface.

Bed tags are not an appropriate method for monitoring bed material entrainment under some circumstances. They do not indicate entrainment of unconstrained particles (i.e., clasts resting on top of the bed with no lateral points of contact), particularly for grains much smaller than a tag. When tags were placed in patches of fine gravel and sand (e.g., the left side of Rows 3 and 4 in the May A reach upstream of the bar), they would not move if only the grains forming the surface were entrained. However, none of the bars had fine-textured patches. Conversely, a tag placed next to a boulder might be dislodged even if the boulder was immobile. Large boulders were located at two points where tags would have been placed at Swamp A and B. Rather than placing tags at these boulders, the boulders were observed to be stable throughout the experiments. These observations are included in the results.

The bed tags were placed in gravel bars and inventoried during three periods: Oct 1997-March 1998; Oct 1998-March 1999, and Oct 1999-Dec 1999. The dates of inventories at each bar are provided in Table 4.2. Bed tags were inventoried at each of the sites periodically (every 1 to 4 weeks during the wet seasons) particularly after significant rainfall. The position (row number and distance from left bank) of each missing tag was recorded and the tag was replaced. In a few cases, a tag appeared to be missing but was actually buried. A buried tag was usually uncovered when a new tag was placed at that position, though occasionally a buried tag was found later in the season. In these cases, the last recorded instance when the tag was missing was revised to indicate that the tag had been present.

### 4.4.3. Hydraulic conditions at field sites

Each bar was surveyed using a level, tape, and stadia rod. The peak stage was recorded at two crest-stage gages along each gravel bar. The crest stage recorders were constructed from steel rods ("rebar") driven into the stream bed near the bank. Hook-and-loop fabric tape (Velcro®) was fastened along the exposed rod such that debris suspended in the stream flow (e.g., fine sediment, particulate organic material, leaves) would collect in the hooks and loops leaving an easily identified high-water mark. At each bar, the gages were separated by 10 to 20 m. The maximum error in water surface slopes is estimated to be ± 0.001 based on 0.5 cm precision in stage measurements and a 10 m distance between gage in a reaches.

The peak total boundary shear stress, $\tau_0$, was calculated for each period between inventories assuming uniform flow across the bar (Equation 4.2). The hydraulic radius for each flood peak was calculated as the wetted cross-sectional area divided by the wetted perimeter at the surveyed section using the maximum recorded stage. The energy gradient was estimated using the water surface slope between the two gages, which assumes the velocity of the current was constant along the bar. The water surface slope varied at all sites with stage (Table 4.1). A stage-slope relationship was developed at each site and used to estimate the energy gradient in shear stress calculations.

Dimensionless shear stress ($\tau_0$*) was calculated using Equation 4.3 where $\gamma_{sediment}$ was 2700 kg/m$^3$, $\tau_0$ was total boundary shear stress, and $D_{50}$ was the median of the particle-size distribution of the surface material of the bar. The specific weight of the sediment is based on the average value for samples collected from each creek. Given the potential for bias and error when using a pebble count to sample surface bed material (e.g., Kellerhals and Bray, 1971; Hey and Thorne, 1983; Diplas and Sutherland, 1988; Wolcott and Church, 1991), the average percentage errors in $D_{50}$ for all sites corresponding to the lower and upper bounds of the 95% confidence interval were propagated using Equation 4.3 to estimate a 95% confidence interval for $\tau_0$*.

### 4.4.4. Results of field experiments

The bed tags were inventoried on 103 occasions at the seven gravel bars from October 1997 to December 1999. Table 4.2 provides the date of each inventory and the fraction of tags missing from each bar ($PE_{bar}$). Figure 4.6 shows the observed values of $PE_{bar}$ and stream hydrographs for WY1998 and 1999. Generally, bed tags were inventoried after a single flood peak capable of entraining bed material. However, there were three flood peaks in Swamp Creek between the bed tag inventories on 16 November 1998 and 15 December 1998. At each bar, there was at least one inventory when no tags had moved since the previous inventory, providing an estimate of the maximum value of $\tau_0$* at each site when the bed was largely stable.

Bed tags provide information about the stability and movement of bed material at the locations where they are placed in the stream bed. Inferences about entrainment over larger areas of the stream bed have an uncertainty associated with the size of the sample (i.e., number of tags) used at a site and the spatial distribution of the probability of entrainment. The sampling error associated with bed tag inventories can be estimated using the binomial distribution given a spatially uniform probability of entrainment. Furthermore, the assumption of a spatially uniform probability of entrainment can be tested using the binomial distribution.

For a given area of stream bed with a spatially uniform and independent probability of entrainment, p, the likelihood of observing m tags missing after a storm is represented by the term $n!/(m!(n-m)!)p^m(1-p)^{n-m}$ of the binomial distribution where n is the total number of tags observed. Bar-scale sampling error is represented by the difference between p and m/n, which is equal to the observed value of $PE_{bar}$. A 95% confidence interval was constructed for each bar by calculating the cumulative binomial distribution function for a given probability and identifying the numbers of tags missing, m, where the cumulative distribution function is 0.025 and 0.975 (i.e., the probability of observing m or fewer tags is 2.5% and 97.5% respectively).

The 95% confidence intervals for $PE_{bar}$, where n = 45 tags, are shown in Figure 4.7 for selected probabilities of entrainment, p = 0.05, 0.15, 0.25, . . . , 0.95, as horizontal error bars around open circles. The x-value of the open circle represents the "true" probability of entrainment for a flood, while the error bars should contain 95% of observed values of $PE_{bar}$. The maximum absolute sampling error occurs when the "actual" probability of entrainment is 0.50, in which case 95% of the observed values of $PE_{bar}$ are expected to be between 0.37 and 0.63 (i.e., 0.50 ±0.13). The error bars in Figure 4.7 over-represent the sampling error of the bed tag inventories since more than 45 tags were used at all of the sites except Swamp Creek A.

Bar surfaces were partially entrained over the range of $\tau_0$* from 0.026 to 0.12 (Figure 4.8). At any value of $\tau_0$*, the extent of partial entrainment of the gravel bars varied between sites and at a site over time. For example, $PE_{bar}$ ranged from 0.02 to 0.48 when $\tau_0$* ~ 0.06. Likewise, the threshold for bed material entrainment was not defined by a consistent value of $\tau_0$* over time: the maximum value of $\tau_0$* at which all tags were observed to be stable at any bar was higher than the minimum value of $\tau_0$* at which some tags moved at that bar (Table 4.3). The minimum values of $\tau_0$* for which at least one tag moved ranged from 0.025 to 0.046 among the bars. The maximum values of $\tau_0$* at which all tags were stable ranged from 0.039 to 0.55 among the bars.

There are two explanations for the lack of a clear threshold of motion for bed material at a site. First, there may have been some bed material movement even if all the tags were stable, in which case the bed surface was not stable under this range of maximum values of $\tau_0$*. Alternatively the critical value of $\tau_0$* at which bed material is initially entrained may vary over time. Changes over time in the relationship between $PE_{bar}$ and $\tau_0$* (including the value of $\tau_0$* at the threshold of entrainment) were observed and are likely to reflect changes in strength of the stream bed produced by stream flow.

The maximum value of $PE_{bar}$ for all periods ranged considerably between bars, from 0.12 for Jenkins B to 0.98 for May A. The maximum values of $PE_{bar}$ at all sites occurred during the period from 21 November to 28 November 1998. During this period, May Creek had a peak discharge rate of 12.5 $m^3/s$ (annual maximum return period ~2.5 years), Jenkins Creek had a peak discharge rate of 5.3 $m^3/s$ (annual maximum return period ~ 2.5 years), Swamp Creek had a peak discharge rate of 4.5 $m^3/s$ (annual maximum return period ~ 1.2 years).

$PE_{bar}$ during a flood can be described by a linear function of the peak $\tau_0$*:

$$PE_{bar} = 12.5\,(\tau_0 * - 0.045) \qquad (4.9)$$

The predictions of $PE_{bar}$ based on Equation 4.9 have a root mean square error of 0.099 which corresponds to an error of ~10% of a bar's surface (Figure 4.8). The vertical error bars around Equation 4.9 in Figure 4.8 correspond to a 95% confidence interval for bed tag sampling errors based on the minimum sample size of 45 tags at the Swamp Creek A gravel bar. The horizontal error bars around Equation 4.9 around Figure 4.8 represent an approximate 95% confidence interval around estimates of $\tau_0$* based on the average 95% confidence interval around estimates of $D_{50}$ at all the sites.

The difference between observed and predicted values is unlikely to be solely a result of sampling error: 54% of the observed values of $PE_{bar}$ were outside the 95% confidence intervals around the values of $PE_{bar}$ calculated from Equation 4.9. The variation of $PE_{bar}$ at a given $\tau_0$*, like the changes in threshold for initial motion, is likely to reflect changes in bed strength over time produced by stream flow. Both types of changes in bed strength are analyzed below.

## 4.5. Spatial patterns of bed material entrainment

Results of the bed tag experiments are analyzed to assess the scale at which bed material entrainment may be represented as a uniform process of random selection of particles from the bed surface. The spatial distribution of the probability of entrainment determines the spatial scale for applying a sediment transport model that has a single valued probability of entrainment, under a given set of hydraulic conditions. Spatial variation in the probability of entrainment is analyzed at intra-bar and intra-reach scales. Field results are compared to predictions based on the cumulative binomial distribution function.

### 4.5.1. Intra-bar patterns

The hypothesis of uniform probability of entrainment along a bar is tested by calculating the probability of the observed values of partial entrainment for individual rows of bed tags ($PE_{row}$) given a reach or bar average value of partial entrainment ($PE_{bar}$). If the probability of observing $PE_{row}$ given a value of $PE_{bar}$ is less than 5%, then the probability of entrainment is unlikely to be uniform along a bar.

There were 372 pairs of $PE_{bar}$ and $PE_{row}$ when $PE_{bar}$ did not equal 0 (i.e., at least one tag on the bar moved in between observations). The pairs are plotted in Figure 4.8. For each pair, the probability of observing $PE_{row}$ given $PE_{bar}$ was determined using the binomial distribution where p is the observed value of $PE_{bar}$, n is the number of tags in the row , and m is the number of tag missing from that row. The probability of observing $PE_{row}$ was calculated based on the number of tags in the row. Figure 4.8, however, shows examples of the 95% confidence intervals for n = 9 tags and n = 20 tags. Each interval is expected to contain 95% of the observed values of $PE_{row}$ for a row with the respective number of tags, if bed material entrainment can be represented as a uniform process over a bar's surface with a probability of entrainment estimated by $PE_{bar}$.

For most the pairs (341 or 91.7%), $PE_{row}$ is within the 95% confidence intervals around $PE_{bar}$. However, there were 31 instances when $PE_{row}$ was outside the confidence intervals (i.e., $PE_{row}$ differed significantly from $PE_{bar}$): the fraction of tags entrained at a row was significantly less than fraction entrained at the bar (i.e., $PE_{row}$ was below the confidence intervals for $PE_{bar}$) in 11 instances (2.9%); the fraction of tags entrained at row was significantly greater than the fraction entrained at the bar in 20 instances (5.4%). The anomalously high and low values of $PE_{row}$ were distributed among 16 rows at all of the sites except Swamp A and occurred during periods with both high and low levels of partial entrainment. Most rows had only one instance when $PE_{row}$ was significantly different than $PE_{bar}$, though six rows that had repeated instances when $PE_{row}$ was significantly different than $PE_{bar}$ (Table 4.4).

The uniformity of bed tag movement was examined at May B over the course of WY 1999. The probability that a tag at a given location was missing for j inventories is:

$$P(j) = \sum_{s} \left( \prod_{i} p_i \right) \qquad 4.10$$

where $p_i$ is the probability of an observation at the location (i.e., a tag was present or missing) for a given inventory, i, based on the bar-average probability of a tag being present or missing; and s is a sequence of i inventories when the tag was missing j times. There are $2^i$ possible sequences of observations (e.g., one possible sequence observations for 11 inventories is 0,0,1,0,1,0,0,1,0,0,0 where 0 indicates a tag was missing and 1 indicates a tag was present).

During WY 1999 there were 11 bed tag inventories at May B with values of $PE_{bar}$ ranging from 0 to 0.87. At any location on the bar, a tag may have been present during all 11 inventories or missing for as many as 10 of the inventories. The probability that a tag was missing 0, 1, 2, . . . , 10 times was calculated using Equation 4.11. The product of P(j) and the total number of tags at the bar gives the expected number of tags that were missing j times. The expected and observed distributions of the number of tags are shown in Figure 4.9. There was a 32% probability that a tag was missing 3 times from any location; 26% of the locations (22 out of 85) had a tag missing three times. In general, the observed distribution is more variable with more "stable" locations (i.e., those where tags were seldom missing) and more "unstable" locations (i.e., those where tags were frequently missing) than would be expected if the probability of bed material entrainment were uniform along the bar for each flood.

### 4.5.2. Reach-scale patterns

Rows of bed tags were placed throughout the reach around May Creek A to characterize how bed material entrainment and stability varies with respect to different channel forms. The binomial distribution is used to assess spatial variation in the probability of bed material entrainment for individual rows within the reach, which includes a bar, a pool, and a plane bed interspersed with large boulders. Preferential entrainment of rows in May Reach A is evaluated using the cumulative binomial distribution function. The cumulative binomial distribution function gives the likelihood of observing "m" tags, or fewer, missing given a uniform probability of entrainment and the total number of tags.

During the period from 1 to 9 January 1998, 45% of the tags (83 out of 185 tags) in May Reach A were missing. At both Rows 9 and 12, 65% of the tags moved. It is unlikely ($p = 0.08$) that 65% or more of the tags in any row would have moved if the probability of movement for every tag had been 0.45. Instead, bed material in rows 9 and 12 was less stable than the reach as a whole. At Row 5, only 11% of the tags moved during the period. It is unlikely ($p < 0.01$) that the probability of a tag moving in this row was equal to the reach-average probability. Instead, Row 5 was more stable than the reach as a whole.

The same patterns of preferential instability at Rows 9 and 12 and stability at Row 5 are evident from the fraction of tags stable for the longer period from December 1997 to March 1998. For all tags in the May A reach, 35% were stable throughout the winter period. Based on observations that 12% of tags were stable in Row 7 and 6% in Row 10, it is highly unlikely ($p <0.05$) that the probability a tag was stable throughout the winter at these reaches was equal to the reach-average value of 0.35. In contrast, Rows 5 and 6 had greater fractions of stable tags (72 and 56% respectively) than would be expected ($p<0.05$) if the probability of stability was equal to the reach-average value.

The least stable rows are located upstream of the crest of a transverse bar (Row 9) and at the tail end of a pool (Row 12). Sections acting as hydraulic controls under low flow conditions are immediately downstream from these rows. As a result, the current velocity is slightly less and the bed material is finer grained at these sections than for most of Reach A. As stage increases, the downstream section no longer acts as a hydraulic control; the water surface slope and current velocities increase at Rows 9 and 12 to levels representative of the reach as a whole. In contrast, the most stable rows (5 and 6) are located on a shallow foreset slope of a bar. The bed material is relatively coarse in these sections reflecting relatively high shear stress values at low and intermediate stages. The patterns of stability at these rows indicates that the local particle-size distribution of the bed surface may be set by size-selective deposition and transport during intermediate and lower flows.

## 4.6. Variation in the partial entrainment of a gravel bar at a given shear stress

The values of $PE_{bar}$ vary among floods with similar values of $\tau_0$*. The variation in $PE_{bar}$ at a given value of $\tau_0$* exceeds the 95% confidence interval representing sampling error associated with the bed tag inventories. Apparent differences between streams may be biased by measurement errors (e.g., energy gradient, particle-size distribution), but the variation in $PE_{bar}$ for a given value of $\tau_0$* is as large at some bars (e.g., Swamp B, May A and B) as it is between bars.

Since neither sampling nor measurement errors are likely to account for the variation of $PE_{bar}$ at a given value of $\tau_0$*, two other sources of variation are analyzed in this section: (1) differences in the cumulative extent of bed surface entrained between inventories with similar peak values of $\tau_0$*; and (2) stochastic changes in the structure of the bed surface (i.e., particle-size distribution and the position and arrangement of particles) that modify the stability of the bed over time.

### 4.6.1. Cumulative extent of bed surface entrainment over time

$PE_{bar}$ is a cumulative measure of the extent of bed material entrainment over a an interval of time. The value of $PE_{bar}$ will reflect both the magnitude and duration of bed load transport, but $\tau_0$* is an instantaneous measurement that does not necessarily account for the duration of sediment transport between inventories. Long duration floods, or multiple flood peaks between inventories should produce relatively high values of $PE_{bar}$ at a given shear stress $\tau_0$*. While the duration of sediment-transporting flows is not analyzed here, the cumulative extent of entrainment over multiple flood peaks at a site is analyzed here.

Inventories conducted after multiple flood peaks indicate higher values of $PE_{bar}$ relative to $\tau_0$* than inventories conducted after a single flood peak. For example, a large fraction of tags was missing ($PE_{bar}$ = 0.96) at Swamp B for the inventory on 15 December 1998 when $\tau_0$* was 0.10. In comparison, $PE_{bar}$ = 0.56 for the 16 November 1999 inventory at Swamp B when $\tau_0$* was also 0.10. These points appear as the X's in the upper right quadrant of Figure 4.7. The hydrograph for Swamp Creek (Figure 4.6) shows three distinct peaks in discharge for the period prior to the 15 December 1998 inventory whereas there was only one peak for the period prior to the 16 November 1999 inventory.

The value of $PE_{bar}$ for an inventory after multiple storm peaks can be calculated assuming independent entrainment of bed tags from flood to flood:

$$PE_{bar} = 1 - \prod_{j} (1 - PE_{bar\ j}) \qquad (4.11)$$

where 1-$PE_{bar\,j}$ is the fraction of the bar that was stable in the jth flood. If $PE_{bar\,j}$ was 0.6 for each of the three peaks, then the expected value of $PE_{bar}$ on 15 December 1998 would be 0.95 as compared to the observed value of 0.96. The agreement of the observed and calculated values suggests that the location of missing bed tags may be independent from flood to flood, and the cumulative fraction of tags entrained can be calculated according to Equation 4.9.

### 4.6.2. Stream-flow mediated changes in the strength of the bed surface

The variation in the observed values of $PE_{bar}$ for a given value of $\tau_0$* is likely a result of structural changes in the bed surface (e.g., armoring, armor breaching, particle clustering, and packing) over the duration of the bed tag experiments. Changes in bed surface at the bars are inferred by comparing $PE_{bar}$ for pairs of floods with similar magnitudes. For each pair, the cumulative binomial distribution function was applied iteratively to find the entrainment probability at which the likelihood of the observed values of $PE_{bar}$ was equal, which is approximately equal to the mean of the two values of $PE_{bar}$. When the likelihood of the two observed values of $PE_{bar}$ was less than 5% given a common entrainment probability, the values of $PE_{bar}$ for the two inventories are identified as significantly different.

For all sites, there were six pairs of inventories when $\tau_0$* was approximately equal for each inventory in a pair but the values of $PE_{bar}$ were significantly different (Table 4.5). $\tau_0$* ranged from 0.045 to 0.082 among the different pairs. In five of the pairs, $PE_{bar}$ was greater for the first inventory than for the second inventory. Other pairs of observations show a pattern of increasing bed stability over time, though the differences in $PE_{bar}$ are not statistically significant. Only one pair of inventories showed a significant decrease in stability over time when a higher value of $PE_{bar}$ was observed for the second inventory than the first.

The surface strength of the gravel bars did not always increase over time. For the 21 November 1998 inventory at May B, $PE_{bar}$ = 0.35 and $\tau_0$* = 0.058 which was consistent with a later inventory on 3 December 1998 when $PE_{bar}$ = 0.32 and $\tau_0$* = 0.060.

The values of $PE_{bar}$ for these inventories are high in comparison to all other inventories when $\tau_0* \approx 0.06$ (Figure 4.8) indicating a relatively unstable bed condition, though a lower value of partial entrainment could have been expected for the second flood. Either the first flood did not stabilize the bed or, perhaps, a large intervening flood ($\tau_0* = 0.118$) on 26 November 1998 weakened the bed.

The influence of the magnitude of previous floods on bed stability was analyzed by comparing the values of $PE_{bar}$ for groups of inventories when $\tau_0*$ had intermediate values varying only from 0.055 to 0.070. The inventories were divided into three groups based on the value of the peak dimensionless shear stress for the previous inventory ($\tau_{previous}*$). Group A (7 inventories) had the lowest magnitude previous floods ($0.062 < \tau_{previous}* < 0.068$). Group B (5 inventories) had intermediate magnitude previous floods ($0.070 < \tau_{previous}* < 0.082$). Group C (5 inventories) had the highest magnitude previous floods ($0.087 < \tau_{previous}* < 0.118$). The inventories when $\tau_0*$ ranged from 0.055 to 0.070 were selected for this analysis because the observed variation of $PE_{bar}$ was large (from 0 to 0.48) but not correlated to variation in $\tau_0*$ over this range.

The partial entrainment of a gravel bar during a flood depended not only on the magnitude of the flood but on the magnitude of previous floods as well. The average value of $PE_{bar}$ was 0.12 for Group A (previous floods with the lowest magnitudes), 0.08 for Group B (previous floods with intermediate magnitudes), and 0.20 for Group C (previous floods with highest magnitudes). The average values of $PE_{bar}$ for Groups B and C were significantly different ( $p < 0.05$ based on a one-tailed Student's t distribution). Intermediate magnitude floods ($0.070 < \tau_0* < 0.082$) strengthen the stream bed surface whereas larger floods ($\tau_0* > 0.082$) leave a weaker bed surface.

The bed tag results conform to a model where floods with intermediate peak magnitudes transport small and unconstrained particles, cluster particles, and pack down the bed surface, whereas higher magnitude floods erode such structures and transport bed material indiscriminately with respect to particle size (e.g., Gomez, 1983a; Kuhnle, 1989; Hassan and Reid, 1990; Chin et al., 1994). The marginal extent of bed material entrainment declines over a series of intermediate magnitude floods, while higher magntiude floods continue to entrain bed material. After a large flood, the critical shear stress of the bed surface is lower than it would be after a moderate flood. The duration of intermediate and higher stream flows, in addition to their magnitude, is likely to influence the strength of the stream bed (e.g., Reid and Larrone, 1995; Church et al., 1998), though it was not examined in this investigation.

The changes in bed condition were not associated with any significant change (< 5 mm) in the median diameter of stream bed material at the sites from summer-to-summer. There may have been transient changes in bed material texture between storms during the winter that were not evident from the results of pebble counts conducted during periods of lower flow (e.g. Gomez, 1983a). Moreover, structural modification of a bed surface, such as the clustering and packing of particles into stable structures on the bed surface, change its stability but may not be indicated by changes in bed material texture (Church et al., 1998).

## 4.7. Results of probabilistic sediment transport models

The sediment transport models developed by Einstein (1950), Wilcock (1997), and Gessler (1970) were modified to predict the extent of surface material entrainment and applied to the gravel bar at May Creek B for values of $\tau_0$* ranging from 0.02 to 0.12. Each of the analytical predictions indicate that partial entrainment of a stream bed surface occurs over a wider range of $\tau_0$* values than was observed for May B (Figure 4.10). The values of $PE_{bar}$ based on Wilcock (1997) agree most closely with observed values of $PE_{bar}$ over the range of floods observed at May B.

The results of the models generally agree with the relatively high observed values of $PE_{bar}$ for inventories when $\tau_0$* ranges from 0.04 to 0.07. The relatively high values of $PE_{bar}$ indicate an “unconditioned” bed surface with an abundance of mobile (i.e., small or unconstrained) particles that correspond to the conditions for each model's development. Indeed, the bed surface was initially unarmored in Gessler's (1970) experiments, an armor layer did not develop in Wilcock and McArdell's (1997) experiments because entrained sediment was recirculated, and Einstein (1950) relied on data from sand-bedded rivers which would not have been strongly armored.

In contrast, there were numerous bed tag inventories at low values of $\tau_0$* when less entrainment occurred than was predicted by any model. These inventories represent transient periods when the stream bed was exceptionally stable, as would be expected under conditions of low sediment supply and steady high flows where the bed surface was armored. The models were developed for a stream bed that is initially armored.

While there were only three inventories in May B when $\tau_0$* > 0.08, the apparent divergence of the predictions from the reported values could result from a bias in the bed tag results or in the estimates of shear stress. Since bed tags may be entrained at locations next to stable particles, the inventory results for high magnitude floods do not indicate the extent of individual stable particles. Additionally, $\tau_0$ may over-estimate $\tau_g$ during high flows at the field sites.

## 4.8. Conclusions

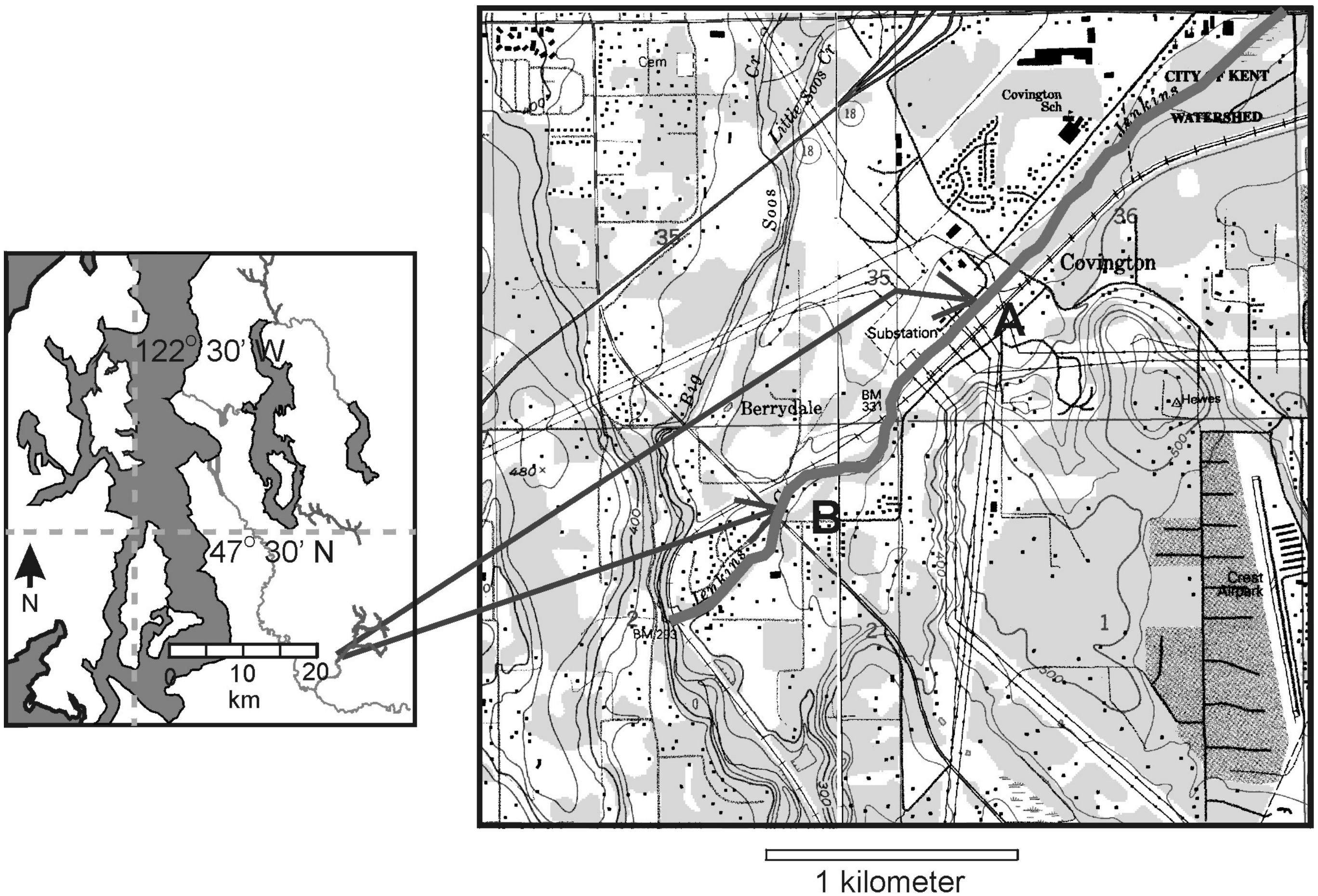
Entrainment of bed material during floods is a frequent form of disturbance in Puget Lowland gravel-bed stream ecosystems. During three winters, floods entrained at most only a portion of the mineral and organic materials forming the bed surface from widely distributed locations along the bed at seven gravel bars in three streams. The partial entrainment of a gravel bar's surface during a flood can be estimated as a linear function of peak reach average shear stress scaled by the intermediate axis of the median surface particle. Partial entrainment of a gravel bar was observed over a 3-fold range in shear stress. Half of the tags at any bar ($PE_{bar}$ = 0.50) are expected to be entrained at $\tau_0$*

in the range from 0.06 to 0.10, though $PE_{bar}$ was observed to vary from 0.01 to 0.96 over this range. Predicted values of parital entrainment, based on the linear equation $PE_{bar} = 12.5(\tau_0^* - 0.045)$, have a root mean square error of 0.099 compared to observed values of $PE_{bar}$.

The fraction of the bed surface entrained during a flood can be equated to the probability of entrainment at a point on the stream bed if the probability of entrainment is uniformly distributed. The results of bed tag experiments show that the probability of entrainment is only approximately uniform over a gravel bar. Some locations on a gravel bar are frequently more stable than the bar as a whole while others are less stable. Specific sections of a channel may be more or less stable than a bar or reach as a whole depending on their form. Higher entrainment probabilities were observed at the steep foreset slope (downstream face) of a bar (May A.10), in a section of converging flow (May B.3), or places where the channel is actively widening or migrating (May A.7 and Swamp B.4 respectively). The two rows where the probability of entrainment was repeatedly lower were located upstream of the bar crest in slightly wider sections (May A.5 and Swamp B.1). The variation in local hydraulic and geomorphic conditions at these sites is typical of many streams in the region and influence the spatial patterns of bed material entrainment during floods.

The lowest values of $\tau_0^*$ at which some bed tags were missing ranged from 0.025 to 0.046 among the sites, which were less than the maximum values of $\tau_0^*$ (0.039 to 0.055) for inventories when all the tags were stable at the sites. Variation in both the observed values of $PE_{bar}$ at a given $\tau_0^*$ and the "threshold" value of $\tau_0^*$ between a stable and partially mobile stream bed may result from a number of factors: the precision of $PE_{bar}$ estimates is not very high; $\tau_0^*$ does not account for the duration of sediment transport, including the cumulative entrainment from multiple floods; and the size and position of particles forming the surface of bed material may change over time such that the distribution of critical shear stress over the bed is not constant.

Variation between $PE_{bar}$ and $\tau_0$* over time show changes in the strength of a stream bed surface as a result of flow conditioning. $PE_{bar}$ was lower on average, indicating a stronger bed surface, in a flood when the previous flood was of intermediate magnitude ($0.07 < \tau_{previous}* < 0.085$) than when the previous flood was of higher magnitude ($\tau_{previous}* > 0.085$). Due to the influence of preceding flows on the strength of the stream bed, the probability of entrainment at a point on the bar is only approximately independent from event-to-event.




contour interval 25 feet

Figure 4.1(a): Bed tags sites in Jenkins Creek.
Source: USGS (1994) Auburn and Black Diamond, WA quadrangles, 7.5 minute series, 1:24,000 scale

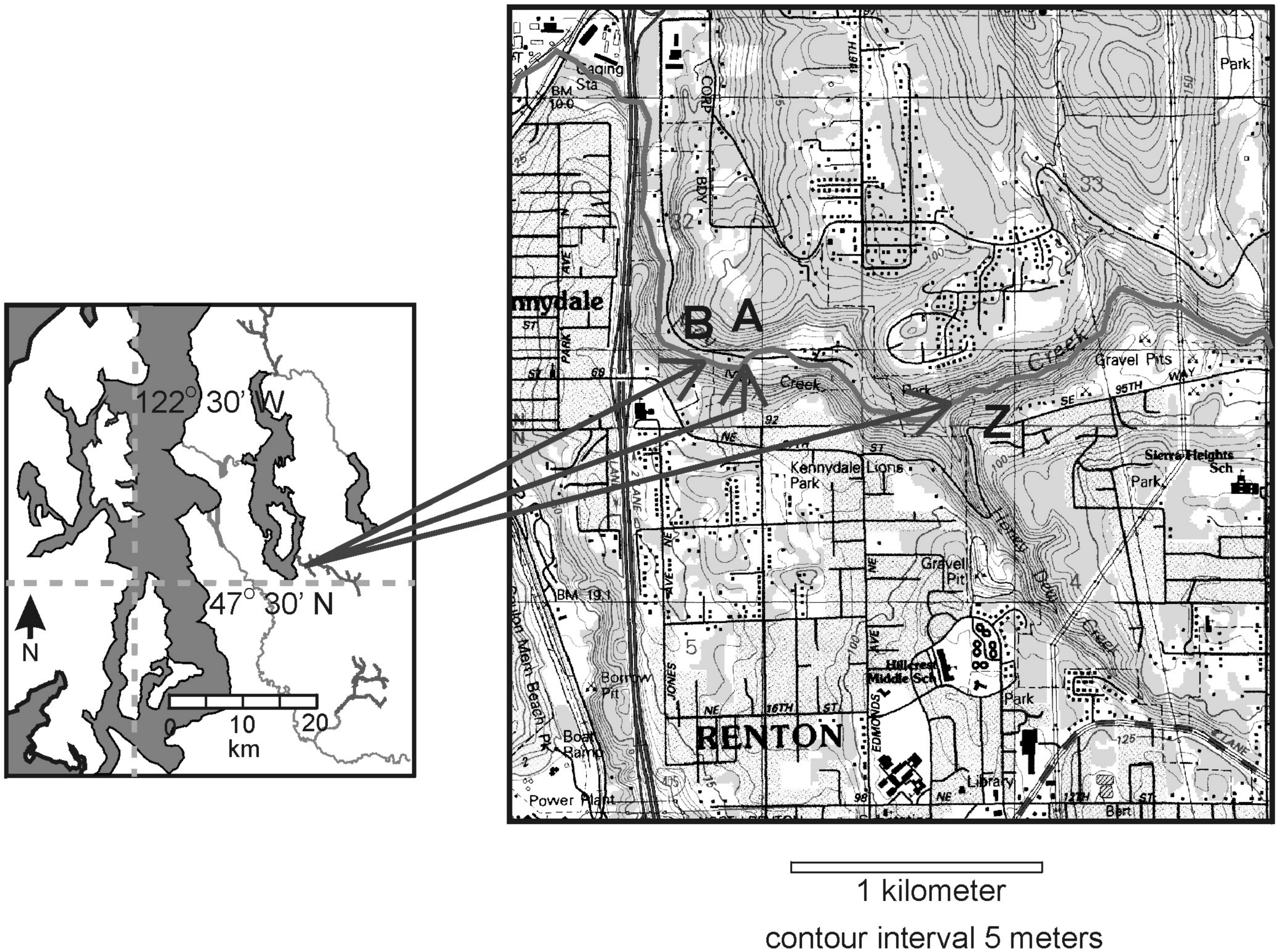



Figure 4.1(b): Bed tag sites in May Creek.
Source: USGS (1983) Bellevue South, WA quadrangle, 7.5 x 15 minute series, 1:25,000 scale topographic map

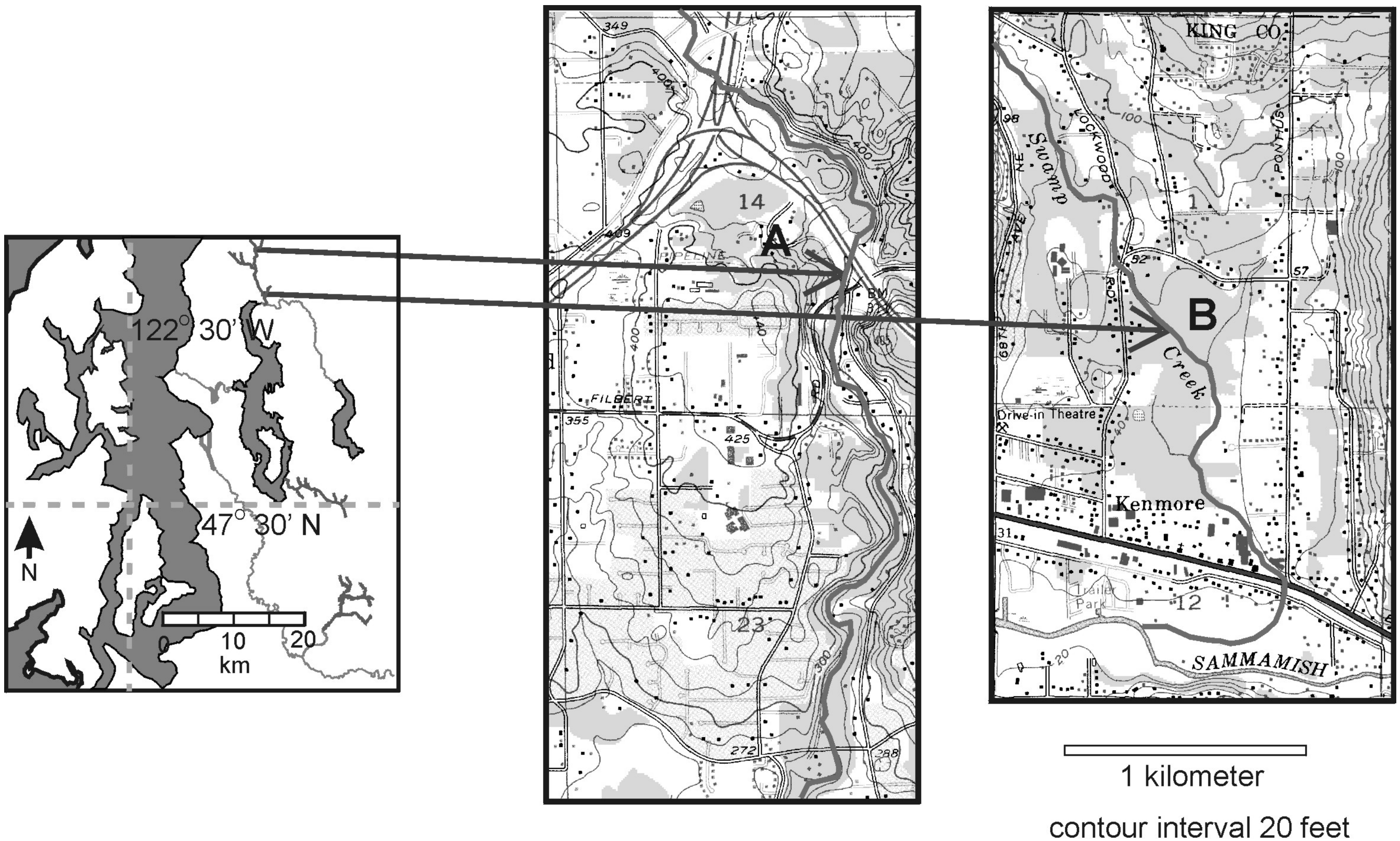



Figure 4.1(c): Bed tag sites in Swamp Creek.
Source: USGS (1981) Edmonds East, WA quadrangle, 7.5 minute series, 1: 24,000 scale

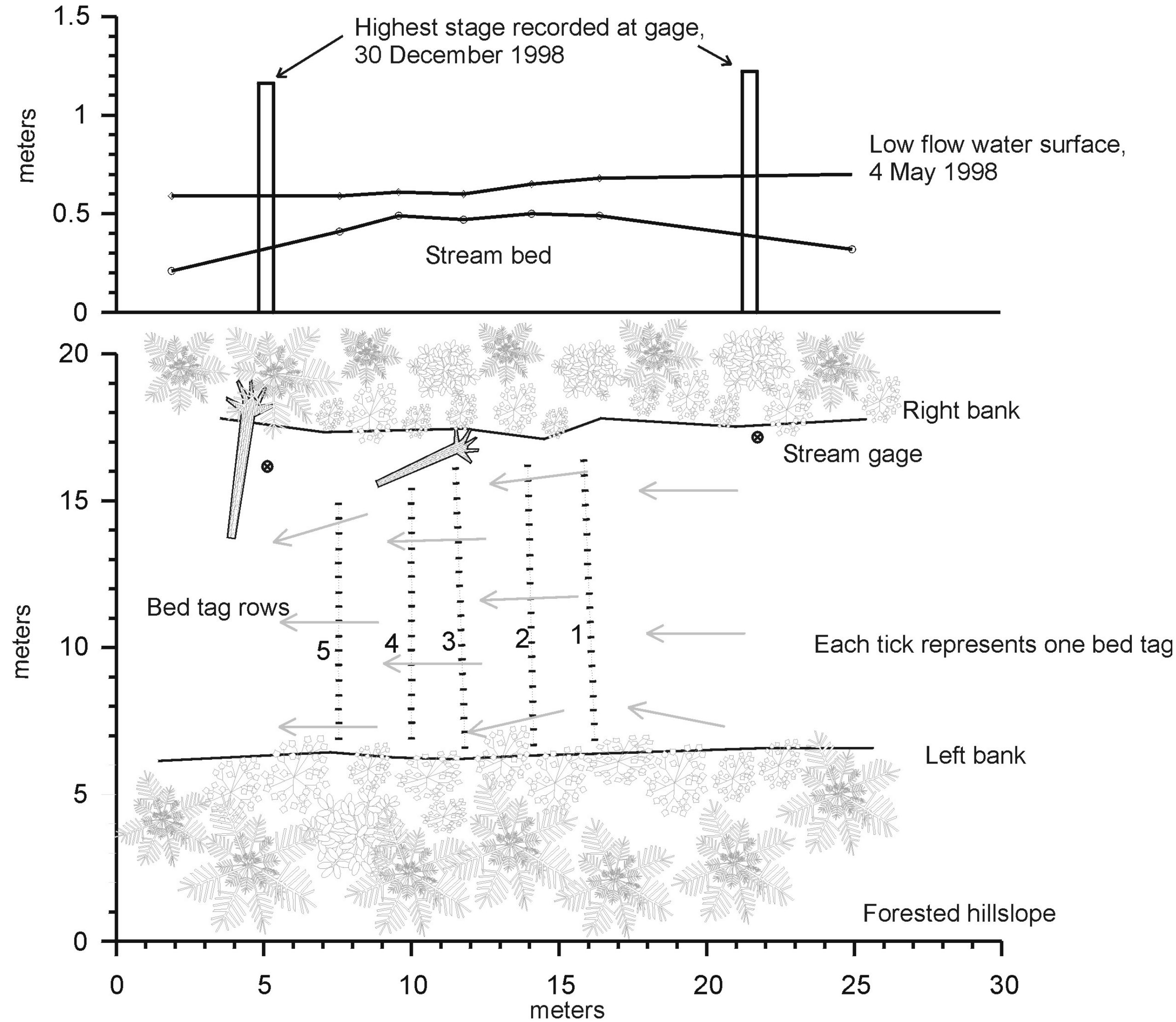



Figure 4.2a: Jenkins Creek Reach A.

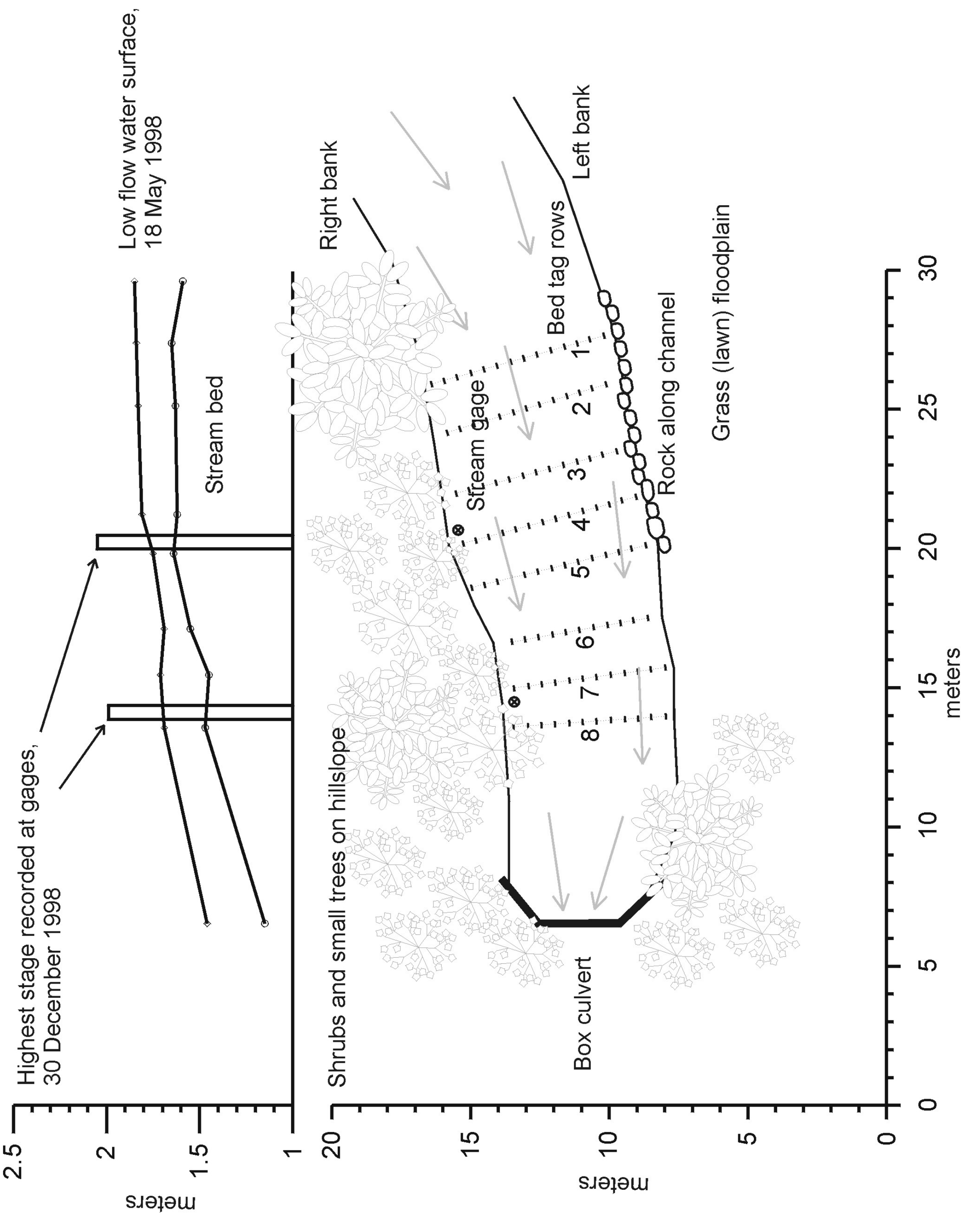



Figure 4.2b: Jenkins Creek Reach B.

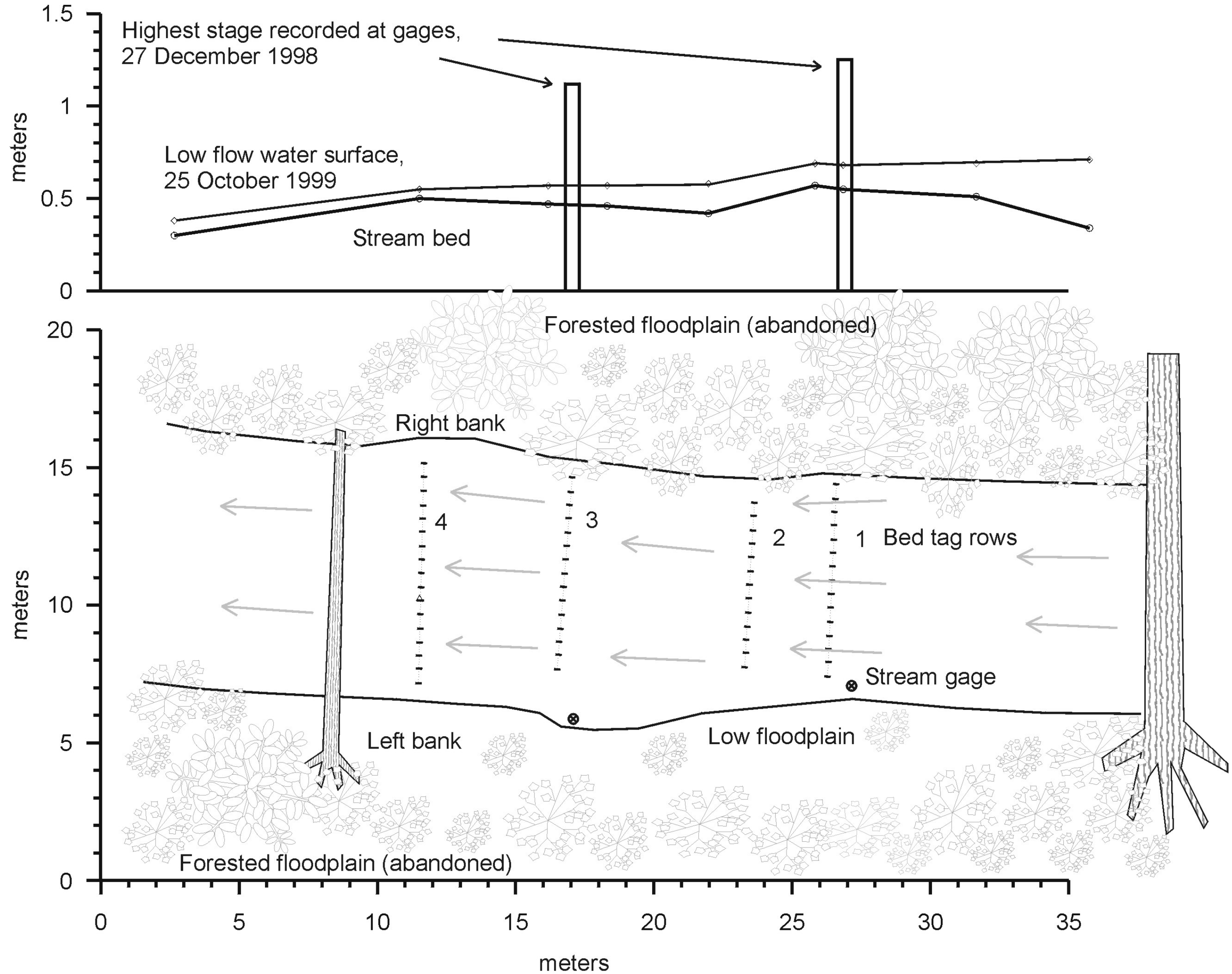



Figure 4.3a: May Creek Reach Z.

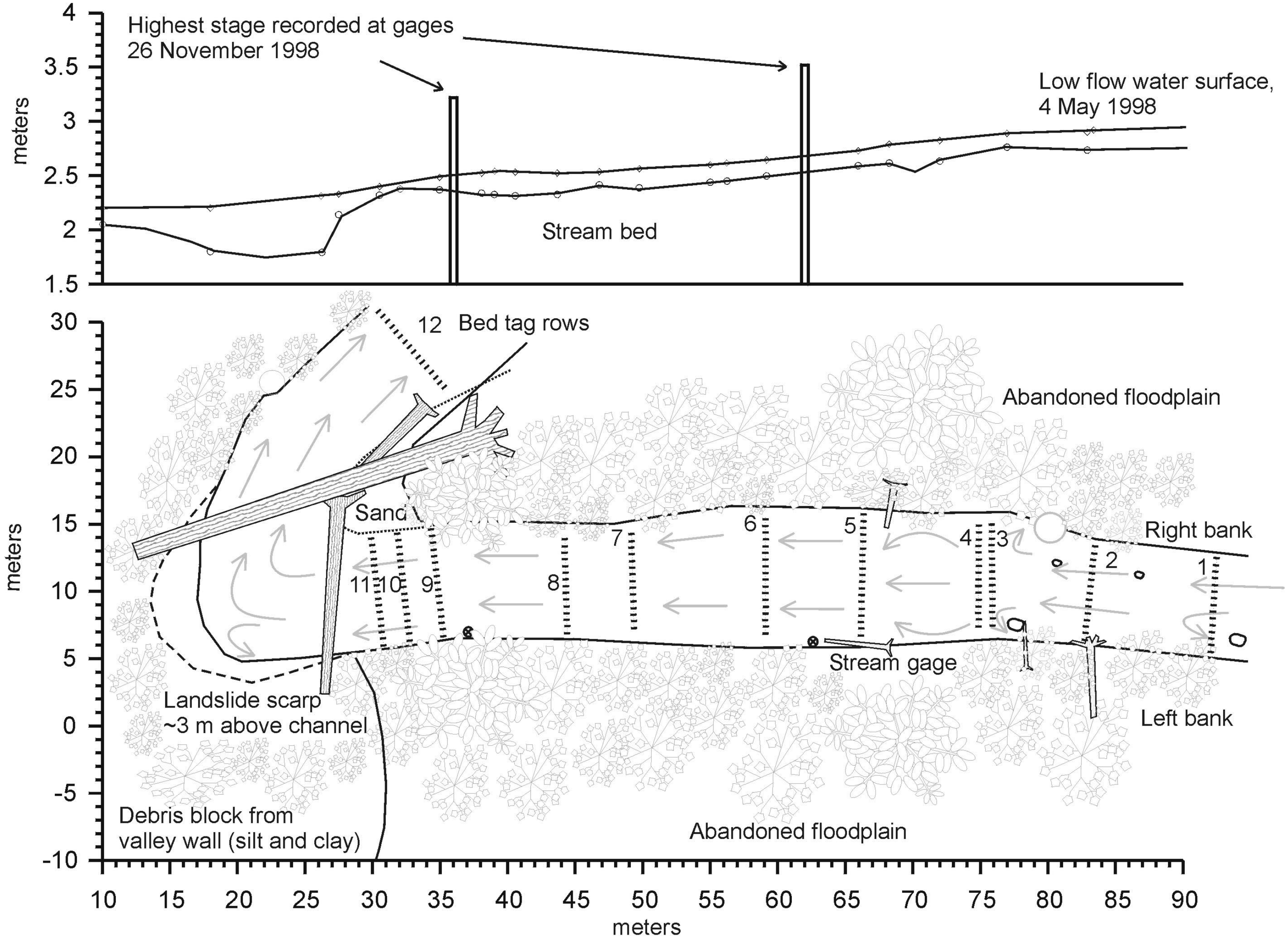



Figure 4.3b: May Creek Reach A.

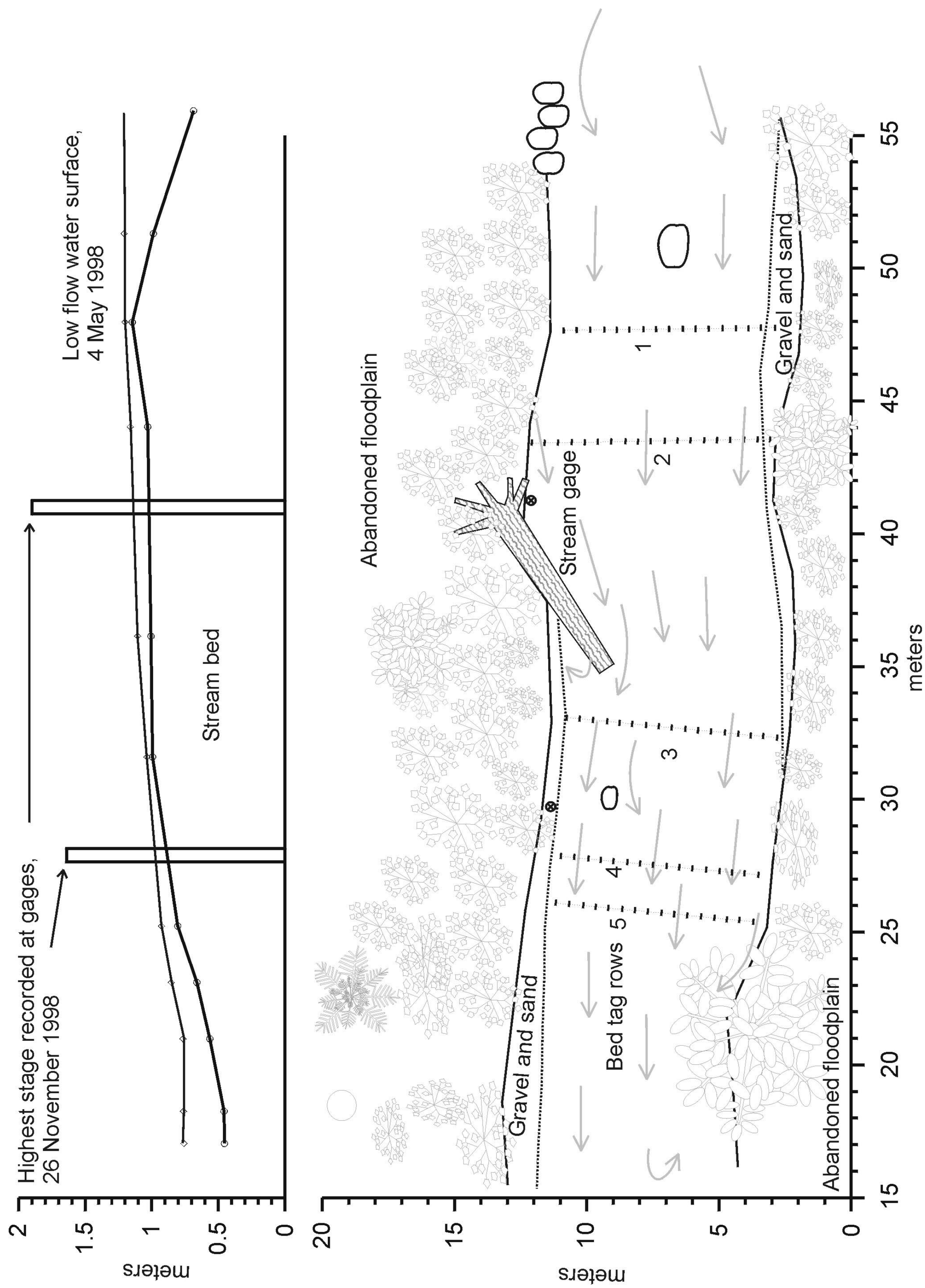



Figure 4.3c: May Creek Reach B.

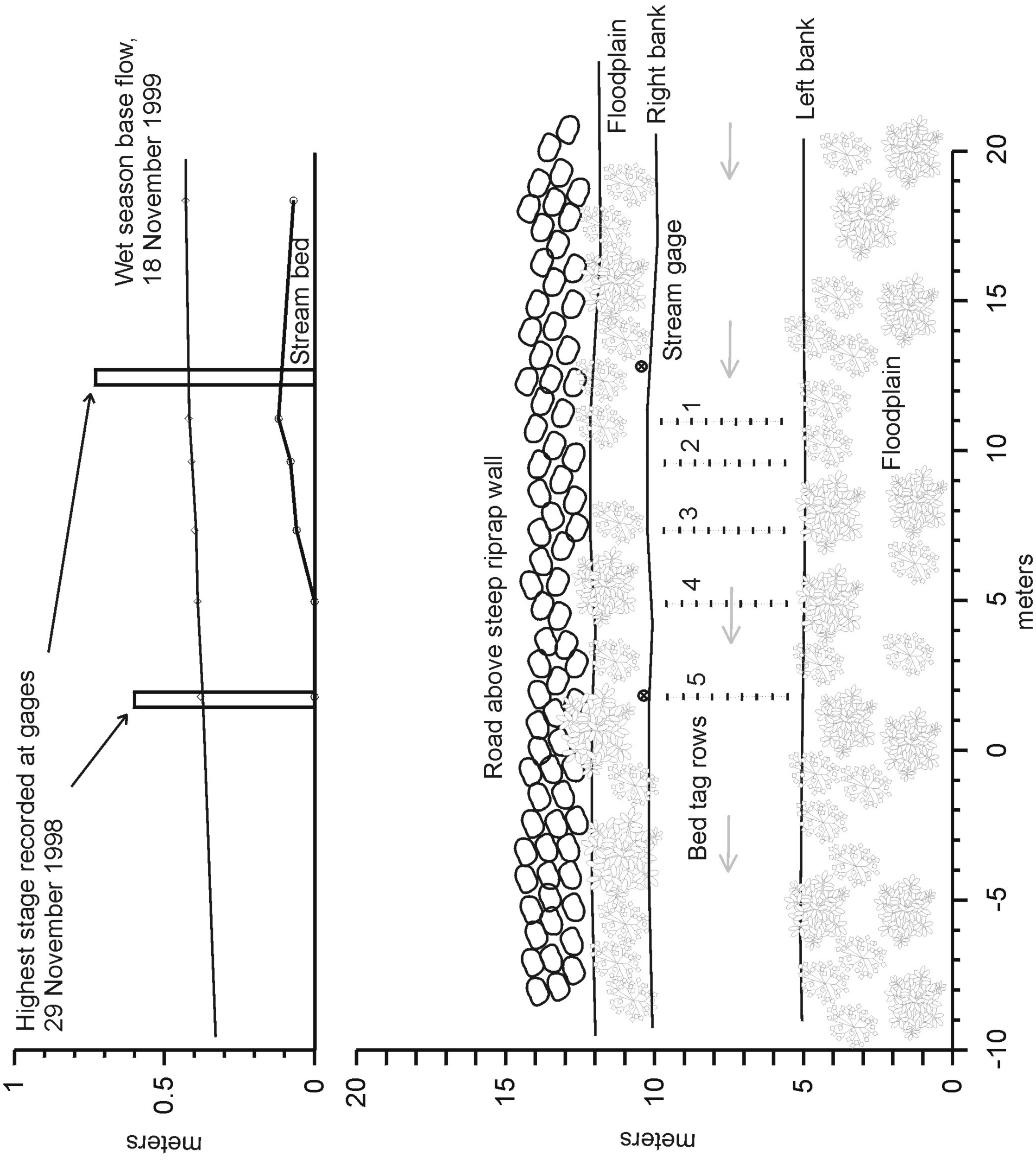



Figure 4.4a: Swamp Creek Reach A.

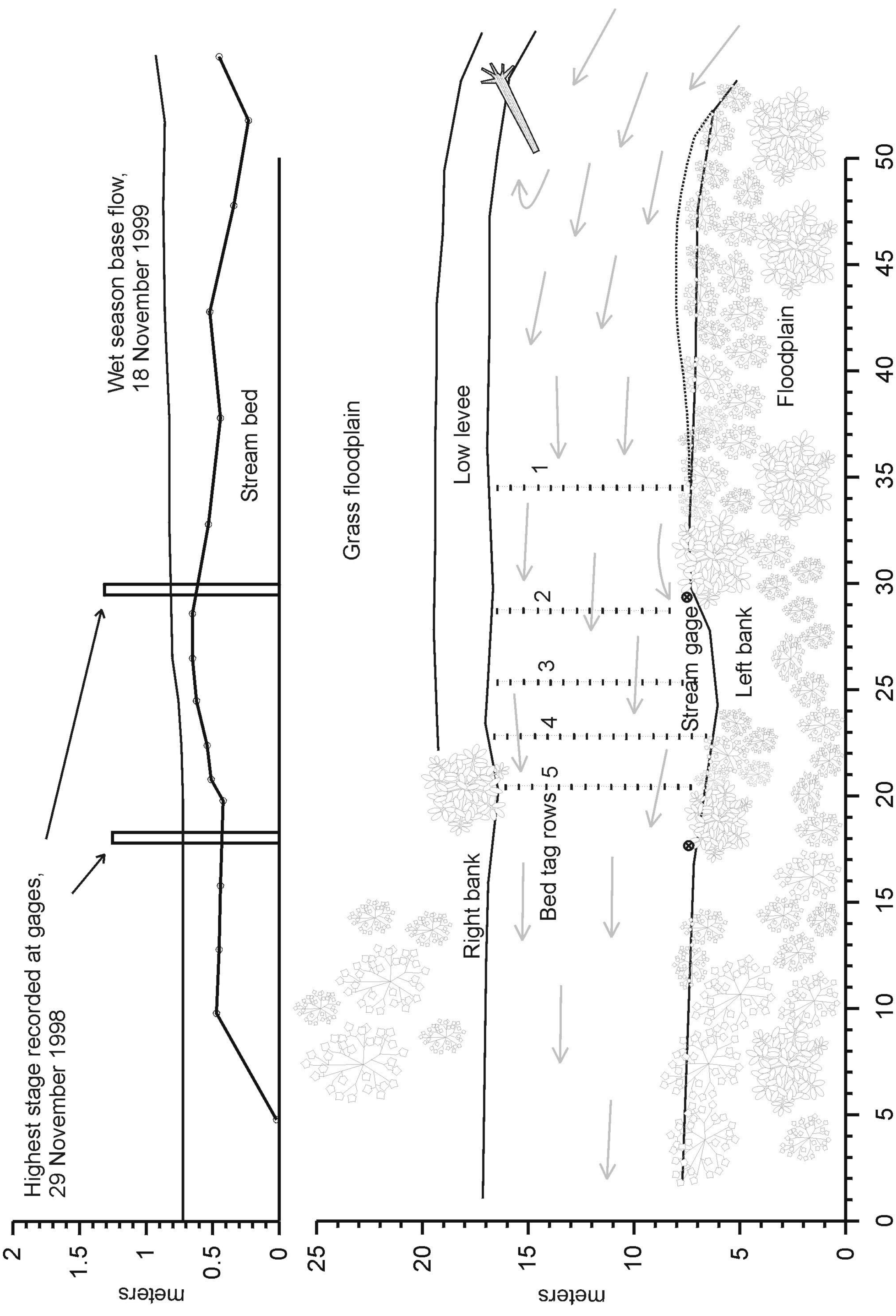



Figure 4.4b: Swamp Creek Reach B.

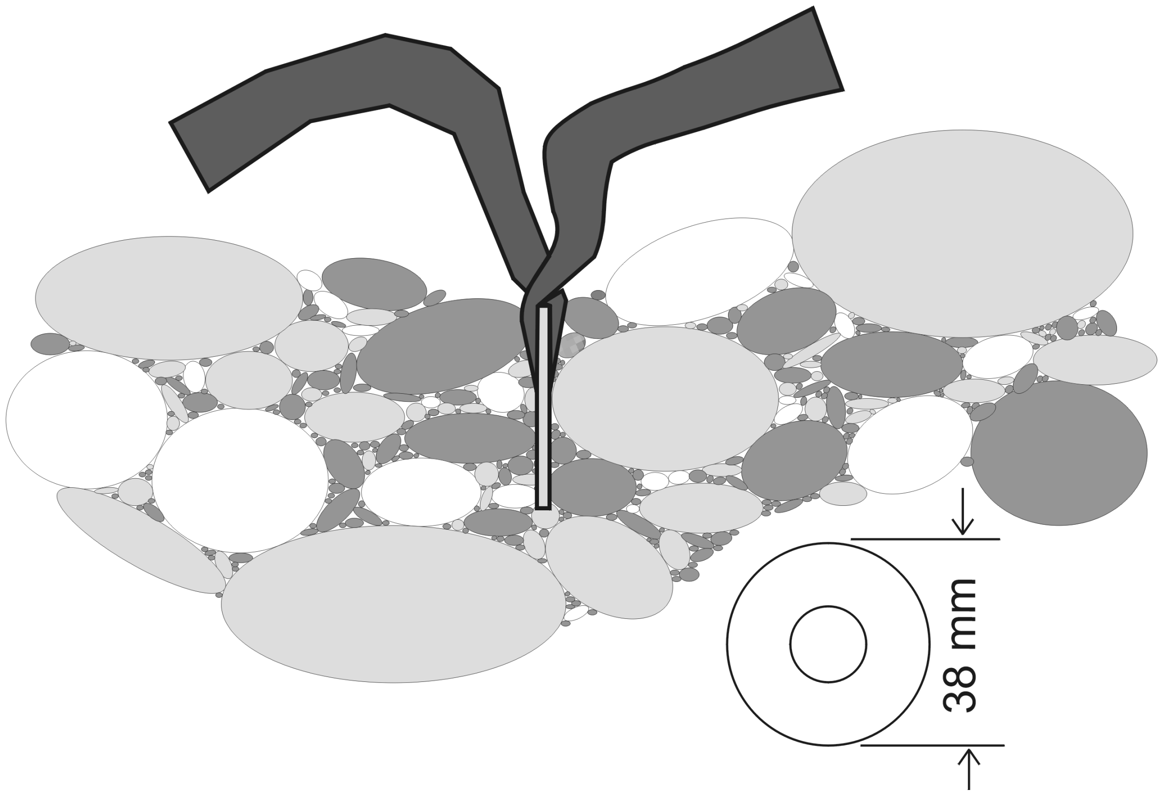
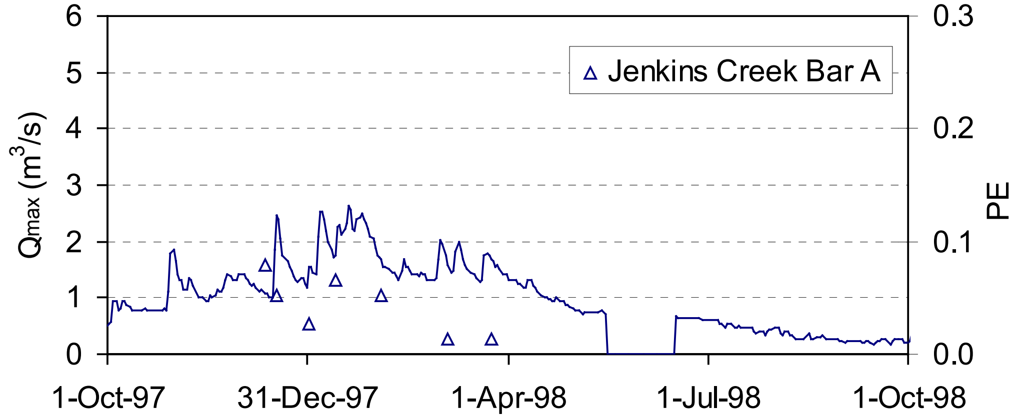
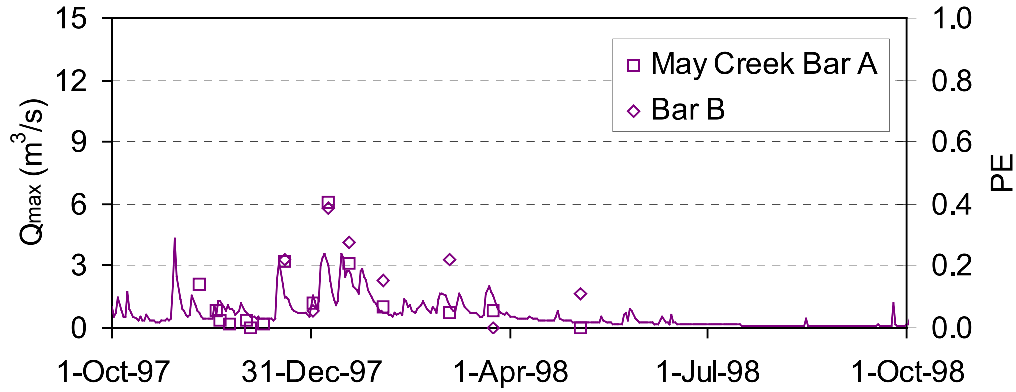



Figure 4.5: Cross-section of bed tag placement in a stream bed.

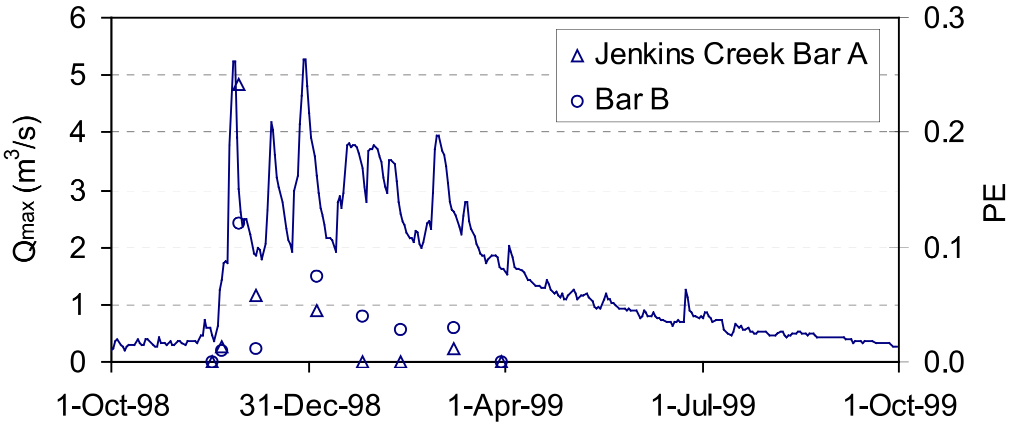



(a)

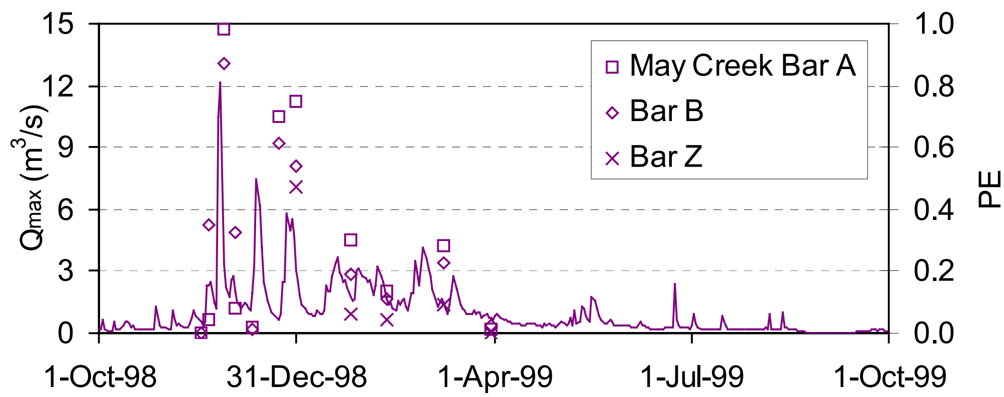



(b)

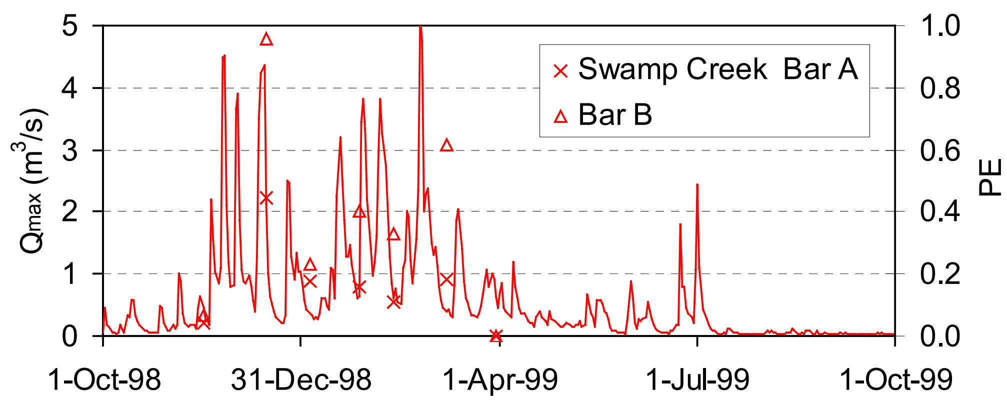
Figure 4.6: Continuous hydrographs of maximum daily discharge ($Q_{max}$) and the fraction of bed tags missing from each bar (PE) for (a) Jenkins Creek WY 1998, (b) May Creek WY 1998, Jenkins Creek WY 1999; (d) May Creek WY 1999; and (e) Swamp Creek WY 1999.




(c)




(d)




(e)

Figure 4.6 continued.

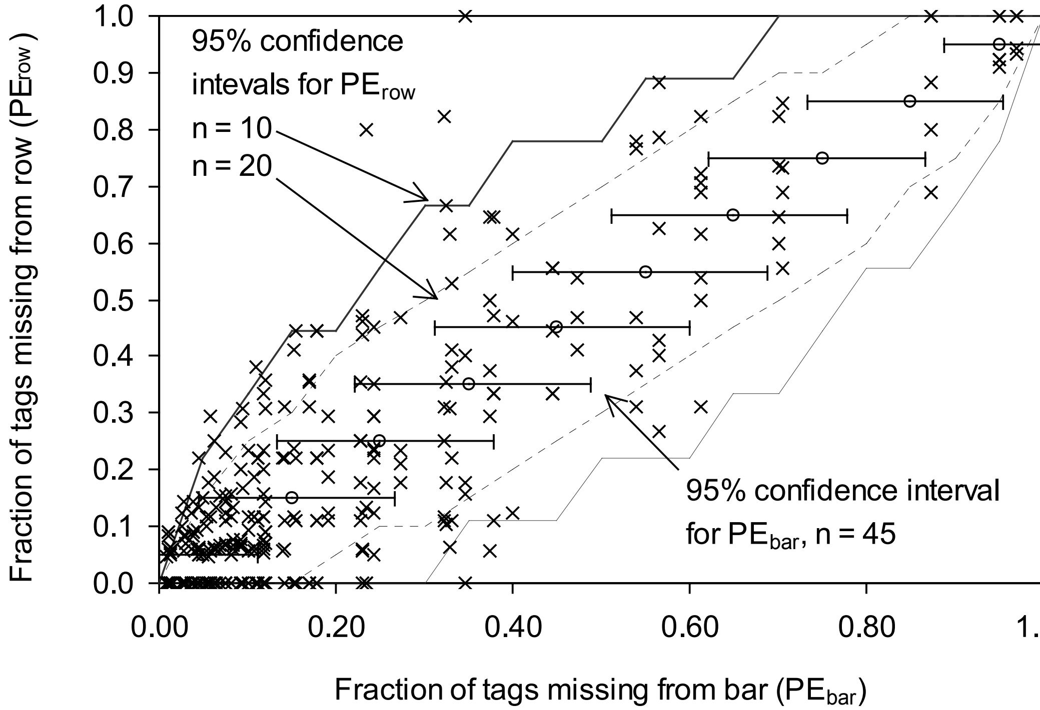



Figure 4.7: Comparison of bed tag inventory results between bars ($PE_{bar}$) and rows ($PE_{row}$) with confidence intervals based on random sampling from a binomial distribution where p = $PE_{bar}$, n = number of tags.

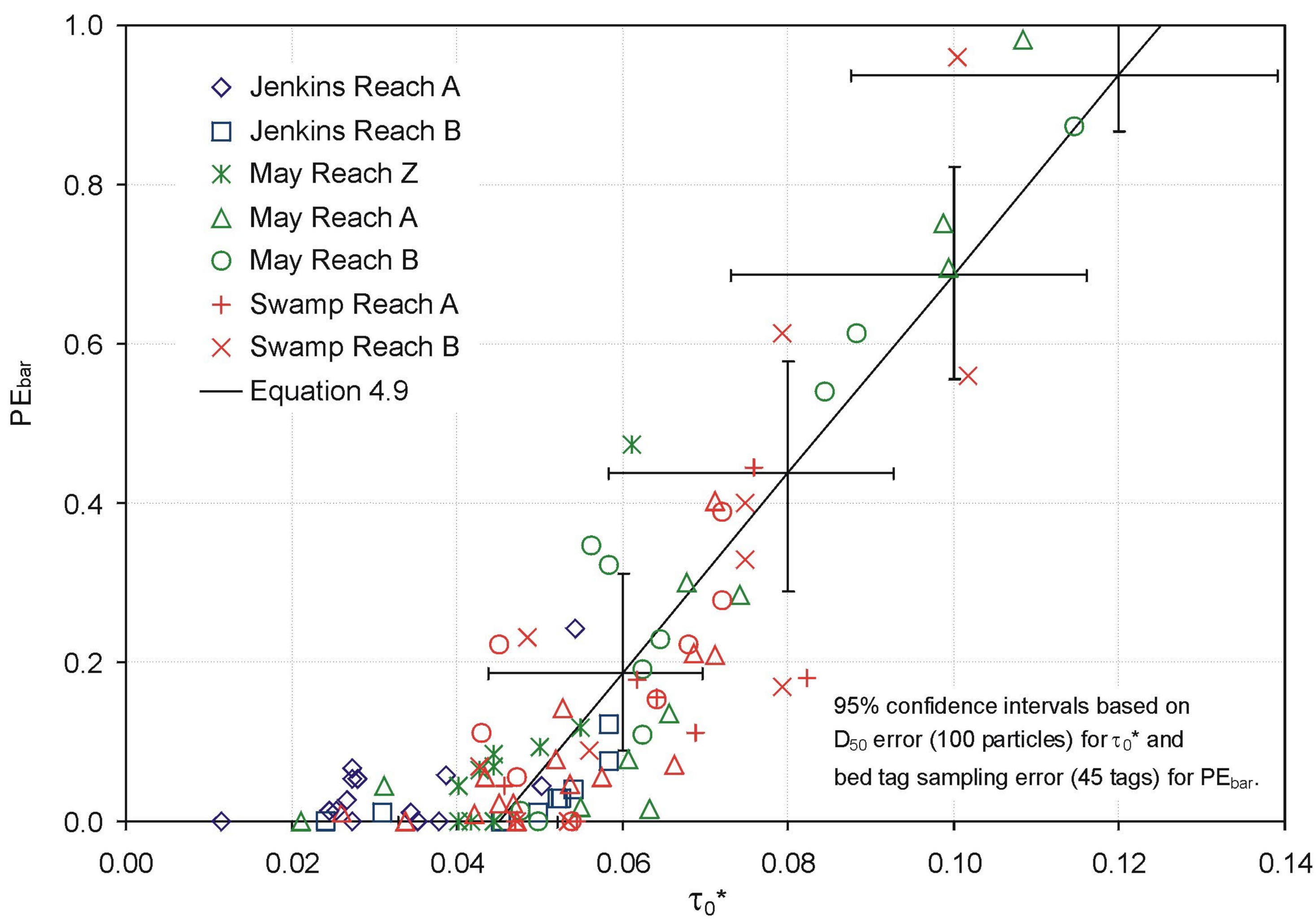



Figure 4.8: Partial entrainment ($PE_{bar}$) at seven gravel bars as a function of peak dimensionless shear stress ($\tau_0^*$).

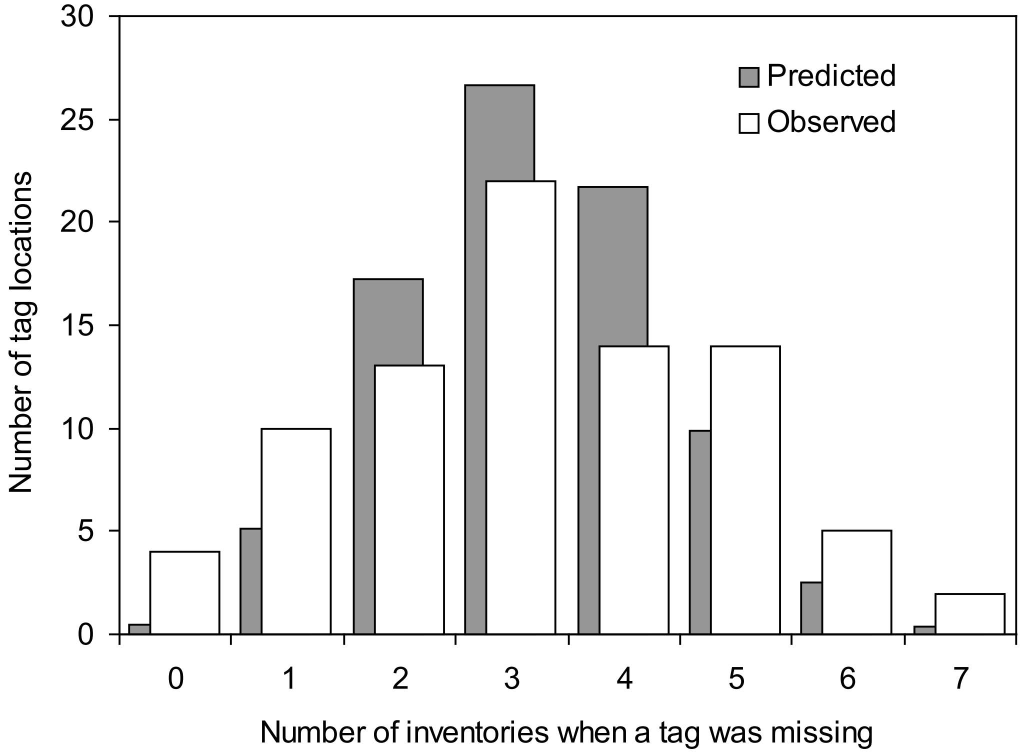



Figure 4.9: Predicted and observed distributions of the frequency that bed tags were missing from locations at May B (85 tag locations, 11 inventories).

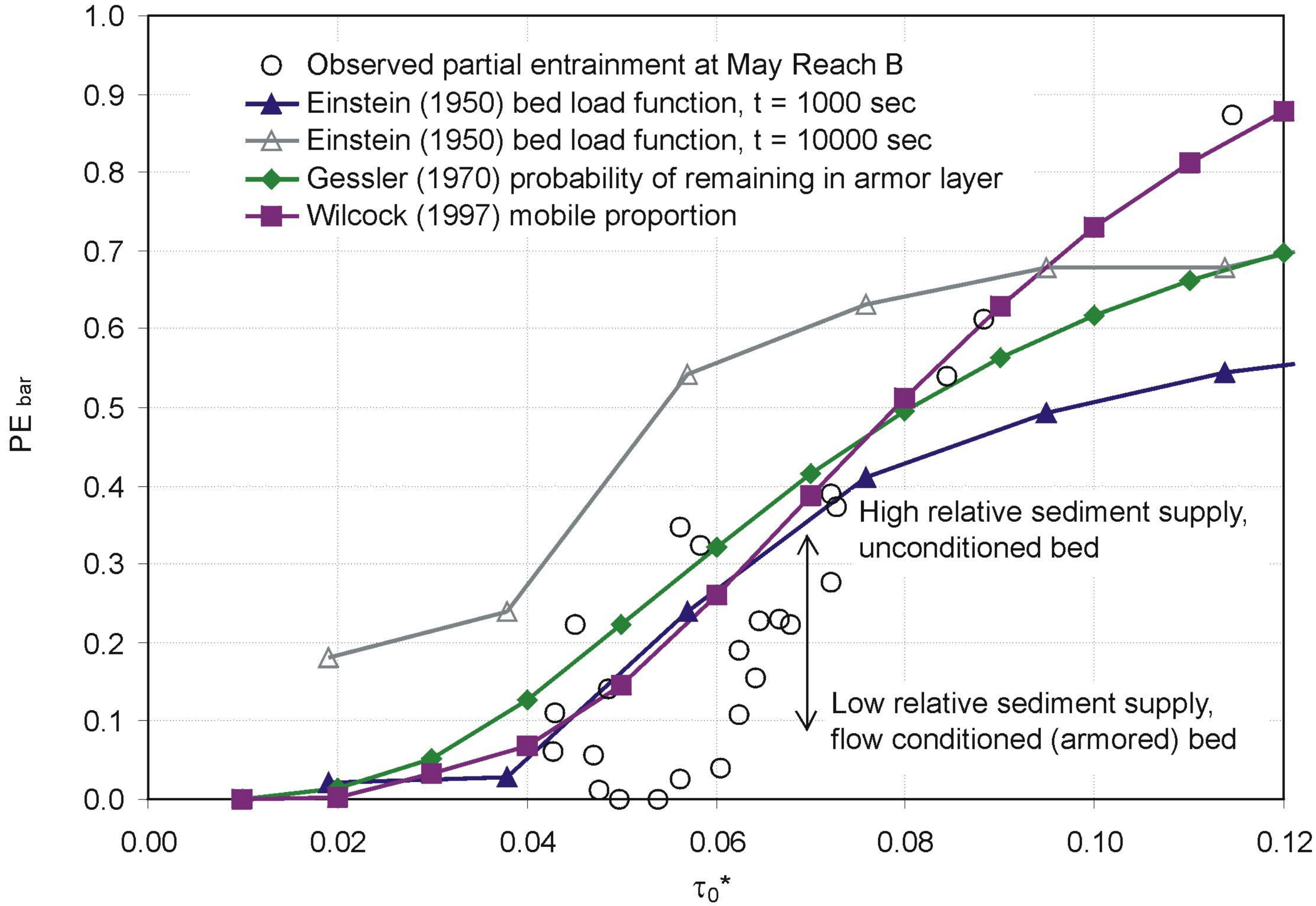



Figure 4.10: Predicted and observed partial entrainment ($PE_{bar}$) for May B as a function of dimensionless shear stress ($\tau_0^*$).

Table 4.1: Physical characteristics of streams at bed tag experiment sites.

| | Channel width | Bar length | Bar amplitude | Water surface slope | | Intermediate axis length for percentiles of the particle size distributions | | | |
|---|---|---|---|---|---|---|---|---|---|
| | m | m | m | Low water | High water | sample size | 10% mm | 50% mm | 90% mm |
| Jenkins A | 10.5 | 23 | 0.5 | 0.004 | 0.001 | 300 | 6 | 31 | 65 |
| Jenkins B | 7 | 24 | 0.3 | 0.016 | 0.006 | 200 | 11 | 48 | 111 |
| May Z | 9 | 34 | 0.2 | 0.016 | 0.011 | 100 | 10 | 45 | 135 |
| May A | 10 | 45 | 0.6 | 0.006 | 0.011 | 500 | 7 | 40 | 95 |
| May B | 9 | 35 | 0.5 | 0.008 | 0.016 | 500 | 8 | 31 | 88 |
| Swamp A | 5 | 16 | 0.1 | 0.003 | 0.012 | 200 | 4 | 40 | 130 |
| Swamp B | 9 | 45 | 0.2 | 0.002 | 0.007 | 400 | 3 | 20 | 54 |

| | Channel type | Bed material conditions |
|---|---|---|
| Jenkins A | pool-riffle | Well-sorted armor layer over bimodal gravel-sand mixture |
| Jenkins B | pool-riffle | Armored gravel bed |
| May Z | pool-riffle | Armored gravel bed |
| May A | plane-bed | Poorly-sorted gravel armor over bimodal gravel-sand mixture |
| May B | pool-riffle | Poorly-sorted gravel armor over bimodal gravel-sand mixture |
| Swamp A | plane-bed | Poorly-sorted gravel |
| Swamp B | pool-riffle | Gravel armor over bimodal gravel-sand mixture |

Table 4.2: Partial entrainment ($PE_{bar}$) observed in bed tag inventories.

| | Jenkins Creek | | May Creek | | | Swamp Creek | |
|---|---|---|---|---|---|---|---|
| | Bar A | Bar B | Bar Z | Bar A | Bar B | Bar A | Bar B |
| Number of rows | 5 | 8 | 5 | 5 | 5 | 5 | 5 |
| Number of tags | 95 | 103 | 60 | 84 | 85 | 45 | 77 |
| Date of inventory | $PE_{bar}$ | | $PE_{bar}$ | | | $PE_{bar}$ | |
| 10-Nov-97 | | | | 0.14 | | | |
| 18-Nov-97 | | | | 0.06 | | | |
| 20-Nov-97 | | | | 0.02 | | | |
| 24-Nov-97 | | | | 0.01 | | | |
| 2-Dec-97 | | | | 0.02 | | | |
| 4-Dec-97 | | | | 0.00 | | | |
| 10-Dec-97 | | | | 0.01 | | | |
| 12-Dec-97 | 0.08 | | | | | | |
| 17-Dec-97 | 0.05 | | | | | | |
| 19-Dec-97 | | | | 0.21 | 0.22 | | |
| 1-Jan-98 | 0.03 | | | 0.08 | 0.06 | | |
| 8-Jan-98 | | | | 0.40 | 0.39 | | |
| 13-Jan-98 | 0.07 | | | | | | |
| 18-Jan-98 | | | | 0.21 | 0.28 | | |
| 3-Feb-98 | 0.05 | | | 0.07 | 0.15 | | |
| 5-Mar-98 | 0.01 | | | 0.05 | 0.22 | | |
| 25-Mar-98 | 0.01 | | | 0.05 | 0.00 | | |
| 4-May-98 | | | | 0.00 | 0.11 | | |

(Shaded areas indicate no bed tags were installed at the bar, blanks indicate no inventory at the bar)

Table 4.2 (continued)

| | Jenkins Creek | | May Creek | | | Swamp Creek | |
|---|---|---|---|---|---|---|---|
| | Bar A | Bar B | Bar Z | Bar A | Bar B | Bar A | Bar B |
| Date of inventory | $PE_{bar}$ | | $PE_{bar}$ | | | $PE_{bar}$ | |
| 16-Nov-98 | | | | | | 0.04 | 0.07 |
| 17-Nov-98 | 0.00 | 0.00 | | 0.00 | 0.00 | | |
| 21-Nov-98 | 0.01 | 0.01 | | 0.04 | 0.35 | | |
| 28-Nov-98 | | | | 0.98 | 0.87 | | |
| 29-Nov-98 | 0.24 | 0.12 | | | | | |
| 3-Dec-98 | | | | 0.08 | 0.32 | | |
| 7-Dec-98 | 0.06 | 0.01 | | | | | |
| 11-Dec-98 | | | | 0.02 | 0.01 | | |
| 15-Dec-98 | | | | | | 0.44 | 0.96 |
| 23-Dec-98 | | | | 0.70 | 0.61 | | |
| 31-Dec-98 | | | 0.47 | 0.75 | 0.54 | | |
| 4-Jan-99 | 0.04 | 0.07 | | | | 0.18 | 0.23 |
| 25-Jan-99 | 0.00 | 0.04 | 0.06 | 0.30 | 0.19 | | |
| 27-Jan-99 | | | | | | 0.16 | 0.40 |
| 11-Feb-99 | | | 0.04 | 0.14 | 0.11 | | |
| 12-Feb-99 | 0.00 | 0.03 | | | | 0.11 | 0.33 |
| 8-Mar-99 | | | | | | 0.18 | 0.61 |
| 9-Mar-99 | 0.01 | 0.03 | 0.09 | 0.28 | 0.23 | | |
| 31-Mar-99 | 0.00 | 0.00 | 0.00 | 0.01 | 0.03 | 0.00 | 0.00 |
| 10-Nov-99 | | | 0.12 | | 0.06 | | |
| 12-Nov-99 | | | | | | | 0.00 |
| 16-Nov-99 | | | 0.08 | | 0.37 | | 0.56 |
| 29-Nov-99 | | | 0.07 | | 0.14 | | 0.09 |
| 13-Dec-99 | | | 0.00 | | 0.04 | | 0.17 |
| 17-Dec-99 | | | 0.00 | | 0.33 | | 0.00 |

Total number of inventories

| | | | | | | | |
|---|---|---|---|---|---|---|---|
| | 16 | 9 | 10 | 26 | 24 | 7 | 12 |

Table 4.3: Peak dimensionless shear stress ($\tau_0$*) bracketing initial movement of bed material.

| Site | Maximum value of $\tau_0$* when no tags were missing | Lowest $\tau_0$* when at least one tag was missing (% of tags missing) |
|---|---|---|
| Jenkins Creek | | |
| A | 0.039 | 0.025 (1%) |
| B | 0.047 | 0.039 (1%) |
| May Creek | | |
| Z | 0.045 | 0.040 (4%) |
| A | 0.049 | 0.027 (1%) |
| B | 0.055 | 0.044 (1%) |
| Swamp Creek | | |
| A | 0.048 | 0.047 (4%) |
| B | 0.049 | 0.044 (7%) |

Table 4.4: Rows where the fraction of tags missing ($PE_{row}$) was significantly different than bar-average values ($PE_{bar}$) for more than one inventory.

| $PE_{row} < PE_{bar}$ | Number of inventories when bed material moved | Number of inventories when $P[PE_{row}$ = Pebar] < 0.05 |
|---|---|---|
| May A.5 | 23 | 2 |
| Swamp B.1 | 9 | 4 |
| $PE_{row} > PE_{bar}$ | | |
| May A.7 | 23 | 4 |
| May A.10 | 22 | 2 |
| May B.3 | 15 | 3 |
| Swamp B.3 | 9 | 2 |

16 of the 38 bed tag rows had at least one observation when the probability that $PE_{row}=PE_{bar}$ was less than 0.05

Table 4.5: Stochastic changes in partial entrainment ($PE_{bar}$) at four gravel bars for pairs of inventories when $\tau_0^*$ was approximately equal.

| | Date of first inventory | $\tau_0^*$ | $PE_{bar}$ | Date of second inventory | $\tau_0^*$ | $PE_{bar}$ | Value of p in binomial distribution for equal likelihood of observed values of $PE_{bar}$ | Likelihood of observed values of $PE_{bar}$ given p from previous column |
|---|---|---|---|---|---|---|---|---|
| (a) increase in bed stability over time (i.e., significant decrease in fraction of tags missing for a pair of floods) | | | | | | | | |
| Jenkins A | 29-Nov-98 | 0.056 | 0.24 | 4-Jan-99 | 0.052 | 0.04 | 0.126 | 0.0014 |
| May A | 8-Jan-98 | 0.073 | 0.4 | 18-Jan-98 | 0.073 | 0.21 | 0.302 | 0.029 |
| Swamp B | 4-Jan-99 | 0.045 | 0.23 | 12-Nov-99 | 0.047 | 0.00 | 0.098 | 3.7E-04 |
| Swamp B | 8-Mar-99 | 0.082 | 0.61 | 13-Dec-99 | 0.082 | 0.17 | 0.382 | 4.1E-05 |
| May A | 25-Jan-99 | 0.070 | 0.30 | 11-Feb-99 | 0.068 | 0.14 | 0.218 | 0.032 |
| (b) decrease in bed stability over time (i.e., significant increase in fraction of bed tags entrained for a pair of floods ) | | | | | | | | |
| May B | 1-Jan-98 | 0.046 | 0.06 | 5-Mar-98 | 0.044 | 0.22 | 0.136 | 0.020 |
| (c) no change in $PE_{bar}$ between observations | | | | | | | | |
| May B | 21-Nov-98 | 0.058 | 0.35 | 3-Dec-98 | 0.060 | 0.32 | 0.336 | 0.59 |

# Chapter 5: Patterns and control of stream bed disturbance during floods in gravel-bed streams

Floods are a primary form of disturbance in stream ecosystems, entraining and transporting bed material along with periphyton, benthic organism, and organic debris downstream. Local biological conditions of streams, such as composition and abundance of periphyton and benthic invertebrates assemblages, are influenced by the movement of bed material during floods (Stehr and Branson, 1938; Power and Stewart, 1987). Broader biological conditions in a stream, such as the population levels of aquatic organisms and the structure and composition of lotic communities, may be influenced by the stream's flood disturbance regime, or patterns of stream bed disturbance in space and over time (Resh et al., 1988; Townsend, 1989; Poff and Ward, 1989; Poff and Allen, 1995).

The spatial extent of stream bed disturbance during a flood is determined by the shear and pressure forces applied by the flood waters and the strength of the material forming the stream bed. These factors are not independent: the shear and pressure forces applied by stream flow reflect the form and resistance of the channel; while the strength of the stream bed depends on the size and position of particles deposited by stream flow. The dual influences of stream flow, on the magnitude of floods and the form and materials of a stream channel, serves as the basis for a hypothesis of hydrologic control of the stream bed disturbance regime in gravel-bed streams. Under this hypothesis, the extent of stream bed disturbance during a flood is expected to vary with the ratio of the flood's discharge rate to some reference discharge rate. The reference discharge represents the suite of flows that determine the strength of the stream bed by selectively transporting small and unconstrained particles from the bed and re-positioning the remaining particles into a more stable configuration.

Hydrologic changes in a stream basin may be sufficient to produce changes in the stream bed disturbance regime if the frequency and magnitude of floods in the stream increase but the form and materials of the stream's channel do not completely adjust to the change in flood patterns. In this case, stream bed disturbance should be more frequent and extensive in the stream than it was prior to the hydrologic change. The motivation for examining the hypothesis is to assess whether the changes in stream flow patterns resulting from urban development manifest in a greater extent and frequency of stream bed disturbance.

In this investigation, the flood disturbance regime in Puget Lowland streams is characterized in terms of the estimated fraction of a gravel bar's surface that is entrained ($PE_{2\ yr}$) during the median annual (2-yr) flood. $PE_{2\ yr}$ is a simple but physically meaningful description of a stream bed disturbance regime as it represents the minimum spatial extent of bed surface disturbance in half of the years for a period of record. $PE_{2\ yr}$ does not describe all of the ecologically relevant aspects of floods, such as seasonal timing or sediment transport rates, but it provides an index of disturbance patterns in space and time.

## 5.1. Biological effects of floods

Floods disturb stream ecosystems when they entrain surface material from stream beds, causing changes in the biologic conditions of the stream. The biologic effects of flood disturbances include decreased periphyton biomass (Douglas, 1958; McCormick and Stevenson, 1991; Fisher et al., 1982; Power and Stewart, 1987), decreased densities and populations of benthic invertebrates and vertebrates (Stehr and Branson 1938, Anderson and Lehmkuhl, 1968; Orser and Shure, 1972), decreased taxonomic diversity (Gorman and Karr, 1978), increased taxonomic evenness (i.e., the relative abundance of different species) (McAuliffe, 1984), decreased predator abundance (Closs and Lake, 1994), and increased dominance of diatoms rather than detritus as trophic base (Fisher et al., 1982). The biological conditions in streams typically recovery quickly after a flood (Fisher et al., 1982; DeBray and Lockwood, 1990; Boulton et al., 1992; Jones et al.

1995), though the prevailing structure and composition of a lotic community may be influenced by the longer term flood disturbance regime of a stream (Shelford and Eddy, 1929; Odum, 1956; Horwitz, 1978; Fisher et al., 1982; Power et al., 1988; Poff and Ward, 1989; Death and Winterbourne, 1995; Poff and Allan, 1995; Poff et al., 1997).

The biological effects of a flood or other disturbances depend on the internal characteristics of the disturbance, the frequency of disturbance, and other factors external to the disturbance. The internal characteristics of a disturbance include its duration, time of year it occurs, the size and distribution of disturbed patches, and the intensity of disturbance within patches. Factors external to a disturbance, including habitat diversity (Gorman and Karr, 1978; Gurtz and Wallace; 1984) and biotic interactions (McAuliffe, 1984; Feminella and Resh, 1990; McCormick and Stevenson, 1991; Wootton et al., 1996) mediate the effects of disturbances in an ecosystem. Nonetheless, changes in the disturbance regime of a stream are expected to have biological effects.

Different types of streams have distinct flood disturbance regimes. In sand-bed channels, high flows ientrain bed material frequently (i.e., many times during a year) such that the bed surface in these channels is continually disturbed and, thus, organisms are adapted to or otherwise tolerate disturbance (Minshall, 1984; Resh et al., 1988). Conversely, a bedrock channel may not be disturbed at all by floods (Gurtz and Wallace, 1984). Given their extreme disturbance regimes, sand-bed and bedrock stream ecosystems may be relatively insensitive to all but the most profound anthropogenic changes in flood frequencies and magnitudes.

The flood disturbance regimes of gravel bed stream ecosystems span a much wider range of frequency and area domains. For example, the spatial extent of bed disturbance during a flood on 26 - 27 November 1998 ranged from 12% to 96% at the seven gravel bars in three Puget Lowland streams, described in Chapter 4. Likewise, Resh et al. (1988) and ASCE Task Committee (1992) identify gravel bed stream ecosystems as particularly sensitive to anthropogenic hydrologic modification because of the potential for changes in their flood disturbance regimes.

## 5.2. Hydrologic control of flood disturbance regime in gravel-bed streams

Stream flow influences both components of the force balance determining the stability/mobility of stream bed material. The force applied during flood typically varies with a flood's discharge rate. Likewise, the strength of a gravel stream bed reflects the forces applied during intermediate flows that condition the stream bed surface through size-selective sediment transport and particle clustering and packing. The influence of stream flow on both components of the force balance that determines whether or not stream bed material is entrained during a flood forms the basis for a hypothesis of hydrologic control of flood disturbance regimes in gravel-bed streams.

Hydrologic control of the flood disturbance regime in gravel-bed streams has two conditions. First, the shear stress applied to the stream bed ($\tau_0$) during a flood and, consequently, the spatial extent of stream bed disturbance varies with the discharge rate (Q) in the stream. The applied shear stress of stream flow can be estimated in uniform flow by the total boundary shear stress:

$$\tau_0 = \gamma_{water} R \ S \qquad 5.1$$

In uniform flow, $\tau_0$ increases with the hydraulic radius (R), which is the cross-sectional area of stream flow divided by the wetted perimeter of the channel, or the energy gradient (S). Generally, $\tau_0$ will be directly related to Q because the near-bed velocity and velocity gradient in most streams increase with Q. This condition contrasts with "hydraulic control" where local factors (e.g., channel form, vegetation and other obstructions to flow) influence the relationship between Q and $\tau_0$. For example, $\tau_0$ may decrease with Q where a backwater condition is created at high flows.

The second condition for the hypothesis of hydrologic control is that the shear strength of the stream bed surface is determined by the shear stress applied during a suite of intermediate flows capable of transporting bed material at low rates. As a result, there should be a direct, monotonic relationship between the shear strength of the stream bed and a reference discharge rate that indexes these flows. The relationship between stream flow and the shear strength of the bed is complex: a transient discharge may have a persistent effect on shear strength of the bed; the bed may not reach an equilibrium with any discharge that can transport bed material; and other factors (e.g., sediment supply) mediate the effects of stream flow on the bed's shear strength.

Stream flow changes the shear strength of a stream bed through transport and deposition of bed material. The strength of a bed surface increases when stream flow entrains fine-grained or unconstrained particles from the bed, moves them to more stable locations (e.g., into clusters with other grains or to a downstream reach with lower shear stress), and deposits coarser particles from upstream reaches. Alternatively, the strength of the bed surface decreases when stream flow deposits finer grained particles over coarser material or breaches a coarse surface layer (i.e., armor) by transporting material indiscriminately with respect to size. In all these cases, stream flow influences the strength of the bed but the strength of the bed is not a direct, monotonic function of discharge.

## 5.3. Factors influencing the strength of a stream bed

The shear strength of the stream bed surface is the measure of a surface material's ability to resist entrainment by a downstream shear force. The shear strength of a surface is typically expressed as the maximum force per area or critical shear stress ($\tau_{cr}$) at which the material is stable and above which it begins to move (Selby 1982). White (1940) showed that $\tau_{cr}$ for a spherical particle in a uniform sediment varies with its diameter. $\tau_{cr}$ of a stream bed comprising different sized particles (i.e., a mixed sediment) is spatially variable (Lane and Kalinsky, 1940; Einstein, 1950; Miller and Byrne, 1966, Grass, 1970; Fenton and Abbot, 1977; Li and Komar, 1986; Kirchner et al., 1990; Wilcock and McArdell, 1997). Moreover, inter-grain structures formed by packing, imbrication, and clustering of particles can increase $\tau_{cr}$ of a stream bed (Laronne and Carson 1976; Brayshaw et al., 1983; Church et al., 1998) without a change in its particle-size distribution. In any event, the shear strength of a unimodal, mixed sediment generally varies with particle-size distribution of the sediment and, in particular, with the diameter of the median of the particle-size distribution ($D_{50}$) (Wilcock, 1993).

Many factors influence the particle-size distribution of a stream bed including the rate and duration of stream flow, the particle-size distributions of upstream sediment sources and the initial bed surface, the rate of sediment supply, and channel morphology. Each of these factors has a dominant influence on the particle-size distribution of the bed surface under specific conditions, which are generally related to sediment transport rates. Table 5.1 provides examples of the influence of each factor drawn from flume experiments, field studies of streams, and theoretical models. The examples are divided into three columns based on whether their focus was the particle-size distribution of the bed at equilibrium ($dD_{50}/dt = 0$) or the response of the particle-size distribution of the bed surface to changes in a factor over time ($dD_{50}/dt < 0$ or $dD_{50}/dt > 0$).

Each factor modifies the particle-size distribution of the bed surface by controlling either the applied shear stress ($\tau_0$) or the particle-size distribution of the sediment supply. Under high sediment transport rates, the particle-size distribution of a stream bed does not vary as a function of the applied shear stress, particularly at an equilibrium state ($dD_{50}/dt = 0$). Instead, the particle-size distribution of the stream bed surface approaches the particle-size distribution of the sediment supply, which may be an upstream source (e.g., an aggrading stream or a sediment-feed flume) or the subsurface bed material (e.g., breaching of an armor layer, a degrading stream, a sediment-recirculating flume).

$\tau_0$ has a direct influence on the particle-size distribution of a stream bed during periods of lower rate sediment transport where upstream sources of sediment are limited (e.g., clear water flumes) or share the same particle-size distribution as the initial bed surface (e.g., sediment-recirculating flumes and long reaches of gravel bed streams with uniform morphology and materials). In these cases, the particle-size distribution of the stream bed will vary with $\tau_0$ provided the sediment is well graded (i.e., a wide range of particle sizes are available for transport) and sediment transport is size-selective (i.e., larger particles are immobile while smaller ones are mobile). If sediment transport ceases, or increases such that all particles are mobile, then the particle-size distribution of a bed surface will no longer vary as a direct function of $\tau_0$.

The particle-size distribution of a stream bed surface is expected to vary with $\tau_0$ only in streams where the rate of sediment supply does not overwhelm the stream's transport capacity but is high enough to prevent a static armor from forming and where the sediment supply to the stream bed is well graded (i.e., has a wide range of particle sizes). The first condition is likely to be satisfied in pool-riffle and plane-bed reaches of gravel bed streams, which Montgomery and Buffington (1997) characterized as moderately transport-limited. Both conditions are likely to be satisfied in low to moderate gradient reaches of gravel-bed streams in the Puget Lowland where well-graded glacial drift provides a wide range of particle sizes for transport by streams. These conditions limit the scope of this analysis as there are streams where the particle-size distribution and, consequently, the strength of the bed surface does not vary with $\tau_0$. In such streams, the stream bed disturbance regimes are not under hydrologic control.

## 5.4. Reference discharge

The hypothesis of hydrologic control of stream bed disturbance regimes requires that the strength of the bed surface and, in particular, its particle-size distribution vary with the applied shear stress at a reference discharge. The particle-size distribution of a bed surface varies with the applied shear stress only during periods when a marginally higher discharge rate would increase the transport rate of finer grains relative to the transport rate of coarser grains and a marginally lower discharge rate would increase the deposition of finer grains relative to coarser grains. Sediment transport is size-selective during these periods because the transport rate of an individual size-class of bed material relative to the total bed load transport rate is not equal to the fraction of the bed surface occupied by that size class.

Thus, the particle-size distribution of the stream bed surface will vary with the applied shear stress for only a limited range of flows when sediment transport is size-selective. While any single-valued reference discharge can only serve as an index for a suite of geomorphically effective flows (Andrews and Nankervis, 1995), a primary objective of this investigation is to evaluate whether the particle-size distribution of a stream bed surface does vary with the applied shear stress associated with a reference discharge for gravel bed streams.

The reference discharge has a physical basis rather than just a statistical relationship to the particle-size distribution of the bed surface. The reference discharge represents a suite of flows that determine the particle-size distribution of a stream bed. If the reference discharge is not competent to transport bed material, then the particle-size distribution is static at the reference discharge and would not vary with marginal changes in the magnitude of the reference discharge. Likewise, if the reference discharge is competent to move all particles on the stream bed, then the particle-size distribution would not vary with changes in the magnitude of the reference discharge. The reference discharge should be associated with conditions of low-rate, size-selective sediment transport.

### 5.4.1. Shear stress-based criteria for the reference discharge

A reference discharge is posited here to represent the range of flows when marginal changes in $\tau_0$ produce marginal changes in the particle-size distribution of the bed surface. In streams where the applied shear stress at the reference discharge is high, the particles forming the bed surface are expected to be coarse. In streams where the applied shear stress at the reference discharge is low, the particles forming the bed surface are expected to be fine.

The applied shear stress of stream flow is represented here by the total boundary shear stress ($\tau_0$) using Equation 5.1, which assumes uniform flow conditions. $\tau_0$ accounts for all of the forces resisting the stream flow divided over the area of stream bed. The shear stress acting on bed material, which is the grain shear stress ($\tau_g$) or skin friction, can be estimated by partitioning $\tau_0$ into a form drag component produced in regions of flow separation (e.g., at bar forms) and the grain component acting on bed material (Einstein and Barbarossa, 1952). Alternatively, the local shear stress acting on the stream bed can be estimated by solving the Prandtl-von Karman logarithmic velocity distribution for turbulent flow (Schlichting, 1979) using near-bed velocity measurements (Nece and Smith; 1970).

In either case, a grain or local shear stress is only a point estimate of the spatial distribution of the applied shear stress over a stream bed. Since $\tau_0$ is the total of all forces resisting flow in a reach averaged over the surface area of the reach, it is a reasonable single-valued estimate of the distributed values of local shear stress where form drag is relatively small and current velocities are approximately uniform.

The relationship between the particle-size distribution of a bed surface and an applied shear stress can be expressed as a dimensionless shear stress, $\tau_0$*:

$$\tau_0{}^* = \frac{\tau_0}{(\gamma_{sediment} - \gamma_{water})D_{50}} \qquad 5.2$$

where $D_{50}$ is the median of the particle-size distribution, $\gamma_{sediment}$ is the specific weight of the bed material and $\gamma_{water}$ is the specific weight of water.

Previous investigations of gravel bed streams have estimated the value of $\tau_0$* for low bed load transport rates and size-selective transport. Andrews (1983) suggests that bed material entrainment begins approximately when $\tau_0$* is 0.02, though individual particles may be entrained from a bed at lower values of $\tau_0$* (Fenton and Abbott, 1977; Komar, 1987). Buffington and Montgomery (1997) proposed an approximate lower limit of 0.03 for incipient motion in gravel bed streams depending on the specific application. In the experiments described in Chapter 4, none of the bed tags at any site moved when $\tau_0$* was less than 0.026, indicating that much or all of the bed surface at the sites was stable.

In this investigation, $\tau_0$* is posited to be greater than 0.02 at the reference discharge. At lower discharges, the particle-size distribution of the bed will be static and independent of marginal changes in the discharge rate. While other structural changes to the bed surface (e.g., packing, clustering, imbrication) may occur at lower values of $\tau_0$*, they are assumed to have little influence on the particle-size distribution of the gravel bar. This is not the case, however, in reaches with high sediments loads where fine material is deposited over the stream bed during low flows.

The upper limit on $\tau_0$* for the reference discharge must be less than the level at which bed material is transported indiscriminately with respect to particle size. Under high flows that transport bed material of any size, the particle-size distribution of the bed surface will vary as a function of the sediment supply (upstream sources or subsurface bed material) rather than discharge rate. The transition from size-selective transport to indiscriminate transport is indicated by the fining of the particle-size distribution of the bed surface, such as when an armor layer is destroyed. Little and Mayer (1976) showed that armored bed surfaces were destroyed in a clear water flume at $\tau_0$* between 0.06 and 0.10, where $\tau_0$* was calculated using the $D_{50}$ of the initial, unarmored bed surface, depending on the initial sorting of the bed material. Likewise, the results of Wilcock and McArdell (1993) indicated that the bed surface in a sediment re-circulating flume coarsened with increasing shear stress until $\tau_0$* was approximately 0.06 (based on the $D_{50}$ of the initial bed surface) at which point the bed surface began to fine. The values of $\tau_0$* for size-selective sediment transport calculated by Little and Mayer (1976) and Wilcock and McArdell (1993) are likely to be lower than the estimates of $\tau_0$* presented in this investigation, which are based on total boundary shear stress and the particle-size distribution of armored gravel bed surfaces. Buffington and Montgomery (1997) compare different methods for calculating $\tau_0$*.

Results of the bed tag experiments, described in Chapter 4, provide evidence for an upper limit on $\tau_0$* for a reference discharge. Intermediate floods, when $\tau_0$* was between 0.070 and 0.085, increased the strength of the stream bed. The bed surface was relatively weak after those (larger) floods when $\tau_0$* was greater than 0.085. Thus, the strength of the bed may be directly related to $\tau_0$* only when $\tau_0$* is less than 0.085 which serves the maximum value of $\tau_0$* for the evaluation of a reference discharge. During flows when $\tau_0$* is greater than 0.085, much of the bed surface is likely to be entrained and its particle-size distribution will approach that of the sediment supply (e.g., upstream sources or subsurface bed material).

Other investigations of gravel bed streams have also found an upper limit on $\tau_0$* for size-selective bed load transport. Andrews (1994) described a condition of "marginal" bed load transport in Sagenhen Creek, CA over the range $0.02 < \tau_0^* < 0.06$ (based on the $D_{50}$ of surface material) where "a majority of the particles on the bed surface are in motion" (p. 2241) at the upper limit. Parker et al. (1982) analyzed bed load transport data for Oak Creek, OR, which is an armored, gravel bed stream, and found that the bed load particle-size distribution was similar to, though somewhat coarser than, the subsurface particle-size distribution when $\tau_0$* was greater than 0.042 (based on surface

$D_{50}$). They concluded that the particle-size distribution of the subsurface material, rather than the applied shear stress, determines the particle-size distribution of the bed load in this case. While Parker et al. indicated the bed surface was partially entrained under this condition, they did not rule out the possibility of further coarsening of the bed surface when $\tau_0$* was greater than 0.042. Ashworth and Ferguson (1989) found that the particle-size distributions of bed load and the bed surface converged, indicating the upper limit on size-selective transport, over a wide range of $\tau_0$* (from 0.03 to 0.2 based on surface $D_{50}$) for three gravel bed streams.

Marginal changes in $\tau_0$* modify the strength of a stream bed surface only during a limited range of flows associated with size-selective transport. In this investigation, the lower limit on $\tau_0$* during these flows is assumed to be approximately 0.02 while the upper limit is 0.08. When $\tau_0* < 0.02$, bed load transport rates are likely to be very low and changes in $\tau_0$* will not influence either the particle-size distribution or the strength of the stream bed. When $\tau_0* > 0.08$, bed load transport rates are likely to be high and the particle-size distribution and the strength of the bed surface may not increase with $\tau_0$*. A stream bed may continue to strengthen (i.e., armor) when $\tau_0$* exceeds 0.085 where the sediment supply is limited (e.g., downstream of a large reservoir), but such streams are not examined here. Since marginal changes in the applied shear stress at these discharges may cause marginal changes in the particle-size distribution of the bed surface, the applied shear stress at a reference discharge is evaluated relative to the range of $\tau_0$* from 0.02 to 0.08.

### 5.4.2. The hydrologic basis for a reference discharge

The strength of a gravel stream bed does not adjust instantaneously to an applied shear stress. Stream flow must work to change the particle-size distribution of a stream bed by depositing or entraining bed material selectively with respect to size. In cases where the particle-size distribution of the bed surface changes (e.g., high sediment supply and transport rates), the reference discharge may be the magnitude of the most recent discharge that exceeded the threshold of bed load transport. While high flows can quickly change the particle-size distribution of the bed surface, lower flows may also have an effect on the strength of the bed surface if they occur for a sufficiently long duration (Church et al., 1998). In moderately transport-limited gravel-bed streams, the duration and range of flows influencing the particle-size distribution the bed surface are uncertain. Using the hydraulic criteria defined by $0.02 < \tau_0* < 0.08$, I examined four stream flow statistics, spanning a range of temporal domains: the annual mean discharge rate($Q_{mean}$), the discharge rate exceeded 10% of the time ($Q_{0.1}$); the discharge rate exceeded 5% of the time ($Q_{0.05}$); and the median annual peak discharge ($Q_{2\ yr}$).

The annual mean discharge rate ($Q_{mean}$) is evaluated here as a reference discharge because it provides a measure of common magnitude of stream flow that is influenced, in particular, by the magnitude and duration of high flows. Leopold and Maddock (1953) introduced the annual mean discharge rate ($Q_{mean}$) as a hydrologic index for hydraulic geometry relationships, which related width, depth, and velocity to discharge. They argued that $Q_{mean}$ was a reliable index of geomorphically effective stream flows because "the mean annual rates of discharge at all points on a large number of rivers are equaled or exceeded about the same percentage of time," and the frequency of discharge at any one point is about the same as the frequency of discharge at any other point for the river they investigated (Leopold and Maddock, 1953, p. 3). Moreover, any discharge could have been used as an independent variable for hydraulic relationships provided that the dependent variables (width, depth, and velocity) were measured at that discharge.

While $Q_{mean}$ may be competent to transport bed material in the large, low gradient rivers examined by Leopold and Maddock (1953), $Q_{mean}$ is likely to be less than the threshold for bed load transport in gravel bed streams. As a result, there is little physical basis for using $Q_{mean}$ as a reference discharge for the particle-size distribution of the stream bed. Furthermore, the duration that the stream flow rate exceeds $Q_{mean}$ varies inversely with the extent of urban development in Puget Lowland stream basins (see Chapter 2). Consequently, $Q_{mean}$ is not a consistent indicator of the magnitude of sediment transporting flows for gravel bed streams with different extents of urban development.

Other investigations have attempted to provide a "mechanistic" basis for the relationship between channel form and stream flow by incorporating a discharge rate that accounts for the greatest cumulative sediment transport over time. Inglis (1949) introduced a relationship between "dominant discharge" and meander wave length. Leopold and Wolman (1957) used bankfull discharge, along with channel slope, to distinguish braided from meandering channels based on the hypothesis that bankfull discharge indexes an intermediate range of discharges that transport the most sediment over time. Henderson (1963) concluded that stream channels are at the threshold of motion during bankfull discharge and speculated that channel form is determined by relationships of hydraulic geometry at the threshold of motion.

A dominant, effective, or bankfull discharge has been widely adopted as the independent variable in equilibrium (i.e., regime and hydraulic geometry) equations for gravel bed streams (Kellerhals, 1976; Li et al., 1976; Parker, 1979; Hey and Thorne, 1986). The hydrologic basis for selecting a particular equilibrium discharge has been analyzed in terms of both flow duration (Wolman and Miller, 1960; Andrews, 1984) and flood frequency (Harvey, 1969; Pickup and Warner, 1976; Williams, 1978). If the particle-size distribution of the stream bed changes quickly in response to a change in discharge, then a frequent flood may be a reliable reference discharge. The median annual flood ($Q_{2\ yr}$) is evaluated here a reference discharge representing the case where the textural response of a gravel stream bed is rapid during floods and insensitive to lower flows.

Investigations of sediment transport in flumes and gravel bed streams provide evidence that a given shear stress must be applied for some period of time before the stream bed reaches an equilibrium state (Little and Mayer, 1970; Garde et al., 1977;

Gomez, 1983; Shen and Lu, 1983; Wolcott, 1990; Chin et al., 1992; Wolcott, 1990; Reid and Larrone, 1995; Wilcock and McArdell, 1997; Church et al., 1998). Accordingly, a reference discharge may need to persist for some minimum duration of time before the particle-size distribution reflects its magnitude. Discharges exceeded 10% of the time ($Q_{0.1}$) and 5% of the time ($Q_{0.05}$) represent relatively high discharge rates, which are likely capable of transporting some bed material, and persist for longer periods (on average 36 and 18 days per year respectively) than most floods (see Figure 2.15). The strength of gravel stream bed surfaces may, as a result, have time to "adjust" to the applied shear stress at these discharges. These two stream flow statistics are evaluated as reference discharges to indicate the influence of flow duration on the particle-size distribution of the bed surface.

## 5.4. Methods for estimating the extent of stream bed disturbance at a reference discharge

The extent of bed disturbance was estimated at 19 gravel bars in 13 Puget Lowland streams. Figure 5.1 shows the location of the streams. A stream gage is located no more than 1 km away from each site. Table 5.2 identifies the streams, the basin area at each gage, and road density in the basin (the ratio of total road length to basin area). The stream basins span the range of urban development in the Puget Lowland region, as indicated by road density from less than 3 km/km$^2$ (Big Beef, Huge, and Rock Creeks) to over 7 km/km$^2$ (Des Moines, Leach, Miller, and Swamp Creeks) with many streams having intermediate levels of urban development (Big Bear, Covington, Jenkins, May, Newaukum, and Soos Creeks)

The stream basins in the analysis display the range of physiographic features found in the Puget Lowland including glacial till-mantled plateaus (e.g., Big Bear and Big Beef Creeks), glacial outwash plains and valleys (e.g., Rock, Jenkins, and Miller Creeks), lakes and wetlands (e.g., Jenkins, Covington, and Big Bear Creeks), ravines (e.g., Miller, May, and Des Moines Creeks, and Cedar River Tributary 0308), broad floodplains (e.g., Swamp Creek) and shallow groundwater (e.g., Jenkins Creek). Additionally, Big Beef, Newaukum, and May Creeks have high-elevation headwaters in bedrock mountains. The diversity of physiographic features represented in these stream basins produce a wide range of hydrologic patterns particularly at lower levels of urban development. The results of the analysis, thus, should be applicable to gravel bed streams throughout the Puget Lowland region and other temperate, maritime regions.

The extent of bed material entrainment during a flood is equal to the total area where the local shear stress applied by stream flow exceeds the local shear strength of the bed surface (Lane and Kalinsky 1940, Grass 1970). Field sites were selected so that $\tau_0$* can be used as a common indicator of the local shear stress distribution among streams. The sites are located on gravel bars in straight channels with uniform widths. Since the sites are limited to gravel bars, hydraulic conditions (e.g., water surface slope, depth) and cross-sections are relative uniform. Flow is well distributed across the channel and unencumbered by large obstructions or vegetation in the channel with no large zones of flow separation or other severe cross-channel velocity gradients.

For each stream, a straight reach with a transverse or mid-channel bar (Church and Jones 1982) was identified. The bars form a riffle in the stream at most sites, which are pool-riffle channels, except in Swamp Creek where the bed is relatively planer and the bar' amplitude is low. Multiple reaches in some streams were analyzed to provide replicate sites within a stream (Miller, May, and Jenkins Creeks) or because there are two gages in the stream (May and Swamp Creeks).

Each reach was surveyed to construct a longitudinal profile of the reach and cross-section of the channel across the foreset (downstream) slope of the bar. The particle-size distribution of the surface material on the bar was determined using a Wolman (1954) pebble count, in which 100 particles were selected at random from the stream bed within 5 m of the surveyed cross-section, and their intermediate axis length was measured to the nearest mm. Sand grains (< 2 mm) were noted and included in the count but represented less than 10% of the particles in all samples. Table 5.2 lists the water surface slope and $D_{50}$ of the particle-size distribution of surface material for each bar.

The analysis of stream disturbance regime relies on a series of hydraulic calculations for four stream flow statistics ($Q_{mean}$, $Q_{0.1}$, $Q_{0.05}$, and $Q_{2\ yr}$). The statistics were estimated from discharge records for water years (WY) 1989 to WY 1998, though data were not available for every stream gage in all of these years. The specific period of record for each stream is listed in Table 5.2. $Q_{2\ yr}$ represents either an "instantaneous" peak or the maximum discharge rate for a 15-minute interval; the other statistics are based on daily mean discharge data. As show in Figure 2.8, the bias in flow duration quantiles introduced by using daily discharge, rather than 15 minute discharge, is negligible for discharges exceeded more than 1% of the time.

The hydraulic radius for each discharge rate was calculated using the laws of mass conservation and Manning's equation for the mean velocity of uniform flow:

$$Q = uA = \frac{S^{0.5} R^{0.67}}{n} RP \tag{5.3}$$

where u is mean velocity through a channel cross-section, A is the wetted cross-sectional area, S is the downstream energy gradient of the stream flow, P is the wetted perimeter, n is a roughness coefficient, and R is hydraulic radius.

Manning's roughness coefficient (n) must be specified in Equation 5.3 before calculating the hydraulic radius. The roughness coefficient represents the effects of forces resisting the flow of water on its mean velocity at a section. Flow resistance in streams depends on the size, pattern, and concentration of surface roughness elements, vegetation and organic debris, channel form, obstructions in the channel, flow depth, and the stability of the free surface (Keulegan, 1938; Chow, 1959; Rouse, 1965; Ikeda and Isumi, 1990). The roughness coefficients for cross-sections in May, Swamp, and Jenkins creeks were calibrated using Manning's equation and mean current velocity from measurements made during periods of storm flow ($Q \approx 1$ to 2 $m^3/s$).

For all other streams, the roughness coefficient had to be estimated. Several approaches have been developed to account for the many sources of flow resistance (e.g., Einstein and Barbarossa, 1952; Leopold and Wolman, 1957; Chow 1959; Barnes, 1967; Hey, 1979). Two empirical equation, developed by Jarrett (1984) and Bathurst (1985) for gravel-bed streams with slopes of 0.002 to 0.04, were used here to estimate the roughness coefficients in those streams where stage and discharge were not measured (Miller, Leach, Huge, Soos, Newaukum, Covington, Des Moines, Big Bear, Rock, and Big Beef Creeks).

Jarrett (1984) developed an equation for Manning's roughness coefficient based on velocity measurements in 21 high gradient (water surface slopes greater than 0.002) gravel bed streams in Colorado. He found that n could be described as an exponential function of water surface slope and hydraulic radius:

$$n = 0.32 S^{0.38} R^{-0.16} \tag{5.4}$$

The root mean square percentage error of estimates was 28% of the values of n calculated from velocity measurements and Manning's equation (Jarrett, 1984).

Bathurst (1985) developed an empirical flow resistance equation for 15 gravel-bed streams in Britain with high relative roughness (i.e., low flow depths when compared to the protrusion of coarse particles from the stream bed). This equation expresses flow resistance in terms of the Darcy-Weisbach friction factor (ff), an alternative roughness coefficient:

$$\sqrt{\frac{8}{ff}} = 5.62 \log\left(\frac{d}{D_{84}}\right) + 4 \tag{5.5}$$

where, d is the mean water depth (i.e., cross-sectional area divided by wetted channel width) and $D_{84}$ is the length that is greater than the intermediate axis of 84% of the particles on the stream bed. The root mean square percentage error of estimates of ff was 34% of the calculated values of ff (Bathurst, 1985). The Darcy-Weisbach friction factor is related to Manning's roughness coefficient by:

$$n = \sqrt{\frac{ff}{8gR^{\frac{1}{3}}}} \qquad (5.6)$$

The average value of n derived from Jarrett (1984) and Bathurst (1985) was used as a first estimate of Manning's roughness coefficient ($n_1$) in the hydraulic calculations for all of the streams except Jenkins, May, and Swamp Creeks, where n was estimated from current velocity and stage measurements. A second estimate of the Manning roughness coefficient was made to assess the sensitivity of shear stress calculations to n. Equations 5.4 and 5.5 are based, in part, on data collected during large, infrequent floods, when total flow resistance may be largely a result of grain roughness. In smaller floods, such as those considered in this analysis, form drag may contribute considerably to total flow resistance (Parker and Peterson, 1980; Prestegaard, 1983). Thus, Equations 5.3 and 5.4 may under-estimate total flow resistance for the sites in this analysis. For the second estimate of Manning's roughness coefficient, $n_2$, the first estimate of $n_1$ is increased by 50%, which is greater than the root mean square percentage error of either Equation 5.3 or 5.4.

For each reference discharge at a site, the hydraulic radius (R) was calculated at a surveyed cross-section by solving Equation 5.3 iteratively for a series of flow depths until the calculated discharge rate (Q) was equal to the reference discharge. Separate calculations were made, first, using $n_1$ and, then, using $n_2$. Table 5.3 provides the value of the parameters for the hydraulic calculations at each gravel bar. For each reference discharge at a site, $\tau_0$ is assumed to have a uniform probability of occurring within the range defined by Equation 5.1 based on the values of R corresponding to $n_1$ and $n_2$.

The dimensionless shear stress ($\tau_0$*) was calculated for each bar using Equation 5.2 where $\gamma_{sediment} = 26500$ N/m$^3$ and $D_{50}$ is the median of the particle-size distribution of the surface material on each gravel bar. The extent of bed disturbance during the median annual flood is estimated using the results of bed tag experiments described in Chapter 4. The experiments demonstrated a direct relationship between the fraction of bed tags entrained during a flood to the peak $\tau_0$* for that flood. Figure 5.2 is a plot of the peak $\tau_0$* at seven gravel bars in Jenkins, May, and Swamp Creeks during floods in WY 1998 and 1999 and the fraction of bed tags moved from each bar.

The extent of bed disturbance is expressed as partial entrainment, PE, which represents the fraction of a gravel bar's surface disturbed during a flood. A linear equation relates PE to the peak dimensionless shear stress during a flood:

$$PE = \begin{Bmatrix} 0 & \tau_0^* \le 0.045 \\ 12.5\left(\tau_0^* - 0.045\right) & 0.045 < \tau_0^* < 0.125 \\ 1 & \tau_0^* \ge 0.125 \end{Bmatrix} \quad (5.7)$$

Estimates based on Equation 5.7 have a root mean square error of 0.099 and are ±0.31 of the observed values of PE (Figure 5.2).

The necessary condition for a reference discharge is that the particle-size distribution of the bed surface and, in particular, $D_{50}$ varies directly with its applied shear stress, $\tau_0$. If the relationship between $D_{50}$ and $\tau_0$ is linear among the sites, then the dimensionless shear stress the reference discharge ($\tau_{ref}$*) will be constant at the sites:

$$\tau_{ref}^* = C \quad (5.8)$$

This condition is evaluated for each of the hydrologic statistics proposed as potential reference discharges.

If $\tau^*_{ref}$ is constant across the sites in this analysis, then the dimensionless shear stress during a flood and, consequently, the extent of bed disturbance will be proportional to the ratio of the shear stress in the flood to the shear stress of the reference discharge:

$$\tau^*_{flood} \propto \frac{\tau_{flood}}{\tau_{ref}} \tag{5.9}$$

Equations 5.7 and 5.9 formulate the hypothesis of hydrologic control over stream bed disturbance patterns: partial entrainment of a stream bed during a flood must be proportional to the ratio of the applied shear stress of the flood to that of the reference discharge. Since $\tau_{flood}$ is a factor on both sides of Equation 5.9, $\tau^*_{flood}$ may be correlated spuriously with $\tau_{flood}/\tau_{ref}$ (Benson, 1965). Spurious correlation is avoided, however, in the alternative formulation provided by Equation 5.8.

The hypothesis of hydrologic control also requires that the applied shear stress of stream flow, during both floods and the reference discharge, is a function of the discharge rate. A few approximations can be used to determine the functional form of the relationship between discharge and the applied shear stress in a stream. Under uniform flow conditions, R varies approximately as an exponential function of Q assuming a constant value for n and a relatively small change in the wetted perimeter with a change in Q:

$$Q \propto R^{1.67} \tag{5.10}$$

A relationship between the shear stress and the discharge rate for uniform flow in a wide channel results from combining Equations 5.1 and 5.10:

$$\tau_0 \propto Q^{0.6} \tag{5.11}$$

The relationship between partial entrainment of a stream bed and dimensionless shear stress can be recast in terms of a discharge ratio:

$$PE \propto \left(\frac{Q_{flood}}{Q_{ref}}\right)^{0.6} \tag{5.12}$$

This investigation will not test whether 0.6 is the optimal value of the exponent in Equation 5.12. The equation indicates, however, that the functional form of the relationship between PE and the discharge ratio $Q_{flood}/Q_{ref}$ should be exponential.

A first-order uncertainty analysis was used to assess the error in the estimated values of PE. The standard deviation ($\sigma$) of a variable (y) that is a linear function of multiple, uncorrelated variables ($x_i$) is estimated as:

$$\sigma_y = \sqrt{\sum_i \sigma_{x_i}^2 \left(\frac{\partial y}{\partial x_i}\right)^2} \tag{5.13}$$

where the partial differential terms are evaluated at the estimated value of y (Benjamin and Cornell, 1970, p. 184).

Equation 5.13 was applied twice in the uncertainty analysis. First, it was used to calculate the standard deviation of estimates of $\tau_0$* as function of the standard deviation of $\tau_0$ and $D_{50}$. The standard deviation of $\tau_0$ was calculated assuming $\tau_0$ had a uniform probability of falling between the low estimate of $\tau_0$ based on $n_1$ and the high estimate of $\tau_0$ based on $n_{2:}$ $\tau_{high}$ - $\tau_{low}$)/$\sqrt{12}$. The standard deviation of $D_{50}$ was estimated as half of the difference between the 45$^{th}$ and 55$^{th}$ percentiles of the particle-size distribution, which provide nonparametric estimates of one standard deviation below and above, respectively, of the $D_{50}$ for a sample size of 100 (Helsel and Hirsch, 1993, p. 70).

After the standard deviation of $\tau_0$* was calculated, Equation 5.13 was used to calculate the standard deviation of estimated values of PE where the estimated value of PE is a function of $\tau_0$* and the distribution of PE at a given value $\tau_0$*. The standard deviation of PE at a given $\tau_0$* was calculated using assuming PE has a uniform probability of occurring within 0.31 of the values of calculated from Equation 5.7 (i.e., the standard deviation of PE was $0.62/\sqrt{12}$). The interval of ±0.31 around Equation 5.7 contains all observed values of $PE_{bar}$.

## 5.5. Results

### 5.5.1. The particle-size distribution of a gravel bar and the reference discharge

The particle-size distribution of surface material on a gravel bar is generally related to the applied shear stress at any of the possible reference discharges. The value of $D_{50}$ increases at the sites with the $\tau_0$ for any reference discharge (Figures 5.3 to 5.6). The vertical error bars in Figures 5.3 - 5.6 represent the 95% confidence intervals for the median diameters of the bar surface material.

The horizontal error bars in Figures 5.3 - 5.6 represent 95% confidence intervals around the estimate of total boundary shear stress at each site. The points corresponding to gravel bars in Jenkins, May, and Swamp Creeks do not include horizontal error bars because the hydraulic radius is known at each bar to within 1 cm from observations of flow depth for discharge rates up to the median annual flood.

$\tau_0$* is relatively constant among the sites for any reference discharge, indicating an approximately linear relationship between $D_{50}$ and $\tau_0$. For any hydrologic statistic, the values of $\tau_0$* are generally contained within a 2-fold interval as represented by the dashed lines of constant $\tau_0$* in Figures 5.3 – 5.6. The mean values of $\tau_0$* at the reference discharges range from 0.025 for $Q_{mean}$, with a root mean square error (RMSE) of 0.0014 (5.8% of the mean) to 0.077 for $Q_{2yr}$, with a RMSE of 0.005 (6.4% of the mean). The values of $\tau_0$* are least variable for $Q_{0.1}$, which had a mean $\tau_0$* of 0.034 with a RMSE of 0.0017 (5.0% of the mean), and for $Q_{0.05}$, which had a mean $\tau_0$* of 0.043 with a RMSE of 0.0019 (4.7% of the mean).

In addition to having the lowest variability, the values of $\tau_0$* at $Q_{0.05}$ and $Q_{0.1}$ are well within the range of values associated with size-selective transport (i.e., $0.02 < \tau_0^* < 0.08$). In contrast, $\tau_0$* for $Q_{mean}$ was less than 0.02 at 2 sites and $\tau_0$* for $Q_{2\ yr}$ was greater than 0.08 at 10 sites. Since $\tau_0$* is outside the range of size-selective transport for $Q_{mean}$ and $Q_{2\ yr}$, these statistics lack a physical basis for controlling the particle-size distribution of bed material in some streams. Thus, either $Q_{0.05}$ or $Q_{0.1}$ are more appropriate choices for the reference discharge than $Q_{mean}$ or $Q_{2\ yr}$.

### 5.5.2. Disturbance during the median annual flood

The estimates of partial entrainment (PE) based on Equation 5.7 ranged from 0.17 for Jenkins and Soos creeks to 0.96 for Des Moines Creek for the median annual flood ($Q_{2yr}$) with an average value of 0.55 for all sites. Figure 5.7 displays the estimates of PE for each site plotted against its drainage area. The error bars in Figure 5.7 represent one standard deviation above and below the mean estimate of PE. There is no apparent relationship between drainage area and the spatial extent of disturbance for these sites.

None of the highly urban streams are likely to have low levels of disturbance during the median annual flood. PE is less than 0.30 only in streams where road densities are less than 6 km/km$^2$ (Rock, Covington, Big Bear, Soos, and Jenkins Creeks) (Figure 5.8). Leach Creek, with a road density of 9.9 km/km$^2$, had the lowest value of PE (0.32) for the median annual maximum flood which is likely influenced by a large, in-channel stormwater detention reservoir. Low levels of development does not assure low levels of disturbance: Big Beef, Newaukum, and Huge creeks have road densities less than 3 km/km$^2$, but PE is greater than 0.50 during the median annual flood at these sites.

The extent of disturbance during the median annual flood in streams with intermediate levels of urban development ranges widely. At the downstream sites on May Creek, where the road density is 5.0 km/km$^2$, the values of PE were 0.50 and 0.62. In contrast, the sites on Jenkins Creek, where the road density is 5.4 km/km$^2$, the values of PE were 0.01 and 0.09. Indeed, floods approximately equal to the median annual maximum flood were observed in the bed tag experiments, described in Chapter 4, to entrain most of the surface material at the May Creek sites but little of surface material at the Jenkins Creek sites.

Although the extent of stream bed disturbance during the median annual maximum flood may not vary consistently with urban development in a stream basin, stream flow patterns provide a better explanation for the predicted differences in the extent of disturbance between the sites. Under the hypothesis of hydrologic control, the extent of bed disturbance during the median annual flood (PE) should be a function of the ratio of the peak magnitude of the median annual flood to the magnitude of the reference discharge (Equation 5.13). Since $\tau_0$* is relatively constant among streams for any reference discharge and $\tau_0$ increases with Q among the sites, PE will necessarily vary with $Q_{2yr}/Q_{ref}$.

Among the reference discharges, $Q_{0.05}$ had values of $\tau_0$* with the lowest variability and a mean value closer to the center of the hydraulic criteria for size-selective bed load transport than the other hydrologic statistics. PE at the sites varies directly with $Q_{2\,yr}/Q_{0.05}$ (Figure 5.9) illustrating a general trend of increasing extent of bed disturbance as the $Q_{2yr}$ is larger relative to $Q_{0.05}$. The concordance of the sites with the general trend in Figure 5.9, versus the scatter of points in Figure 5.8, supports a hypothesis of hydrologic control over stream bed disturbance patterns regardless of the level of development in a stream basin, though the extent of disturbance during the median annual flood varies widely for streams where the $Q_{2\,yr}$ is 2 to 5 times the magnitude of $Q_{0.1}$.

The variation in the magnitude of annual maximum floods was proposed in Chapter 2 as a control on geomorphic stability of stream channels. In streams with low inter-annual variability as indicated by a low coefficient of variation for annual maximum floods ($CV_{AMF}$), the magnitude of infrequent floods are smaller relative to frequent floods than in streams with high inter-annual peak flood variability. As a result, the extent of bed disturbance caused by frequent floods would be expected be higher in streams with low values of $CV_{AMF}$.

PE at the sites is plotted against $CV_{AMF}$ for WY1989 to 1998 for the sites in Figure 5.10. PE is greater than 0.50 at the five sites where the $CV_{AMF}$ is less than 0.70. At this sites, the value of $Q_{2yr}/Q_{0.05}$ is also greater than 3, so this analysis cannot distinguish whether the high values of PE are a result of lower reference discharges relative to peak flood magnitude or lower inter-annual variation in peak flood magnitude. However, Huge and May Creeks have high inter-annual variation in peak flood magnitude (i.e., $CV_{AMF}$ is greater than 1.0) but are still likely to have extensive bed disturbance for the median annual flood (i.e., PE is greater than 0.50).

## 5.6. Discussion of stream bed disturbance patterns

The relationship between the particle-size distribution of the bed surface and a reference discharge represents a geomorphic equilibrium (Gilbert, 1877; Lacey, 1929; Shulits, 1936; Mackin, 1948; Leopold and Maddock, 1953; Lane, 1955a and b; Leopold and Wolman, 1957; Henderson, 1963; Schumm, 1969). The particle-size distribution of the bed surface is established during periods of low-rate, size-selective sediment transport when changes in the discharge rate can produce changes in the particle-size distribution of bed surface. While the particle-size distribution of the bed surface may not be at a steady-state under the reference discharge (e.g., Gomez, 1983; Wolcott, 1990; Wilcock and McArdell, 1997; Church et al., 1998), its observed consistency from summer-to-summer at gravel bars in Jenkins, May, and Swamp Creeks represents a dynamic equilibrium (Chorley, 1962; Langbein and Leopold, 1964).

The reference discharge used here has a longer duration than typical values of an effective discharge for channel formation or cumulative sediment transport (Wolman and Miller, 1960; Pickup and Warner, 1976; Williams, 1978; Andrews, 1984, Carling, 1988, Whiting et al. 1999). It is likely that the reference discharge here represents only the lower range of geomorphically effective stream flows that structure a stream bed's surface.

Hydrologic changes resulting from urban development elicit a variety of geomorphic responses from stream channels. Increased storm flow volume and rates increases hillslope and fluvial sediment transport resulting in bank erosion, channel incision, headward erosion of canyons, and aggradation of low gradient channels (Hammer, 1972; Ebisemiju, 1989; Whitlow and Gregory, 1989; Booth, 1990; Trimble, 1997). Trimble (1995) provides evidence for a geomorphic disequilibrium, which he defined as when the mean rates of sediment supply and transport are not equal, persisting in urban streams for decades after hydrologic change with different parts of a stream network adjusting at different rates.

However, few investigations have detailed the sedimentological changes in urban streams (Douglas, 1985; Brooks, 1996). Wolman and Schick (1967) found that the spatial extent of fine grained-deposits increased over two Maryland stream beds as a result of increased sediment loads generated by construction activities on hillslopes. The increased sediment load produced by bank erosion may also contribute to a general fining of the stream bed.

Once a stream basin has been developed and land use is relatively stable, the stream channel can be expected to attain a new equilibrium with the urban stream flow patterns (Henshaw and Booth, manuscript in review for publication in Journal of the American Water Resources Association). At a new, post-development equilibrium with increased storm flow rates, bed material distributions would be coarser particularly where sediment sources are limited through erosion control efforts (Lane, 1955a; Dietrich et al., 1989). Indeed, Thoms (1987) found that mean particle size of subsurface bed material from a reach of the River Tame flowing through Birmingham, England was coarser and sand fractions were lower than expected compared to samples from a rural reach of the adjacent River Blythe. The issue of stream bed disturbance, however, is not simply whether the particle size distribution is coarser in urban streams but whether the particle size distribution adjusts sufficiently to limit the frequency and extent of bed disturbance to levels comparable to streams with less urban development.

The particle size distributions of the surface bed material at the sites in this investigation reflect intermediate-magnitude flows. Intermediate flows modify the strength of a stream bed surface when the particle-size distribution of the material transported from and deposited in a reach diverges from the particle-size distribution of the bed surface in the reach. Furthermore, intermediate flows can move particles into more stable positions. These processes occur at relatively low bed load transport rates over time scales that may be longer than the duration of a single flood (Gomez, 1983a; Wolcott, 1990; Wilcock and McArdell, 1997; Church et al., 1998) particularly in the Puget Lowland where floods are produced by rain storms and would typically have a shorter duration than floods produced by snowmelt.

Characteristics hydrologic effects of urban development include reducing the duration of storm flow and increasing the magnitude of flood peaks relative to recessional and base flows (see Figures 2.2 and 2.15). Even if the particle size distribution of an urban stream bed increases in response to higher storm flow rates, it is unlikely to adjust (i.e., become supply limited) to the increased rates because of the short period of time available for sediment transport and the lower relative magnitudes of long duration flows (e.g., $Q_{0.1}$). Thus, bed disturbance in gravel-bed streams will be more frequent and extensive than it was prior to development as a result of hydrologic changes provided sediment continues to be available for transport in the stream.

The "flashy" hydrologic conditions that indicate extensive bed disturbance (i.e., short duration floods with high magnitudes relative to recessional and base flows) are typical of arid regions as well. Reid and Laronne (1995) observed analogous, though more pronounced, conditions in Nahal Yatir, Israel. Nahal Yatir is an ephemeral stream with rapid storm flow recession. As a result, sediment transport rates are very high and there is no armor development at the bed surface.

Likewise, gravel-bed streams in the Puget Lowland with low levels of development are likely to have frequent and extensive disturbance where stream flow patterns are flashy. For example, Big Beef, Huge, May and Newaukum Creeks have values of $Q_{2yr}/Q_{0.05}$ greater than 3 and PE greater than 0.5 for the median annual maximum flood. The stream flow patterns in these basins may be a result of natural physiographic conditions such as mountain headwaters, greater rain volumes and rates during storms, and higher rates of sediment supply to their channels. However, natural physiographic conditions may not be sufficient to produce frequent and extensive levels of stream bed disturbance, since all of their basins have been logged and have large areas without forest cover. In such cases, natural physiographic conditions may dictate the sensitivity of a stream to land use.

Urban streams with relatively attenuated stream flow patterns are likely to have lower levels of bed disturbance. For example, Leach Creek is expected to have only moderate levels of bed disturbance (PE = 0.3) during the median annual flood despite its high level of urban development. The level of disturbance in Leach Creek, however, is typical of other streams with similar hydrologic regimes. In this case, a large in-channel detention pond may be effective at moderating Leach Creek's stream flow patterns and, as a consequence, its stream bed disturbance regime.

Two urban stream flow patterns, high peak discharge rates relative to the discharge rate of longer duration structuring flows, and low variation in annual maximum floods, have a physical basis for maintaining *persistently extensive* bed disturbance in gravel-bed streams. For the streams analyzed here, the magnitude of a flood relative to structuring flows displays a closer relationship to the extent of bed disturbance than does the variation in annual maximum floods.

## 5.7. Conclusions

The flood disturbance regime of gravel bed streams in the Puget Lowland depends on the frequency and magnitude of floods as well as the magnitude of longer duration, intermediate magnitude flows that structure the stream bed. Intermediate flows determine, in part, the particle size distribution of the bed surface, which has a primary influence on the strength of the bed surface. The discharge rate exceeded 5% of the time ($Q_{0.05}$) provided a robust reference discharge for the suite of flows determining the particle size distribution of the bed surface in moderately transport-limited (i.e., pool-riffle and plane-bed) gravel-bed streams.

The fraction of a gravel bar's surface entrained during a flood is predicted to vary with the ratio of the flood's peak discharge rate to $Q_{0.05}$. More than 30% of a gravel's bar surface is likely to be disturbed during a flood with a peak discharge rate that is 3 times $Q_{0.05}$. More than 60% of a bar's surface is likely to be disturbed during a flood that 6 times greater than $Q_{0.05}$. The influence of stream discharge on both the strength of the stream bed and the magnitude of the applied force of a flood provide hydrologic control over stream bed disturbance patterns in gravel bed streams.

The ratio of a flood with a specific frequency (e.g, $Q_{2yr}$) to $Q_{0.05}$ is generally higher in streams with high peak discharge rates and rapid storm flow recession whether a result of urban development of physiographic conditions. Urban development in a stream basins promotes geomorphic instability in gravel bed streams when the magnitude of frequent floods increase, the rate of storm flow recession increases, and the magnitude of intermediate flows decreases. For the Puget Lowland urban streams analyzed here, the peak discharge rate of the median annual maximum flood was more than 3 times $Q_{0.05}$ indicating frequent and extensive bed disturbance is likely in these gravel-bed streams.

Likewise, gravel-bed streams with lower levels of development may have frequent and extensive bed disturbance if the magnitude of frequent floods is large relative to $Q_{0.05}$.

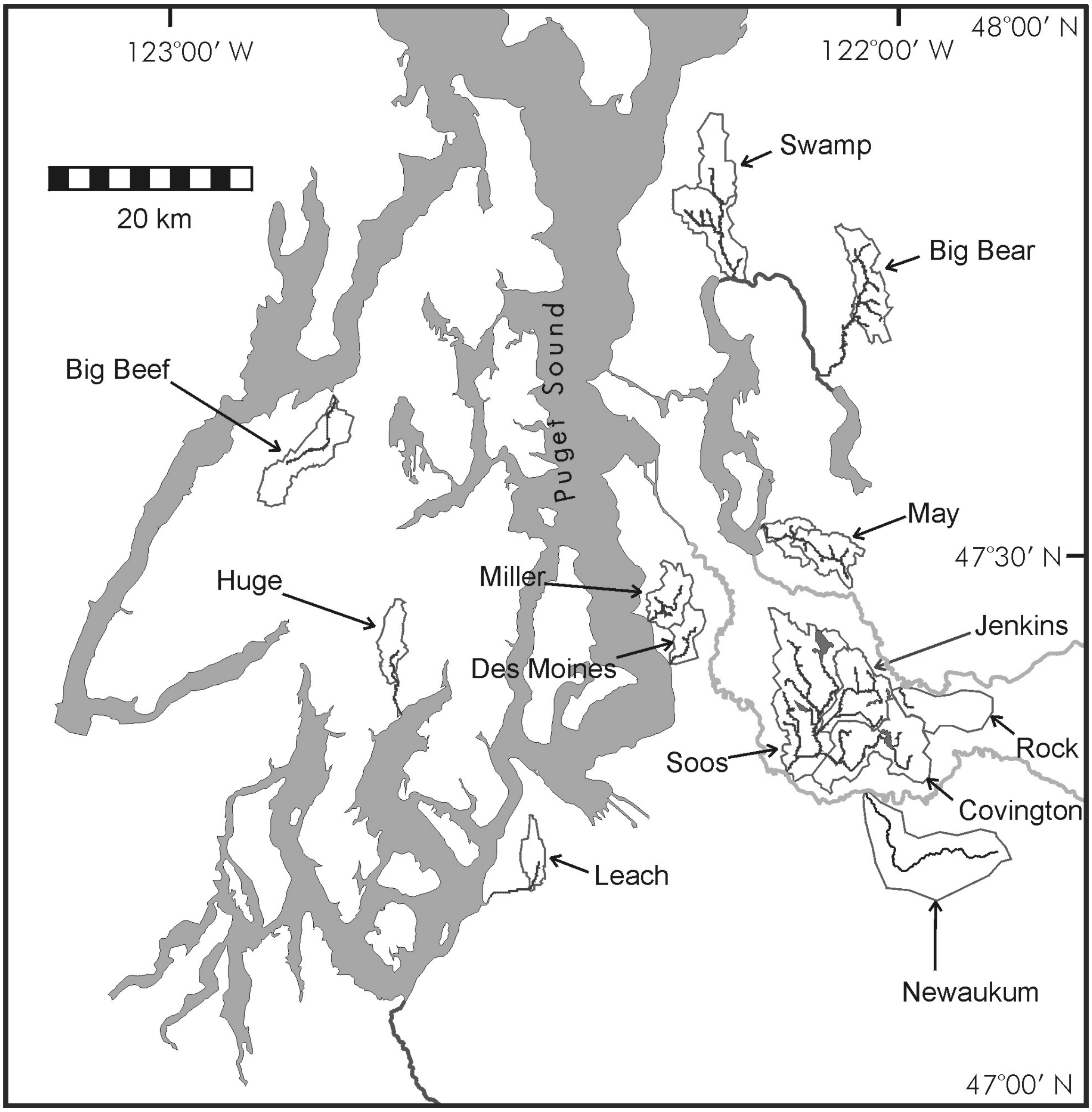



Figure 5.1: Streams used in the bed disturbance analysis.

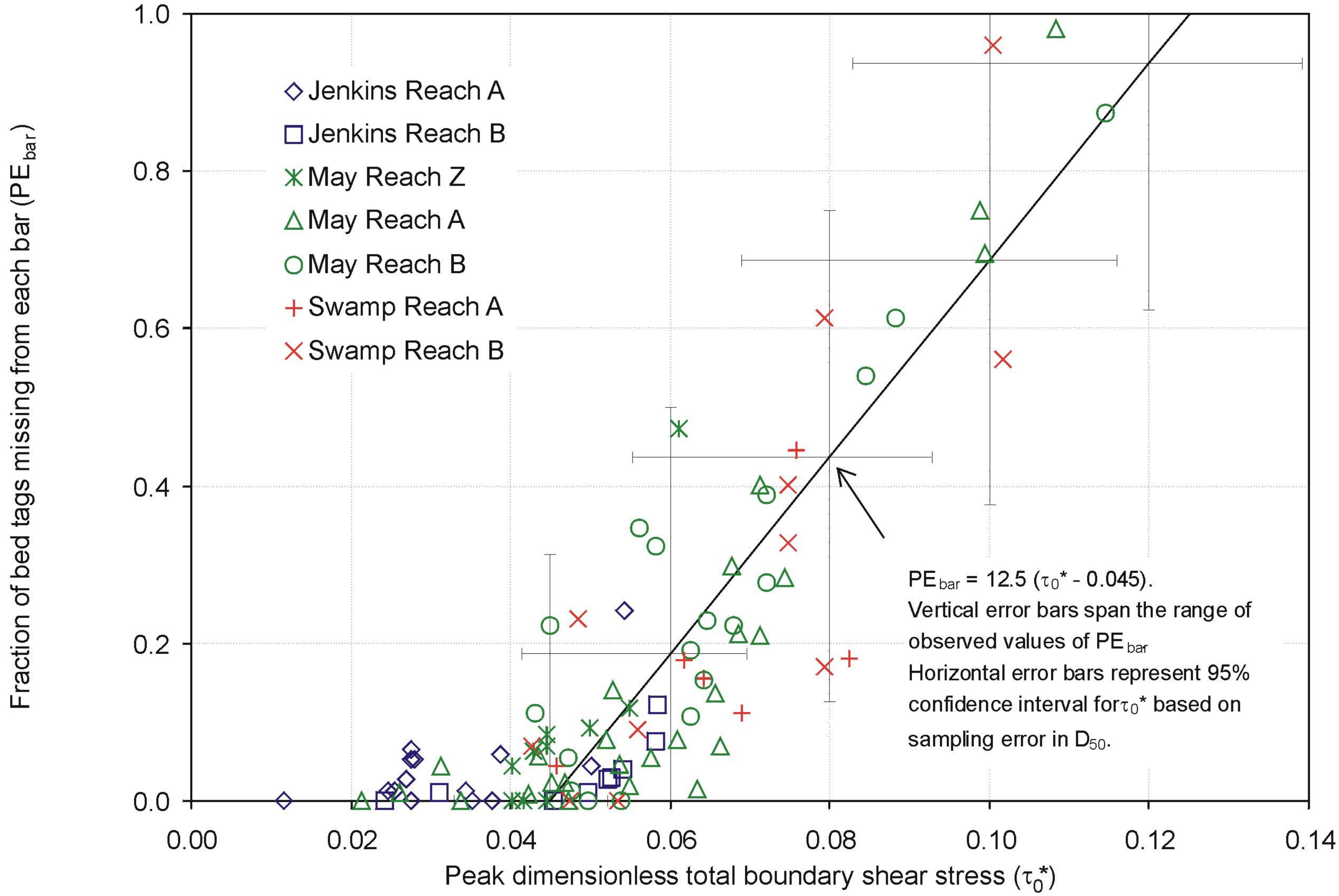



Figure 5.2: Bed tag inventory results used to predict the extent of bed disturbance.

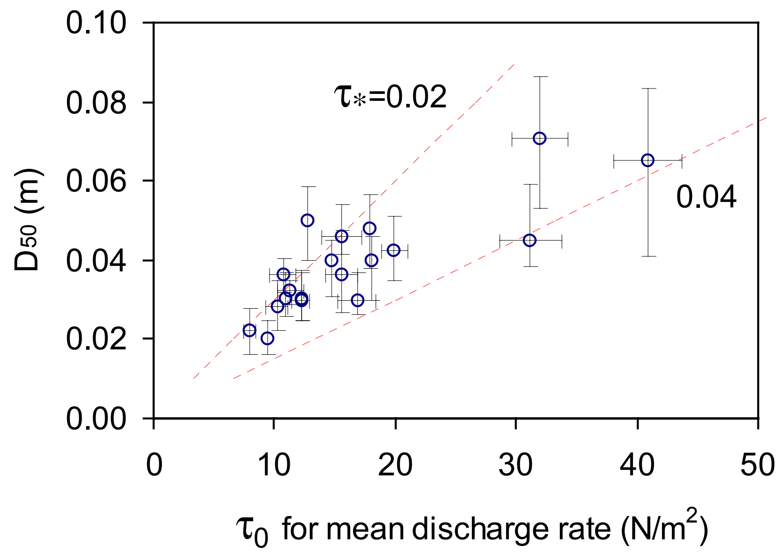



Figure 5.3: Median of the particle-size distribution for the gravel bar surfaces as a function of shear stress for mean discharge rate with 95% confidence intervals.

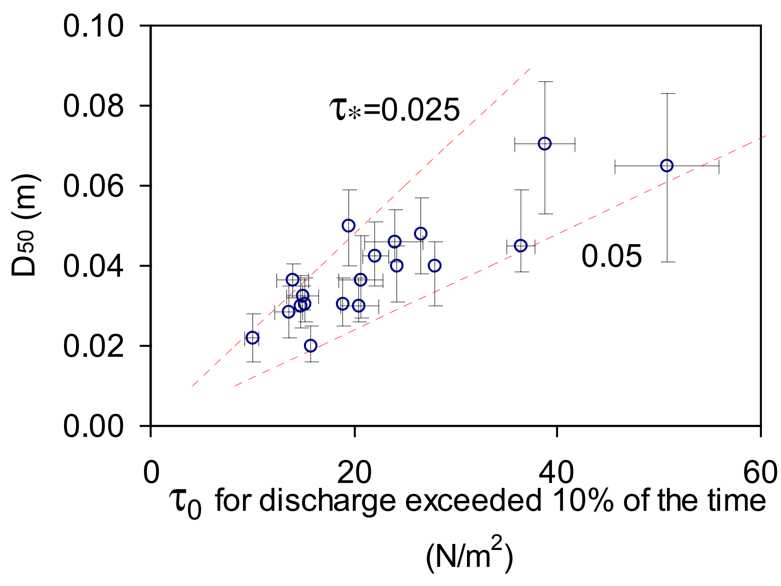



Figure 5.4: Median of the particle-size distribution for the gravel bar surfaces as a function of shear stress for discharge exceeded 10% of the time with 95% confidence intervals.

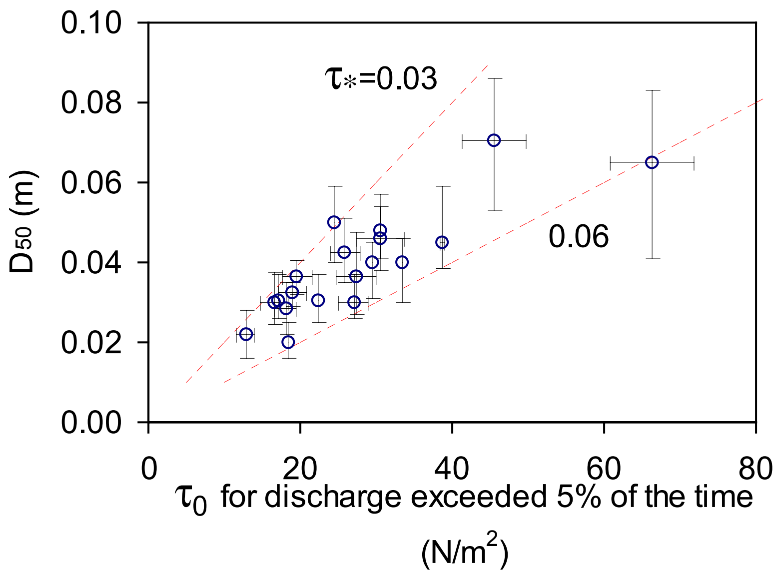



Figure 5.5: Median of the particle-size distribution for the gravel bar surfaces as a function of shear stress for discharge exceeded 5% of the time with 95% confidence intervals.

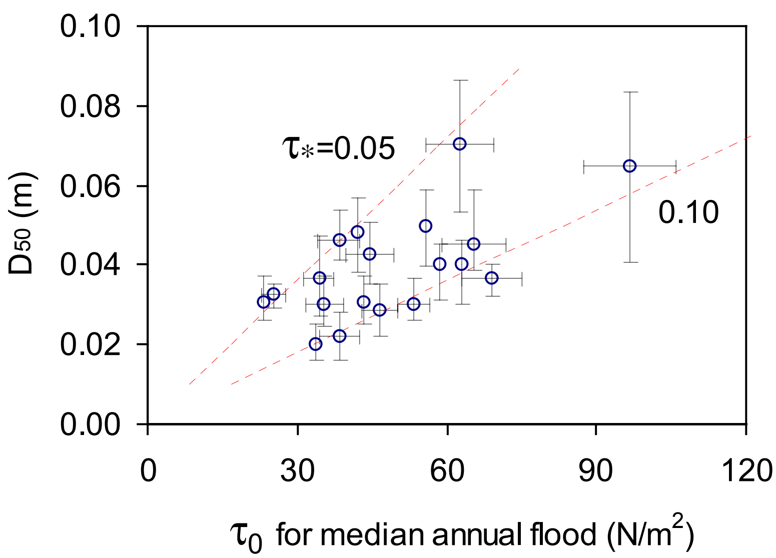



Figure 5.6: Median of the particle-size distribution for the gravel bar surfaces as a function of shear stress for median annual flood with 95% confidence intervals.

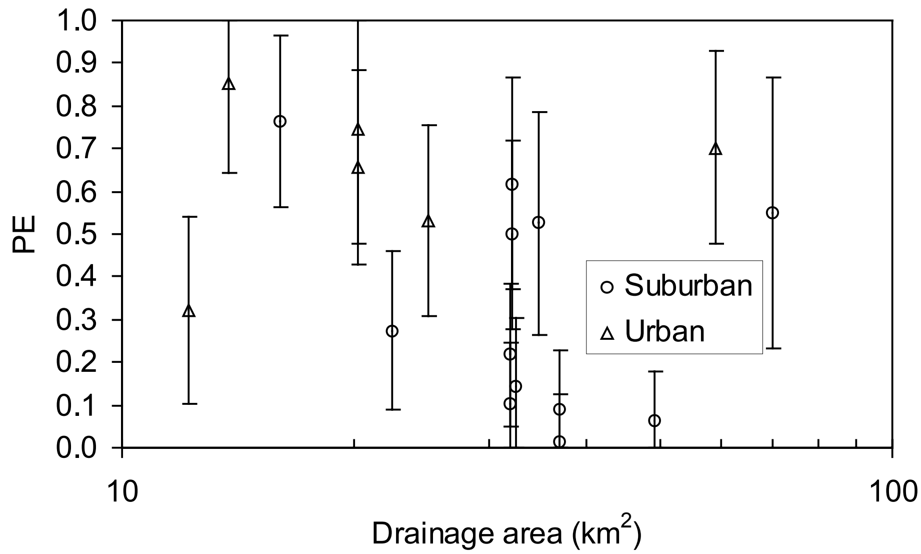



Figure 5.7: Partial entrainment (PE) during the median annual maximum flood, with error bars ± 1 standard deviation, plotted against drainage area.

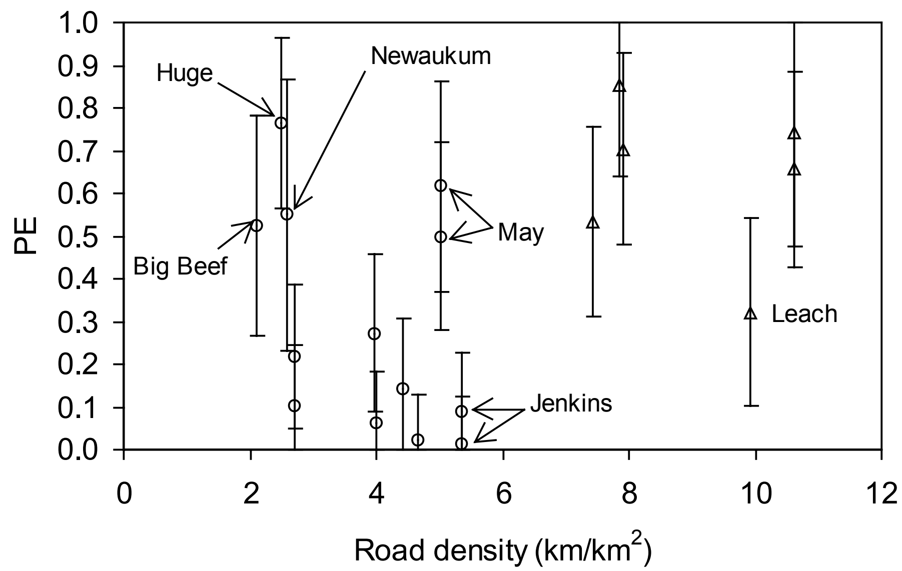



Figure 5.8: Partial entrainment (PE) during the median annual maximum flood, with error bars ± 1 standard deviation, plotted against road density

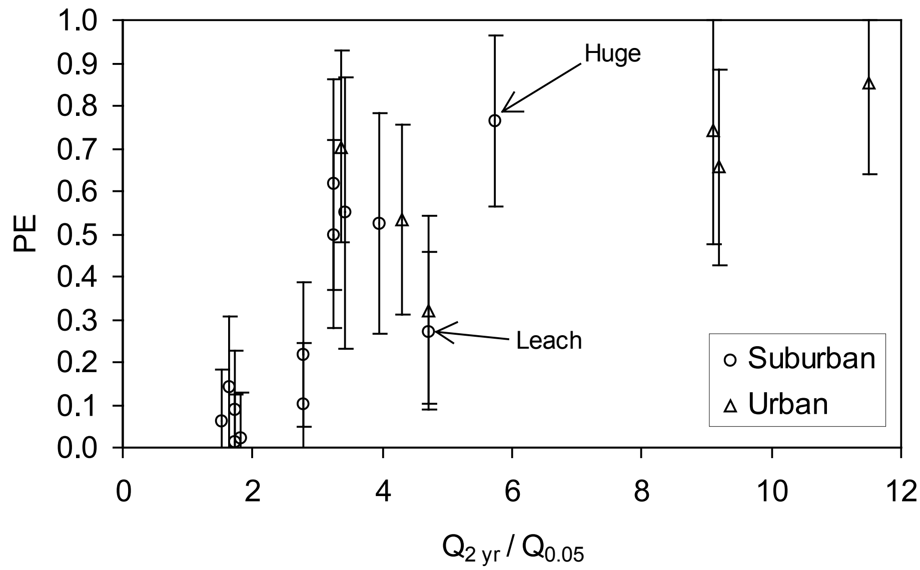



Figure 5.9: Partial entrainment (PE) during the median annual maximum flood, with error bars ± 1 standard deviation, plotted against the ratio of the flood's discharge ($Q_{2yr}$) to the discharge exceeded 5% of the time ($Q_{0.05}$).

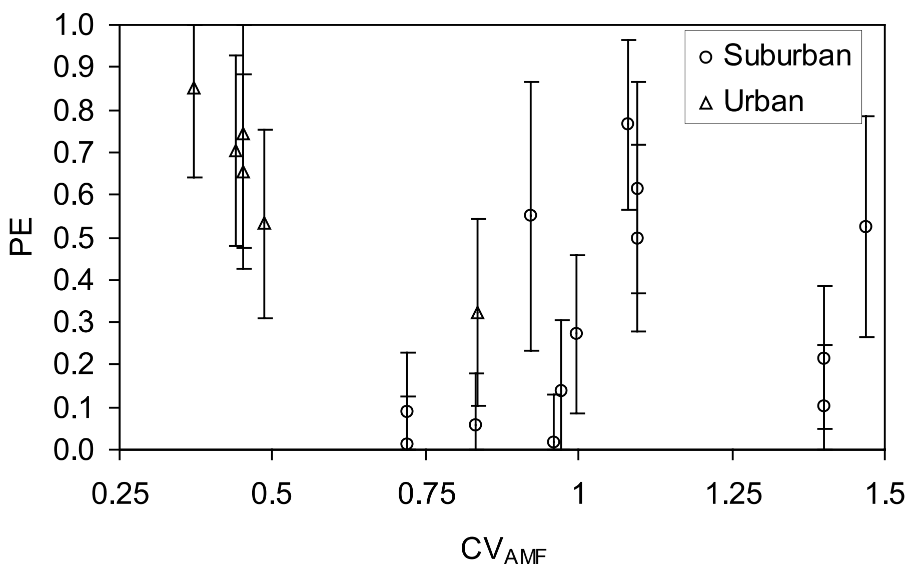



Figure 5.10: Partial entrainment (PE) during the median annual maximum flood, with error bars ± 1 standard deviation, plotted against the coefficient of variation of the annual maximum flood ($CV_{AMF}$).

Table 5.1: Factors that influence the particle-size distribution of a stream bed.

| Particle size distribution of bed surface varies with... | Equilibrium at bed surface $\frac{dD_{50}}{dt} = 0$ | Fining of bed surface $\frac{dD_{50}}{dt} < 0$ | Coarsening of bed surface $\frac{dD_{50}}{dt} > 0$ |
|---|---|---|---|
| *At intermediate to low sediment transport rates* | | | |
| Rate and particle size distribution of sediment supply | Steady-state transport in sediment feed flume (Kuhnle and Southard 1988)<br>Braided gravel streams (Leopold and Wolman 1957) | Increased hillslope input (Wolman and Schick 1967, Rice and Church 1996) | Reducing sediment supply rate in flume (Kuhnle and Southard 1988; Dietrich et al. 1989; Lisle et al. 1993)<br>Degrading channel downstream of dams (Williams and Wolman 1984) |
| Applied shear stress and initial particle size distribution of bed surface | Gravel bed streams (Lane 1955, Henderson 1963, Allen 1965, Shih and Komar 1990; Parker et al. 1982; Parker and Sutherland 1990)<br>Sand bed canals (Lacey 1929)<br>Steady-state transport in recirculating flume (Wilcock and McArdell 1993)<br>Immobile surface layer in clear water flume (Gessler 1970; Little and Mayer 1976; Shen and Lu 1983; Chin et al. 1994) | | Armor development after storms in gravel bed streams (Gomez 1983a)<br>Degrading bed in flume (Garde et al. 1977) |
| Channel morphology and discharge | Downstream fining (Wathen et al. 1995; Seal et al. 1997)<br>Bedrock channels | Change over time in base level (Leopold and Bull 1979), sinuousity (Mackin 1948; Rees 1957), or upstream sediment storage (Rice and Church 1996) | |
| Lithology of stream basin | Gravel bed streams (Hack 1957, Wolcott 1988) | | |
| *At high sediment transport rates* | | | |
| Particle size distribution of sediment supply | Steady-state transport in recirculating (Wilcock and McArdell 1993) or sediment feed flume (Kuhnle and Southard 1988)<br>Gravel bed streams (Parker et al. 1982) | Breaching of a surface armor in gravel bed stream (Gomez 1983a) or flume (Little and Mayer 1976)<br>Aggrading channels (Lisle 1982) | Increasing gravel fraction of bimodal sediment in flume (Ikeda and Iseya 1988) |

Table 5.2: Characteristics of Puget Lowland streams in the disturbance analysis.

| | | Drainage area | Road density | Stream gaging period of record | Slope | Median of particle size distribution of bar surface material | 84th percentile of particle size distribution of bar surface material |
|---|---|---|---|---|---|---|---|
| | | $km^2$ | $km^{-1}$ | Water Years | | m | m |
| Big Beef Creek near mouth | | 35 | 2.1 | 1994 - 2000 | 0.011 | 0.045 | 0.090 |
| Huge Creek near mouth | | 16 | 2.5 | 1989 - 1998 | 0.012 | 0.030 | 0.046 |
| Rock Creek @ pipeline crossing | A | 32 | 2.7 | 1971 - 1976, 1995 - 1998 | 0.024 | 0.071 | 0.143 |
| | B | | | | 0.018 | 0.043 | 0.089 |
| Covington Creek near mouth | | 49 | 4.0 | 1988 - 1994 | 0.011 | 0.046 | 0.089 |
| Big Bear Creek @ NE 133rd St. | | 32 | 4.4 | 1988 - 1991, 1993 - 1998 | 0.007 | 0.037 | 0.081 |
| Newaukum Creek near mouth | | 70 | | 1989 - 1998 | 0.014 | 0.065 | 0.175 |
| May Creek @ Coal Creek Parkway | | 22 | 4.0 | 1991 - 1998 | 0.012 | 0.045 | 0.110 |
| May Creek near mouth | A | 34 | 5.0 | 1989 - 1998 | 0.009 | 0.040 | 0.083 |
| | B | | | | 0.012 | 0.031 | 0.070 |
| Jenkins Creek near mouth | A | 38 | 5.4 | 1989 - 1998 | 0.005 | 0.031 | 0.059 |
| | B | | | | 0.007 | 0.048 | 0.090 |
| Soos Creek near mouth | | 171 | 4.7 | 1989 - 1998 | 0.003 | 0.033 | 0.054 |
| Swamp Creek at Filbert Road | | 24 | 7.4 | 1989 - 1998 | 0.010 | 0.040 | 0.083 |
| Swamp Creek near Kenmore | | 59 | 7.9 | 1981 - 1990 | 0.005 | 0.021 | 0.042 |
| Des Moines Creek near mouth | | 14 | 7.9 | 1992 - 1998 | 0.017 | 0.037 | 0.059 |
| Leach Creek | | 12 | 9.9 | 1989 - 1998 | 0.013 | 0.030 | 0.068 |
| Miller Creek near mouth | A | 20 | 10.6 | 1989 - 1998 | 0.010 | 0.029 | 0.056 |
| | B | | | | 0.006 | 0.022 | 0.060 |

Table 5.3: Hydraulic conditions for reference discharges.

| | | Q | R | Manning's n | |
|---|---|---|---|---|---|
| Mean discharge | | $m^3/s$ | | low | high |
| Big Beef Creek near mouth | | 1.7 | 0.29 | 0.054 | 0.081 |
| Huge Creek near mouth | | 0.3 | 0.14 | 0.044 | 0.065 |
| Rock Creek @ pipeline crossing | A | 0.5 | 0.14 | 0.091 | 0.136 |
| | B | 0.5 | 0.11 | 0.057 | 0.086 |
| Covington Creek near mouth | | 0.8 | 0.14 | 0.046 | 0.069 |
| Big Bear Creek @ NE 133rd St. | | 0.8 | 0.23 | 0.043 | 0.064 |
| Newaukum Creek near mouth | | 1.7 | 0.30 | 0.085 | 0.128 |
| May Creek @ Coal Creek Parkway | | 0.7 | 0.20 | 0.080 [a] | |
| May Creek near mouth | A | 0.7 | 0.10 | 0.028 [a] | |
| | B | 0.4 | 0.11 | 0.050 [a] | |
| Jenkins Creek near mouth | A | 1.1 | 0.22 | 0.052 [a] | |
| | B | 1.1 | 0.26 | 0.060 [a] | |
| Soos Creek near mouth | | 3.5 | 0.46 | 0.034 | 0.050 |
| Swamp Creek at Filbert Road | | 0.4 | 0.15 | 0.040 [a] | |
| Swamp Creek near Kenmore | | 1.0 | 0.19 | 0.033 [a] | |
| Des Moines Creek near mouth | | 0.1 | 0.06 | 0.050 | 0.075 |
| Leach Creek | | 0.2 | 0.10 | 0.060 | 0.090 |
| Miller Creek near mouth | A | 0.2 | 0.10 | 0.047 | 0.070 |
| | B | 0.2 | 0.13 | 0.043 | 0.064 |
| Discharge exceeded 10% of the time | | | | | |
| Big Beef Creek near mouth | | 3.1 | 0.34 | 0.051 | 0.076 |
| Huge Creek near mouth | | 0.5 | 0.17 | 0.042 | 0.063 |
| Rock Creek @ pipeline crossing | A | 1.1 | 0.16 | 0.072 | 0.108 |
| | B | 1.1 | 0.13 | 0.049 | 0.074 |
| Covington Creek near mouth | | 2.2 | 0.22 | 0.044 | 0.066 |
| Big Bear Creek @ NE 133rd St. | | 1.8 | 0.30 | 0.041 | 0.061 |
| Newaukum Creek near mouth | | 3.3 | 0.37 | 0.079 | 0.119 |
| May Creek @ Coal Creek Parkway | | 1.5 | 0.32 | 0.080 [a] | |
| May Creek near mouth | A | 1.5 | 0.16 | 0.028 [a] | |
| | B | 1.5 | 0.17 | 0.050 [a] | |
| Jenkins Creek near mouth | A | 2.1 | 0.31 | 0.052 [a] | |
| | B | 2.1 | 0.39 | 0.060 [a] | |
| Soos Creek near mouth | | 8.0 | 0.61 | 0.033 | 0.049 |
| Swamp Creek at Filbert Road | | 1.0 | 0.25 | 0.040 [a] | |
| Swamp Creek near Kenmore | | 2.5 | 0.32 | 0.033 [a] | |
| Des Moines Creek near mouth | | 0.3 | 0.08 | 0.046 | 0.069 |
| Leach Creek | | 0.3 | 0.12 | 0.050 | 0.075 |
| Miller Creek near mouth | A | 0.5 | 0.14 | 0.043 | 0.065 |
| | B | 0.5 | 0.17 | 0.039 | 0.058 |

[a] Manning's n calculated from velocity measurements

Table 5.3 continued.

| | | Q | R | Manning's n | |
|---|---|---|---|---|---|
| Discharge exceeded 5% of the time | | $m^3/s$ | | low | high |
| Big Beef Creek near mouth | | 4.6 | 0.36 | 0.048 | 0.072 |
| Huge Creek near mouth | | 0.8 | 0.23 | 0.041 | 0.062 |
| Rock Creek @ pipeline crossing | A | 1.4 | 0.19 | 0.068 | 0.103 |
| | B | 1.4 | 0.15 | 0.048 | 0.072 |
| Covington Creek near mouth | | 2.7 | 0.28 | 0.044 | 0.065 |
| Big Bear Creek @ NE 133rd St. | | 2.6 | 0.40 | 0.040 | 0.060 |
| Newaukum Creek near mouth | | 4.8 | 0.48 | 0.078 | 0.117 |
| May Creek @ Coal Creek Parkway | | 1.3 | 0.21 | 0.050 [a] | |
| May Creek near mouth | A | 2.1 | 0.38 | 0.080 [a] | |
| | B | 2.1 | 0.19 | 0.028 [a] | |
| Jenkins Creek near mouth | A | 2.7 | 0.35 | 0.052 [a] | |
| | B | 2.7 | 0.44 | 0.060 [a] | |
| Soos Creek near mouth | | 10.3 | 0.77 | 0.032 | 0.049 |
| Swamp Creek at Filbert Road | | 1.5 | 0.30 | 0.040 [a] | |
| Swamp Creek near Kenmore | | 3.4 | 0.38 | 0.033 [a] | |
| Des Moines Creek near mouth | | 0.4 | 0.12 | 0.045 | 0.068 |
| Leach Creek | | 0.5 | 0.13 | 0.046 | 0.068 |
| Miller Creek near mouth | A | 0.7 | 0.18 | 0.042 | 0.063 |
| | B | 0.7 | 0.22 | 0.037 | 0.056 |
| Median annual maximum flood | | | | | |
| Big Beef Creek near mouth | | 18.1 | 0.61 | 0.044 | 0.065 |
| Huge Creek near mouth | | 4.9 | 0.45 | 0.040 | 0.060 |
| Rock Creek @ pipeline crossing | A | 3.8 | 0.27 | 0.058 | 0.087 |
| | B | 3.8 | 0.25 | 0.046 | 0.069 |
| Covington Creek near mouth | | 4.2 | 0.36 | 0.043 | 0.065 |
| Big Bear Creek @ NE 133rd St. | | 4.2 | 0.50 | 0.039 | 0.059 |
| Newaukum Creek near mouth | | 16.5 | 0.70 | 0.050 | 0.075 |
| May Creek @ Coal Creek Parkway | | 6.1 | 0.47 | 0.045 [a] | |
| May Creek near mouth | A | 6.9 | 0.71 | 0.041 [a] | |
| | B | 6.9 | 0.37 | 0.028 [a] | |
| Jenkins Creek near mouth | A | 4.7 | 0.48 | 0.035 [a] | |
| | B | 4.7 | 0.61 | 0.040 [a] | |
| Soos Creek near mouth | | 18.6 | 1.03 | 0.032 | 0.048 |
| Swamp Creek at Filbert Road | | 6.5 | 0.60 | 0.042 [a] | |
| Swamp Creek near Kenmore | | 11.4 | 0.69 | 0.033 [a] | |
| Des Moines Creek near mouth | | 4.9 | 0.41 | 0.044 | 0.066 |
| Leach Creek | | 2.4 | 0.28 | 0.043 | 0.065 |
| Miller Creek near mouth | A | 6.2 | 0.47 | 0.039 | 0.059 |
| | B | 6.2 | 0.65 | 0.036 | 0.054 |

[a] Manning's n calculated from velocity measurements

# Chapter 6: Conclusions

In the course of urban development, forests are cut down, hillslopes graded, and drainage networks constructed. These changes redistribute rainfall that would have been stored in the forest canopy, soil column, and wetlands and other depressions to overland and shallow subsurface pathways that quickly deliver stormwater to streams. As a result, stream flow increases rapidly during storms, attains higher peak discharge rates, and falls rapidly after rain has ceased in urban streams. Over longer periods of time, urban streams have more frequent peaks in discharge than suburban streams, particularly under dry antecedent conditions.

While biologic conditions may recover quickly after individual disturbances such as a flood or a drought, annual and inter-annual stream flow patterns are likely to have persistent biologic effects on stream ecosystems. Exploratory data analysis of stream flow records for the Puget Lowland, Washington, revealed three differences between urban and suburban stream flow patterns at annual and inter-annual scales: (1) fewer days in a year that the mean discharge rate is exceeded; (2) lower variation in annual maximum floods; and (3) a shorter cumulative duration that the discharge rate of a flood of a given frequency is exceeded. These stream flow changes provided the basis for a general hypothesis that the frequency and extent of hydrologic disturbances (i.e., droughts and floods) are higher in urban streams than suburban streams.

Droughts represent a seasonal form of hydrologic disturbance in the Puget Lowland, particularly for ephemeral streams, which have no surface flow during the summer. Points along a stream channel with a drainage area of 1.2 $km^2$ had a 50% probability of flowing perennially in this investigation, though the drainage areas of individual streams, both ephemeral and perennial, ranged from less than 0.1 $km^2$ to more than 10 $km^2$. The total length of perennial stream (L) in a stream basin was generally related to the basin's area (A) by $L = 0.4\ A + 0.8$ where L has units of km and A has units of $km^2$. The predicted stream lengths from this equation had a root mean square error of

9.8% relative to observed lengths. The extent of perennial streams in a basin was not related to road density or to measures of physiographic conditions (basin shape, valley relief, valley slope). Urban development may, however, reduce the period of continuous flow in ephemeral streams such that the duration and frequency of droughts increases. The effects of such changes would be to reduce the availability of aquatic habitat during spring and early summer rather than later in the summer.

Floods represent a form of high flow disturbance when stream bed material is entrained by the flood. The fractional extent of disturbance, or partial entrainment ($PE_{bar}$), of a gravel bar during a flood ranges from 0 to 1 and is related to the peak dimensionless shear stress of the flood by the equation $PE_{bar} = 12.5\ (\tau_0{}^* - 0.045)$ which has a root mean square error of 0.099 based on 104 bed tag inventories conducted at seven gravel bars. In part, the deviation between observed and predicted values of $PE_{bar}$ reflects changes in the strength of the stream bed from storm-to-storm.

$PE_{bar}$ provides an average probability of bed material entrainment for a gravel bar, though the probability of bed material entrainment during a flood is only approximately uniform over the surface a gravel bar. Likewise, over periods of multiple floods, individual locations on a gravel bar are consistently more or less stable than the bar as a whole. At reach-scales, higher entrainment probabilities were observed at the steep foreset slope (downstream face) of a bar, in a section of converging flow, or places where the channel is actively widening or migrating while lower probabilities were observed upstream of bar crests in slightly wider sections.

Stream bed disturbance regimes depend on the magnitude of floods relative to intermediate flows because intermediate flows determine, in part, the particle size distribution of the bed surface which has a primary influence on the strength of the bed surface. The discharge rate exceeded 5% of the time ($Q_{0.05}$) provided a robust reference discharge for the suite of flows determining the particle size distribution of the bed surface in moderately transport-limited (i.e., pool-riffle and plane-bed) gravel-bed streams.

The fraction of a gravel bar's surface entrained during a flood is predicted to vary with the ratio of the flood's peak discharge rate to $Q_{0.05}$. More than 30% of a gravel's bar surface is likely to be disturbed during a flood with a peak discharge rate that is 3 times $Q_{0.05}$. More than 60% of a bar's surface is likely to be disturbed during a flood that 6 times greater than $Q_{0.05}$. The influence of stream discharge on both the strength of the stream bed and the magnitude of the applied force of a flood provide hydrologic control over stream bed disturbance patterns in gravel bed streams.

The ratio of a flood with a specific frequency (e.g, $Q_{2yr}$) to $Q_{0.05}$ is generally higher in streams with high peak discharge rates and rapid storm flow recession whether a result of urban development of physiographic conditions. Urban development promotes geomorphic instability in gravel bed streams when the magnitude of frequent floods increase, the rate of storm flow recession increases, and the magnitude of intermediate flows decreases. For the Puget Lowland urban streams analyzed here, the peak discharge rate of the median annual maximum flood was more than 3 times $Q_{0.05}$ indicating frequent and extensive bed disturbance is likely in these gravel-bed streams. Likewise, gravel-bed streams with lower levels of development may have frequent and extensive bed disturbance if the magnitude of frequent floods is large relative to $Q_{0.05}$.

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