Sediment-transporting flows in headwater streams

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ABSTRACT

The equilibrium alluvial stream channel has a geometry that allows it to pass the water and sediment supplied from the watershed. At the same time, the equilibrium alluvial channel is built and maintained by the flows and sediment delivered to it. A prerequisite for understanding the creation of the equilibrium channel is an understanding of the sediment conveyance and competence of the flows the channel receives.

This study describes the bed-load transport regime as it is linked to hydrology and geomorphology in 23 headwater gravel-bed streams in snowmelt-dominated parts of central and northern Idaho. At sites, drainage areas range from 1.29 to 381 km^2, stream gradients range from 0.0042 to 0.0747, and median bed surface particle sizes range from 4 to 207 mm. Stream architecture includes riffle-pool, planar, and step-pool beds.

The bed load is much finer than the surface and subsurface material, suggesting selective transport of the finer sizes. Nonetheless, the majority of the load is sand at all flow discharges. Progressively coarser sediment was collected as flow discharge increased, and painted rock experiments documented the transport of coarser particles at higher discharges. The supply of sediment to the streams appears limited, as indicated by observed clockwise hysteresis in bed-load transport rates during each spring snowmelt and by the coarse surface armor observed at sites.

Flows above bankfull discharge move 37% of the bed load, whereas flows between mean annual discharge and bankfull move 57% of the bed load. The bed-load effective discharge has a recurrence interval that averages 1.4 yr and the magnitude of effective discharge averages 80% of bankfull discharge. The recurrence interval of bankfull discharge averages 2.0 yr. The ratio of effective discharge to bankfull discharge is independent of basin size, grain size, and gradient, although the ratio increases with the relative magnitude of large infrequent events.

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INTRODUCTION

It has been said that alluvial streams are the architects of their own geometry (Leopold, 1994). This means that the channel has created an equilibrium form that allows it to pass the water and sediment supplied to it from upstream given the constraints of climate, topography, and geology. A key to understanding the creation of the equilibrium channel characteristics, or the maintenance of these characteristics, is to be able to relate the sediment moved to the magnitude and frequency of flows of water.

A stream can carry a range of sediment sizes ranging from silt to sand to gravel. These particles move as wash load, suspended load, and/or bed load, depending upon stream energy. The sediment sizes most relevant to the creation or maintenance of the channel geometry are those of the material making up the bed and banks. In upland channels, the coarse nature of the load means that much of the material moves along the channel as bed load. The bed and banks are composed of this coarse material, hence the channel geometry is related to the magnitude and frequency of flows transporting bed load. In more lowland, gentle-gradient channels, the load is typically finer and most of the sediment (>95%) is moved in suspension (e.g., Emmett, 1994, cited *in* Nash, 1994). The channel bed and particularly the banks are often composed of materials moving as bed load and suspended load. For these channels, the channel geometry may be related to the magnitude of flows transporting suspended load and/or the finer bed load.

Most studies of the amount of sediment moved by different flows, especially as they relate to channel geometry, have focused on lowland rivers (e.g., Wolman and Miller, 1960) or on the suspended component of the sediment load (e.g., Andrews, 1980; Ashmore and Day, 1988; Nash, 1994). Many of these studies, but not all, have concluded that over the long term the most sediment is transported by flows that are near bankfull (Wolman and Miller, 1960; Andrews, 1980; Andrews and Nankervis, 1995) and have a recurrence interval of about 1.5 yr. The flow moving the most sediment over the long term has been termed the effective discharge (Wolman and Miller, 1960). Large flows move more sediment on a day by day basis, but occur less frequently than the flows that just fill the channel, such that in total, the bankfull discharge is thought to be the most effective discharge in transporting sediment, and by extension, the most important in maintaining channel form and function.

Headwater streams are typically coarser and steeper than lowland channels. As a consequence, headwater streams may transport more of their load as bed load. Despite these differences, the assumption is that smaller headwater channels behave in a manner similar to larger lowland streams. There is limited support for such a view. Emmett (1975) found that bankfull discharge in a variety of streams draining to the Salmon River in Idaho had a recurrence interval of ~1.5 yr, a frequency equivalent to effective discharge in lowland streams. Carling (1988) and Pickup and Warner (1976) found that the effective event in small basins was about that reported for larger basins. There is no essential difference in the relationship between effective discharge and bankfull discharge for small and large basins in the data of Andrews and Nankervis (1995). Nash (1994) found no relation between effective discharge and basin size or precipitation. However, there is evidence that more-extreme, less-frequent events may play a more important geo-

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morphic role in smaller basins than in larger basins (Costa, 1974; Grant et al., 1990). Ashmore and Day (1988) found that the effective discharge for suspended sediment was a rarer and more extreme event in smaller drainage basins. Pitlick (1988) found that the effective discharge for predicted bed load at seven gaging stations in northern California became an increasingly rare event as the gradient and bed material size increased. For slopes less than 1% the recurrence interval was less than 2 yr, whereas at slopes of 2% the effective discharge had a recurrence interval exceeding 75 yr.

The objective of this study was to investigate the range of streamflows that are important in transporting sediment and thus in forming and/or maintaining channel geometry and function in headwater streams in the snowmelt-dominated region of the intermountain west. For the purpose of this study, headwater streams are defined as those streams with the majority of their drainage at elevations above 1200 m and having drainage areas less than 400 km^2.

FIELD SITES

At 23 headwater streams in central and northern Idaho, we collected or there existed, streamflow and sediment transport information. The U.S. Forest Service has operated a series of stream gages in headwater channels for which there are records of discharge and sediment transport. We included in the study 17 sites for which there was 5–15 yr of streamflow and several years of sediment transport information. Near Stanley, Idaho, the U.S. Geological Survey operates several gages in headwater streams for which there are discharge records for 24 or more years. Near these three U.S. Geological Survey (USGS) gages, we began flow measurements in three additional streams. At the USGS gages, and at the additional sites, we made sediment transport measurements. Figure 1 shows locations of sites and Table 1 provides a summary of site attributes.

The 23 streams studied are in the northern Rocky Mountain physiographic province (Fenneman, 1931) in Idaho (Fig. 1). All streams are within the Columbia River basin and, with the exception of Catspur Creek in northern Idaho and Eggers Creek in central Idaho, all are tributaries of the Salmon River or the Clearwater River. Bedrock geology includes the Idaho Batholith, volcanic rock, and metasedimentary rock (Maley, 1987). At least one site was glaciated: Fourth of July Creek (Williams, 1961). Bedrock is exposed locally in the bed and/or banks of several streams. Nonetheless, all sites have substantial alluvial cover that allows for the development of a self-formed channel. The channel reaches studied include A, B, and C type channels (Rosgen, 1994) and the channel architecture at sites is either riffle pool, planar bed, or step pool (Montgomery and Buffington, 1997). There are small irrigation diversions at Fourth of July Creek, Herd Creek, and Valley Creek, but only at Fourth of July Creek does the majority of the diverted water not return to the channel. Diversion, however, does not appear to occur during peak flow periods. Vegetation near the channels ranges from sage and grass, to willows, to conifer forest. Precipitation in the basins is as low as 25 cm to as high as 180 cm (University of Idaho, 1995).

METHODS

Field and Laboratory Methods

Each of the study sites was surveyed and described. Maps over a reach of typically 20 or more channel widths were made to show major planform features and the locations of cross sections and gages. At one channel-width intervals, the water surface along each bank and the channel bed at the channel centerline were surveyed. The elevation of bankfull along the channel was also surveyed. Bankfull was identified as the level corresponding to the top of the channel banks and at the level of the relatively flat, frequently inundated surface adjacent to the channel built by the channel in its current hydrologic regime. Bankfull was surveyed only where there was a distinct morphologic feature corresponding to the flood plain. Surrogates for bankfull such as vegetation, sediment size, or stain lines were not surveyed, although they were useful at times in confirming subtle bankfull features. Care was taken to avoid nearbank slumps. At some sites in 1995, the water level at a stage near bankfull was surveyed along one bank in the vicinity of the gage. Approximately four (two to six) cross sections showing the channel and valley features were established and permanently marked with lengths of rebar.

A stage plate was installed, or the existing stage plate was located, and surveyed at all sites. At three sites without an operating gage, a pressure transducer was installed to monitor continuously the stage. At most U.S. Forest Service (USFS) sites and at the three sites where we installed the pressure transducers, stage was monitored from mid-April to late October. Stream ice prevents stage measurement during winter at most USFS sites: stage is monitored year-round at the USGS gages.

Flow velocity was measured with Price and Pygmy current meters. At low flows or in smaller channels, these current meters were attached to hand-held topset wading rods. At high flows in the larger channels, the meters were suspended from cables or from bridges. Velocity was measured for 60 s typically at 20 equally spaced points across the channel. The velocity measurements and local depth measurements were integrated across the channel to give the flow discharge using the midsection method used by the USGS and shown by Hipolito and Loureiro (1988) to give the most precise measurements of total discharge through a section. Rating curves specifying the relationship between stage and water discharge were built for each site based upon tens to hundreds of stage-discharge pairs.

Bankfull stage was determined for sites from the stage reading associated with intersection with the stage plate of the equation describing the longitudinal elevation of bankfull level. The bankfull discharge was determined from the stage-discharge curve for the gage. For a few sites, projection of the bankfull elevation to the gage produced unlikely results because of local hydraulics near the gage. In these cases, either the difference between the surveyed high water and local bankfull elevation or a portion of the longitudinal profile were used to estimate bankfull stage.

Bed load was measured by the single equal width increment method (Edwards and Glysson, 1988). A 7.6 cm Helley-Smith sampler with a 0.25 mm mesh collection bag was placed on the bed for a uniform period (30 or 60 s) at 10–25 equally spaced intervals across the channel. Two bed-load samples were typically collected during each site visit. The Helley-Smith sampler was either hand-held or suspended from a cable. The particle-size distribution of each bed-load sample was determined by sieving. For selected sites in 1994 and 1995, the intermediate axis (b-axis) of the coarsest particle in samples was measured to 1 mm with a caliper.

Suspended sediment concentration was measured with depth-integrating samplers using the equal discharge increment method (Edwards and Glysson, 1988). A water sample was taken from each of five approximately equal discharge increments. Samples were stored in the dark until suspended-load concentration was determined. Concentration was established to 0.01 mg/L for the fractions coarser and finer than 0.063 mm.

Two transects containing painted rocks were established at selected sites. Particles ranging in size from about D_{25} to D_{90} were dried and painted. The subscripts refer to the percentage that is finer. At each of 20 equally spaced points across the channel, a painted rock was placed on the bed in a pocket between grains with the shortest axis (c-axis) pointing upward. The particle that was placed was chosen randomly from the pool of previously painted grains. Movement of any of the grains was noted during site visits when the flow was clear enough to permit observation of the bed. After the spring snowmelt, those painted rocks that could be located were retrieved for identification and determination of the total downstream displacement of grains.

Surface pebble counts were made at each site (Wolman, 1954). Particles were selected for measurement of their intermediate axis (b-axis) at approximately equally spaced intervals along four or five transects across the bankfull width of the channel. The exact particle measured at each equal increment was chosen blindly. A minimum of 100 particles (but often more) were measured for each pebble count—the exact number depended upon the number of transects and transect width. Each transect was completed once it was begun.

Subsurface particle-size measurements were made at selected sites by USFS personnel. A 55-gallon drum was cut in half, its end removed, and placed on the bed. Surface grains were removed to a depth of about the median diameter of surface grains and then the material below the surface was sampled. The principal dimensions of grains coarser than 64 mm were measured; grains between 16 and 64 mm were wet sieved in the field, and grains finer than 16 mm were sieved in the laboratory. Surface particle-size measurements were also made at locations where subsurface measurements were made.

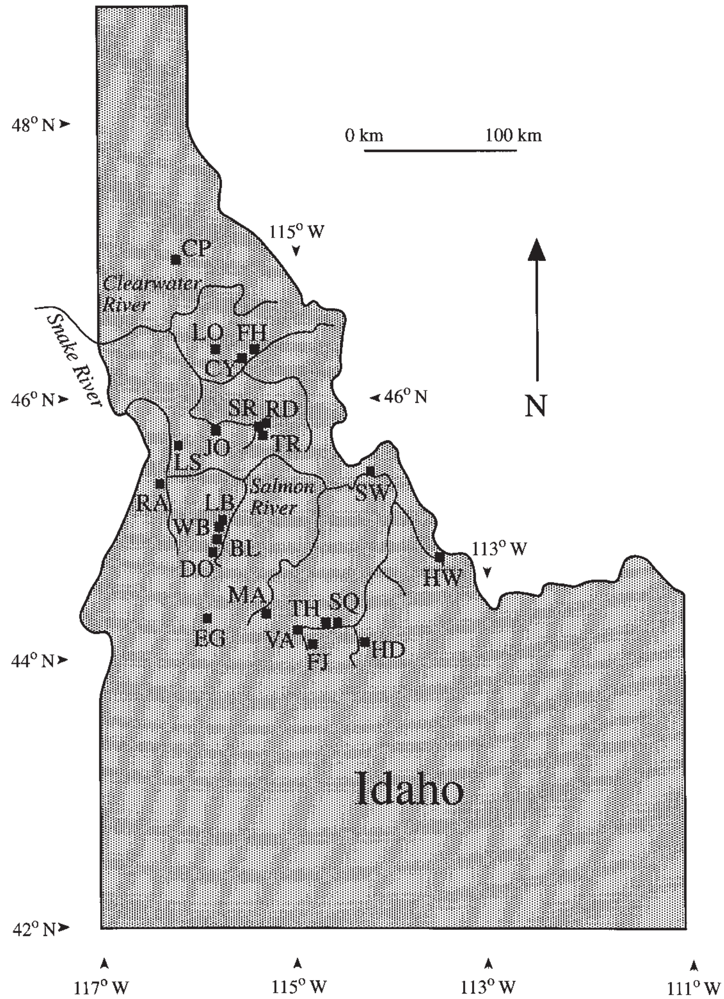


Figure 1. Locations of sites: BL—Blackmare Creek, CP—Catspur Creek, CY—Canyon Creek, DO—Dollar Creek, EG—Eggers Creek, FH—Fish Creek, FJ—Fourth of July Creek, HD—Herd Creek, HW—Hawley Creek, JO—Johns Creek, LB—Little Buckhorn Creek, LS—Little Slate Creek, LO—Lolo Creek, MA—Marsh Creek, RA—Rapid Creek, RD—Red River, SQ—Squaw Creek (USGS/CWRU), SR—South Fork Red River, SW—Squaw Creek (Salmon National Forest), TH—Thompson Creek, TR—Trapper Creek, VA—Valley Creek, and WB—West Fork Buckhorn Creek.

Analysis of Long-Term Hydrologic Records

Flow-duration curves describe the percent of time a given discharge is equaled or exceeded during the period of record. Two important issues arise with respect to the generation of these curves—the completeness of the existing record and the representativeness of the period of record. Most stream gages (other than USGS gages) in this region are operated from snowmelt to freeze-up, which is approximately early April to the end of October. To estimate the flows during periods when gages were not operating, either due to seasonal shutdown or to equipment malfunction, the gages with partial records were correlated to nearby gages that operated year-round. The second issue, the representativeness of the existing record, arises when the flow record is insufficiently long to include the typical range of flows associated with climatologic variation. To estimate the duration of flows over the long term, sites with short-period records were correlated to sites with a longer period of record.

A method for estimating these unmeasured flows (Moog et al., 1999) is a modification of a method suggested by Hirsch (1982). The characteristics of the distribution associated with the record from a base station with a long flow record are transferred to the short-record station to augment the short record. The resulting time series of daily discharges is intended to be a reasonable estimate of an actual long record for the short-record site were it to exist.

For paired data having positive correlation, the maintenance of variance estimation (MOVE.1) equation (Hirsch, 1982) is written

$$\hat{y}(i) = m(y_1) + \frac{S(y_1)}{S(x_1)}\left(x(i) - m(x_1)\right), \quad (1)$$

where y is the short record, x is the base record, subscript 1 indicates a subset consisting of concurrent dates, S is the standard deviation, m is the mean, and ˆ indicates an estimated value. Note that there is no explicit pairing of values between x and y. This is equivalent to an ordinary least squares regression (r), with r set to 1. Consequently, the mean and variance of estimated y for the concurrent period equal the respective values for y_1.

The MOVE.1 technique is a linear function. Scatterplots of concurrent data sets used to estimate the slope and intercept of this function should, therefore, display an essentially linear pattern. For a variety of reasons, the best available long-term stations may show scatterplots of the concurrent data sets having curved point patterns and uneven variation over the range of flows of the long-term station. An approach to this situation is to transform the discharges from

TABLE 1. CHARACTERISTICS OF SITES

Stream	Data collector	Period used in analysis	Drainage area (km^2)	Slope	D_{50} (mm)	Rosgen stream type	Channel architecture	Bankfull Discharge (m^3/s)	Bankfull Width (m)	Bankfull Depth (m)
Blackmare Creek	USGS/USFS	1990–1994	46.1	0.0299	95	B3	Planar	4.73	10.8	0.42
Canyon Creek	Clearwater NF	1992–1995	49.7	0.0508	140	B3a	Step pool	6.48	10.7	0.43
Catspur Creek	Panhandle NF	1987–1995	28.0	0.0111	27	C4	Riffle pool	2.36	4.91	0.35
Dollar Creek	USGS/ USFS	1990–1994	42.7	0.0146	77	B3c	Planar	6.43	8.85	0.44
Eggers Creek	USFS	1975–1984	1.29	0.0747	4	B4a	Planar	0.0478	1.44	0.10
Fish Creek	Clearwater NF	1975–1995	225	0.0266	140	A3b	Planar	20.9	18.8	0.80
Fourth of July Creek	CWRU	1994–1995	44.2	0.0202	51	B4c	Planar	3.88	6.65	0.39
Hawley Creek	Salmon NF	1989–1995	109	0.0184	40	B4	Planar	1.33	6.73	0.28
Herd Creek	CWRU	1994–1995	308	0.0077	67	C3	Planar	5.46	8.15	0.49
Johns Creek	Nez Perce NF	1986–1995	293	0.0207	180	B3	Planar	49.0	19.5	1.12
Little Buckhorn Creek	USGS/ USFS	1990–1994	15.5	0.0509	28	A4	Step pool	N.A.*	N.A.*	N.A.*
Little Slate Creek	Nez Perce NF	1986–1995	162	0.0268	207	B3	Planar	12.2	14.1	0.68
Lolo Creek	Clearwater NF	1982–1995	106	0.0097	90	B3c	Planar	11.7	16.8	0.62
Marsh Creek	CWRU	1994–1995	196	0.0060	56	C4	Planar	20.8	19.0	0.60
Rapid River	Nez Perce NF	1986–1995	280	0.0108	94	B3c	Planar	17.7	24.3	0.54
Red River	Nez Perce NF	1986–1995	129	0.0059	68	C3	Riffle pool	9.34	11.5	0.54
SF Red River	Nez Perce NF	1986–1995	98.9	0.0140	86	C3	Planar	7.25	10.9	0.44
Squaw Creek	Salmon NF	1989–1995	37.0	0.0240	27	B4	Planar	0.623	3.53	0.24
Squaw Creek	USGS/CWRU	1971–1995	205	0.0100	43	C4	Step pool	5.12	11.0	0.39
Thompson Creek	USGS/CWRU	1973–1995	56.4	0.0153	66	C4b	Planar	2.48	6.87	0.24
Trapper Creek	Nez Perce NF	1976–1995	20.7	0.0414	67	B3a	Step pool	2.56	6.43	0.36
Valley Creek	USGS/CWRU	1921–1972, 1992–1995	381	0.0042	49	C4	Planar	28.6	32.0	0.76
WF Buckhorn Creek	USGS/ USFS	1990–1994	58.5	0.0320	180	B3	Planar	5.72	9.35	0.55

Notes: USGS—United States Geological Survey; USFS—United States Forest Service; NF—National Forest; D_{50}—median diameter of surface particles.
*N.A.—Bankfull could not be identified at this site.

one or both stations to more closely approximate a linear function-constant variance relationship, apply the MOVE.1 procedure to the transformed data, and then back-transform to original coordinates. Hirsch (1982) used this general approach by taking the logarithm of the discharges.

A widely used family of transformations is the power transformations discussed by Box and Cox (1964). The scaled system of power transformations can be written as:

$$v^{(\lambda)} = \left\{ \frac{\left(v^{\lambda} - 1\right) / \lambda \; if \; \lambda \neq 0}{\ln \; \lambda \; if \; \lambda = 0} \right\} \qquad (2)$$

in which $v^{(\lambda)}$ is the transformed value of v and λ is the transformation parameter. A useful property of power transformations is that increasing or decreasing λ produces gradual changes in the effects on the point pattern. This system includes most common transformations; reciprocal, logarithm, square root, no transformation, and square. Another useful property of the transformations is that they allow a better representation of the relationship between stations across the range of flows.

We developed a program to visually select the best values of λ to fit the concurrent data sets. The base station selected for extending short records as well as the length of the respective records are listed in Table 2. In some cases, the actual record length was greater than that indicated, but was not used because of concerns regarding data quality or methodology.

Figure 2 provides an example of a 10 yr portion of the discharge record (1986–1995) from Red River based upon measured stream flows and extended streamflows. The bold line represents measured flows, and the finer line represents extended flows. There is general correspondence between measured and extended streamflows, as can be seen in Figure 2.

Flood Frequency

The recurrence interval of instantaneous peak flows was calculated for most sites by the Log Pearson III method (Water Resources Council, 1967). At Fourth of July, Herd, and Marsh Creeks, there were only 2 yr of measured record, which is too short for a meaningful estimate of flood frequency using the Log Pearson III method. For these streams the annual peaks of the mean daily flows during 1994 and 1995 were assigned the frequency of the corresponding flows at Valley Creek (the stream serving as the base station) and the frequency of bankfull and effective discharge interpolated between these values.

RESULTS

Geometry and Hydraulic Geometry

The headwater channels investigated span a large range of geometry and particle size (Table 1). Stream gradients of the reaches range from 0.0042 to 0.0747. All channels have a gravel bed: the median surface particle size ranges from 4 to 207 mm (Table 1). The bankfull width (W_B, in m) and bankfull depth (D_B, in m) increase with drainage area (DA, in km^2) (Fig. 3; Table 1). The equations for bankfull width and depth are:

$$W_B = 1.233 \text{ DA}^{0.473} \; r^2 = 0.752 \qquad (3)$$

$$D_B = 0.106 \text{ DA}^{0.326} \; r^2 = 0.673. \qquad (4)$$

Bankfull discharge (Q_B) increases with drainage area (Fig. 4; Table 1):

$$Q_B = 0.0628 \text{ DA}^{1.010} \; r^2 = 0.748. \qquad (5)$$

The increase in bankfull discharge is nearly linear (exponent equals 1.010) with drainage area. This power function was chosen over a linear function to represent the data because of its ability to describe the relationship in small drainage basins and in larger basins (W.W. Emmett, 1996, unreported data). There is large scatter about the line; several of the relatively low bankfull discharge values—the three lowest values above 100 km^2 (Hawley Creek, Squaw Creek near Clayton, and Herd Creek)—are in relatively drier basins.

Viewing each of the streams as headwater extensions of the Columbia River, downstream hydraulic geometry at bankfull discharge (Q_B, in m^3 s^1) for the sites is described by the following equations:

$$W^B = 4.680Q^{B0.446} \; r^2 = 0.913 \qquad (6)$$

$$D^B = 0.256Q^{B0.328} \; r^2 = 0.930 \qquad (7)$$

$$V_B = 0.891{Q_B}^{0.210} r^2 = 0.692. \qquad (8)$$

TABLE 2. FLOW EXTENSION

Stream	Period of measured flow	Base station for extension	Period of flow record at base station	Number of years
Blackmare Creek	1990–1994	SF Salmon River at Krassel	1967–1982, 1986, 1990–1995	23
Canyon Creek	1992–1995	Selway River at Lowell	1911–1912, 1930–1995	68
Catspur Creek	1987–1995	NF Clearwater	1968–1995	28
Dollar Creek	1990–1994	SF Salmon River at Krassel	1967–1982, 1986, 1990–1995	23
Eggers Creek	1966–1993	Not extended	1966–1993	28
Fish Creek	1992–1995	Selway River at Lowell	1911–1912, 1930–1995	68
Fourth of July Creek	1994–1995	Valley Creek at Stanley	1921–1972, 1992–1995	56
Hawley Creek	1989–1995	Squaw Creek near Clayton	1971–1995	25
Herd Creek	1994–1995	Valley Creek at Stanley	1921–1972, 1992–1995	56
Johns Creek	1986–1995	SF Clearwater at Stites	1965–1995	31
Little Buckhorn Creek	1990–1994	SF Salmon R.at Krassel	1967–1982, 1986, 1990–1995	23
Little Slate Creek	1986–1995	Little Salmon River	1952–1954, 1957–1995	42
Lolo Creek	1986–1995	Lolo Creek near Greer	1980–1995	16
Marsh Creek	1994–1995	Valley Creek at Stanley	1921–1972, 1992–1995	56
Rapid River	1986–1995	Salmon River at White Bird	1911–1995	85
Red River	1986–1995	SF Clearwater at Stites	1965–1995	31
SF Red River	1986–1995	SF Clearwater at Stites	1965–1995	31
Squaw Creek	1989–1995	Squaw Creek near Clayton	1971–1995	25
Squaw Creek	1971–1995	Not extended	1971–1995	25
Thompson Creek	1973–1995	Not extended	1973–1995	23
Trapper Creek	1986–1995	SF Clearwater at Stites	1965–1995	31
Valley Creek	1921–1972, 1992–1995	Not extended	1921–1972, 1992–1995	56
WF Buckhorn Creek	1990–1994	SF Salmon River at Krassel	1967–1982, 1986, 1990–1995	23

Notes: WF—West Fork; SF—South Fork; NF—National Forest.

Figure 5 shows the hydraulic geometry as a function of bankfull discharge. Emmett (1975) found generally similar relations in his study of tributaries of the Salmon River in central Idaho, and when the analysis is limited to the same range of bankfull discharge used by Emmett (1975), the equations are more similar.

The median size of surface particles at sites increases with stream power (Ω) at bankfull discharge (Fig. 6): $D_{50} = 1.130\ \Omega^{0.601}$, $r^2 = 0.819$. Stream power is the product of stream gradient, discharge, and the unit weight of water (Bagnold, 1977).

Hydrology of Sites

These upland channels reach peak flows in April, May, or June in association with spring snowmelt. Some of the northern basins, particularly those at lower elevations, have occasional high flows during fall and winter associated with cyclonic storms and with rain on snow events. In most basins, fall and winter are times of low flow. Figure 2 shows the regular pattern of spring snowmelt and fall and winter low flow at the Red River.

The field surveys and surface and subsurface sediment analyses were conducted in 1994 and 1995. Water year 1994 (October 1, 1993–September 30, 1994) was unusually dry (USGS, 1994). Water year 1995 was far wetter, particularly in southern and central Idaho (USGS, 1995). The peak discharges in 1994 at sites correspond on average to 1.5 yr recurrence interval flows, while in 1995 sites correspond on average to 4.4 yr recurrence interval flows. The period from the mid-1980s to the mid-1990s included several years of drought (Fig. 2).

Flow-Duration Curves

The cumulative frequency of the flow record[1] at sites is shown in Figure 7 (A and B). There is gross similarity in the structure of most curves that in part reflects the similarity of the hydrologic regime at sites—a snowmelt driven hydrograph. Hawley Creek, Lolo Creek, and Squaw Creek near Salmon have a relatively smaller range of flows compared to the other streams. This may reflect a very large proportion of ground-water input to the stream in these basins (B. Riefenberger, 1995, personal commun; B. Spiegal, 1997, personal commun.) (cf. Whiting and Stamm, 1995).

Extension of flow records increased the magnitude of the mean annual discharge (Table 3) over the short period values for most sites. This is because much of the record at many sites is from the mid-1980s to 1995, a period including several years of drought. The ratio of mean annual discharge to bankfull discharge averages 0.21 at the study sites. In his study of Salmon River tributaries, Emmett (1975) found that the ratio of mean annual discharge to bankfull discharge averaged about 0.25 and that the ratio rose from about 0.20 for small drainage areas to 0.27 for larger drainage areas.

Bankfull discharge is reached or exceeded on average of 9.2 days per year at sites. The range is from 0.4 to 24 days per year. The recurrence interval of the bankfull discharge at all sites averages 2.0 yr with a range of 1.1 to 4.8 yr (Table 3).

Sediment Rating Curves

The bed-load transport rates and associated flow discharges are shown in Figure 8 for about one-half of the streams. Figure 8 also shows the

[1]Mean daily discharge values at sites for the period of record are available from the first author.

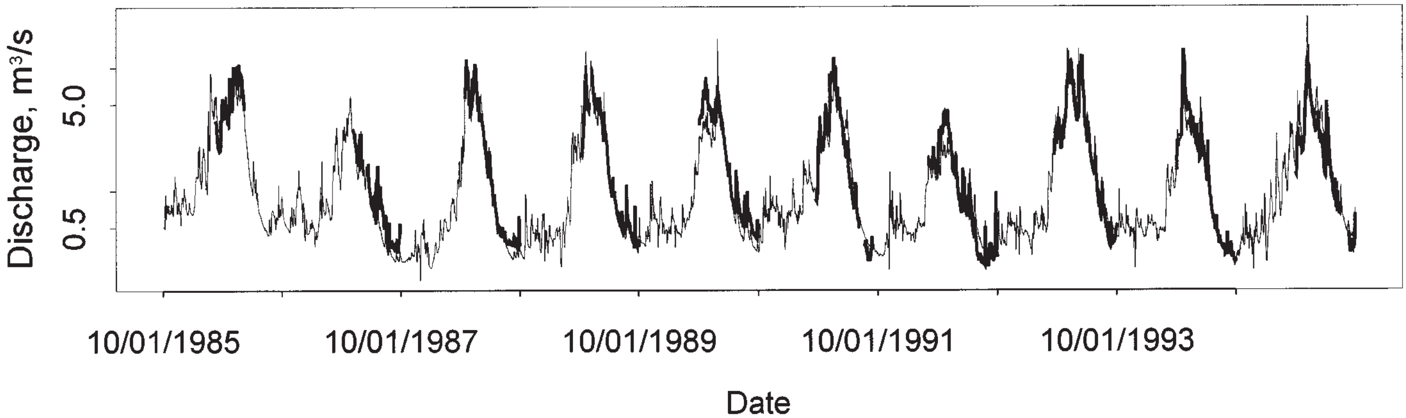


Figure 2. Daily streamflow at Red River for the period 1986–1995. The thick line represents measured streamflow values, and the fine line represents extended streamflow values.

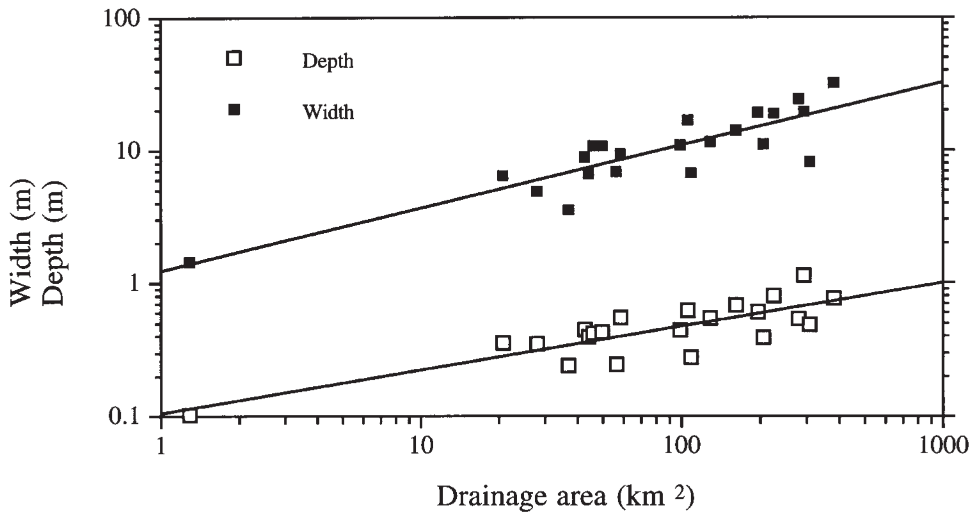


Figure 3. Bankfull width and depth increase with drainage area.

best-fit regression to the data of the form

$$Q_s = k\,Q^x, \tag{9}$$

where Q_s is the bed-load transport rate in metric tons per day, k is a coefficient, Q is the flow discharge in m^3s^{-1} and x is real. Models other than this traditional model were evaluated, but these did not provide consistent and physically reasonable results across the range of sites. Bed load was determined herein as material collected in the Helley-Smith sampler coarser than 0.85 mm, the calculated approximate upper limit to sizes moved in suspension at these sites under the studied flows using the criterion for suspension that the shear velocity must be greater than the particle fall velocity (Bagnold, 1966; and as discussed by Sumer et al., 1996). The sediment coarser than 0.85 mm represents on average 58% of the sediment collected in the Helley-Smith sampler (range = 45%–72%). Two bed-load measurements were typically made during each visit to the site. If stage readings were frequent, each bed-load measurement was associated with a distinct flow discharge, otherwise the two bed-load measurements were associated with a common discharge value. The number of bed-load measurements at sites[2] varies from 15 to 174 (Table 3). The mean of the values of the exponent x in equation 9 (Table 3) is 2.44 (standard deviation, S.D. = 0.61) and the median is 2.37.

At the few sites where there were 10 yr of bed-load data, the relationship between bed load and flow discharge was examined for hysteresis in bed load (Moog and Whiting, 1998). Annual clockwise hysteresis was demonstrated; prior to the first occurrence of a threshold discharge each year, there was greater transport at a given flow discharge than after the threshold. The threshold discharge is greater than the mean annual flow and varies from about one-third to one-half bankfull discharge. Apparently, stores of readily moved material were exhausted by the time rising flows reached the threshold discharge. The observed hysteresis is of secondary importance, compared to flow discharge, in explaining the bed-load transport rates. Because there were not data available at most sites to define the magnitude of hysteresis or threshold discharge at sites, subsequent calculations of the amount of sediment moved are based upon equation 9.

The average annual yield of bed load (Table 3) finer than 76 mm averages 3.58 t km^{-2} y^{-1}. There is large scatter about the relation that we attribute in part to different basin characteristics (geology, slope, precipitation, and land use) and to the size of the Helley-Smith sampler relative to the bed material size. Less transport is estimated on an annual basis in the streams with larger particles that those with smaller bed sediment.

Average annual yield of suspended load was

[2] Bed load transport values and the associated flow discharge are available from the first author.

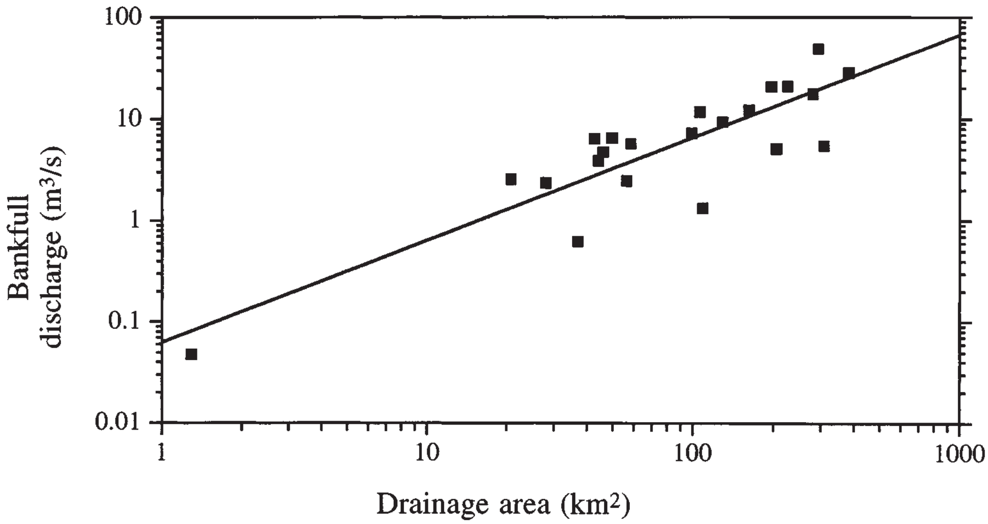


Figure 4. Bankfull discharge increases with drainage area.

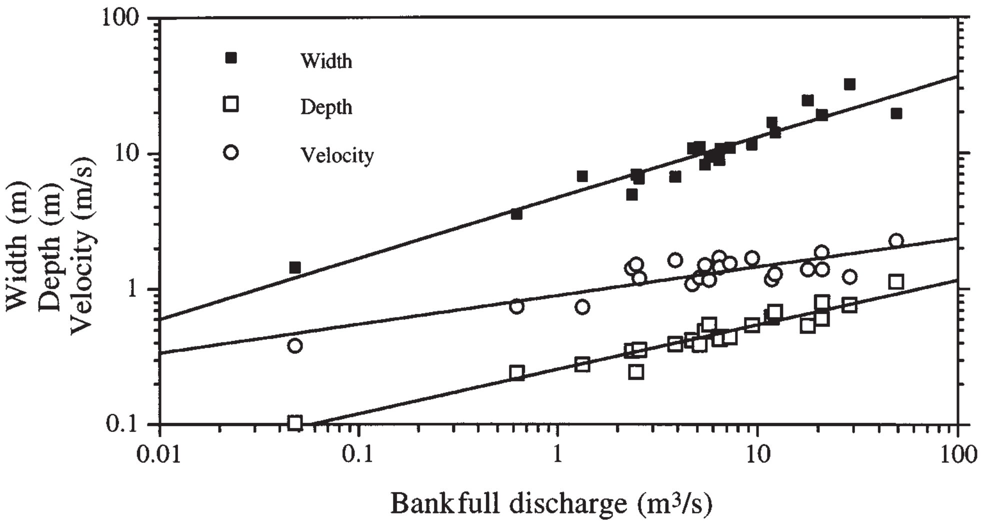


Figure 5. Hydraulic geometry at sites; bankfull depth, width, and velocity as a function of bankfull discharge.

calculated at six sites—Fourth of July Creek, Herd Creek, Marsh Creek, Squaw Creek, Thompson Creek, and Valley Creek. At these sites, bed load is a relatively small proportion of the annual suspended load (7%–33%), although this should be a minimum value since bed load coarser than 76 mm would not be collected due to the sampler orifice. However, these sizes appear to be moving, from various lines of evidence.

Bed Load, Bed Surface, and Subsurface Size Distributions

The size of the coarsest grain collected in each bed-load sampler is plotted in Figure 9 as a function of flow discharge. Only data from six sites are shown because the characterization of the particle sizes was the most detailed for these sites. Although there is substantial scatter in the relationship, that can be expected given the probabilistic nature of sediment entrainment and transport (Einstein, 1948; Dietrich and Whiting, 1989), there are consistent trends of increasing particle size with increasing flows in the data from sites. The increase in particle size with discharge is nearly linear (exponents of 0.93 to 1.32). Results at other sites are substantively similar. Despite the clear increase in particle sizes sampled with increasing discharge, most of the bed load (40%–90%) is sand size or finer (Table 4).

The bed load is finer than the material making up the bed surface and subsurface (Table 4). The bed-load size reported is the transport- and frequency-weighted median grain diameter. Because the orifice of the Helley-Smith sampler (76 mm) limits the size of collected grains, the size distributions were truncated to include only those sizes less than 32 mm, and the median diameter of the truncated samples was recalculated. The bed load is finer than the subsurface material, as indicated by the value of D*—the ratio of the median diameter (D_{50}) of bed load and subsurface material (Lisle, 1995).

These stream beds are armored. The ratio of the size of median surface particle to the size of the median subsurface particle averages 3.3 for eight of the sites where both measurements were made. The range of values is from 1.8 to 5.0.

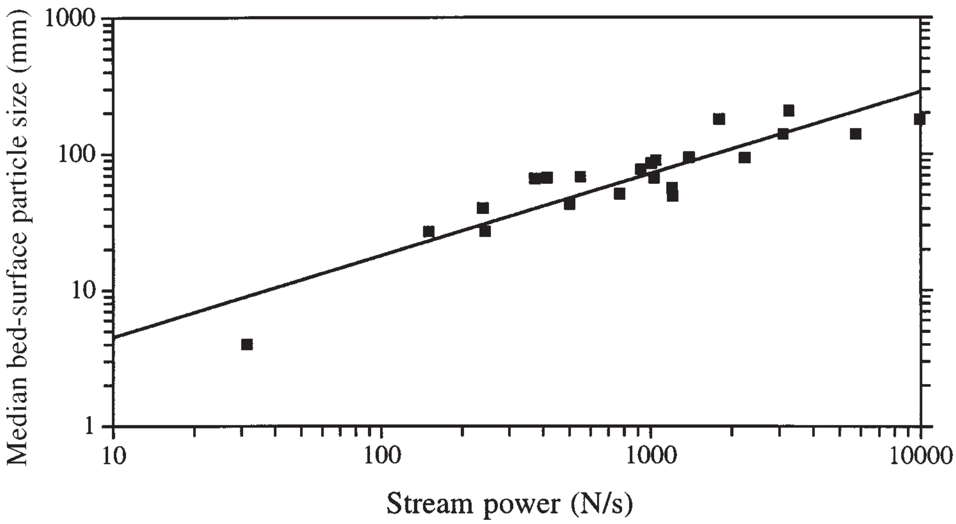


Figure 6. Median bed-surface particle size increases as stream power increases.

Painted Rock Experiments

The orifice of the Helley-Smith bed load sampler limits the size of bed-load that can be measured during sampling. The probabilistic nature of sediment transport (Einstein, 1948) also means that the movement of the larger particles is infrequent, and the likelihood that a sampler will be in position to catch such particles (even if they could fit into the sampler) is very low (Dietrich and Whiting, 1989). One method for estimating the range of sizes that are in transport is to place painted rocks in the stream (e.g., Leopold, 1994; Leopold and Emmett, 1981). A common criticism of such studies is that rocks deployed on the stream bed are inherently more mobile because of their placement, i.e., loose and atop the bed. Particles were placed in stable sites; in pockets of the bed with their short axis pointing upward.

Painted rocks placed at several sites show that particles up to at least the D_{90} of the bed material were capable of being moved by the bankfull and higher flows of 1995 (Fig. 10). Recovery rates of particles ranged from 20% to 100%. Some painted rocks were observed to take multiple steps downstream with periods of repose lasting days or weeks. The multiple steps taken by these grains indicate that they found hydraulically stable pockets for deposition but then moved again, suggesting that initial placement does not explain the mobility of these sizes. Andrews (1984) documented that a fraction of deployed painted rocks moved several times during a snowmelt season.

SEDIMENT TRANSPORT REGIMEN

The amount and size of bed load increases with the streamflow in these gravel-bed streams. The exponents to the sediment rating equation range from 1.62 to 3.93; the median value is 2.37. Using the median value of 2.37, a doubling of the streamflow is associated with a more than five-fold increase in bed-load discharge. Leopold (1994) also found that the slope of the rating curve is most commonly between 2 and 3. The size of particles in motion also increases with the streamflow. As can be seen in Figure 9, the sizes of the coarsest gravel particles captured in the sampler increase nearly linearly with flow discharge. Painted rock experiments suggest that the grains making up the bed surface (grains up to at least D_{90}) are moved by bankfull flows (Fig. 10). The painted rock findings are supported by the size of the coarsest particles showing evidence of recent movement found on bars. Despite the observation that coarse particles are moving, most of the bed load in these gravel-bed channels is sand, as has been found by others (e.g., Leopold, 1992). Sand makes up the majority of bed load at all measured discharges. The annual bed-load yield is at a minimum 7%–33% of suspended-load yield in these channels. For comparison, Emmett (1984) reported that bed load in large mainstem rivers is about 5% of the suspended load.

The bed load is finer than the subsurface particle sizes in streams for which subsurface data were available (Table 4). Lisle (1995) found that the ratio of the median particle size of bed material to the transport- and frequency-weighted mean of median bed-load size (D*) decreased from approximately 3 to approximately 1 at a drainage area of 100 km^2 or at a bankfull discharge of 50 m^3/s. Lisle (1995) used this result to suggest that the equal mobility concept (Parker and Klingeman, 1982; Andrews, 1983), while it may be applicable to larger streams, breaks down in smaller basins. Values of D* at sites are large (2–7), which is consistent with selective transport of finer sizes (cf. Ashworth and Ferguson, 1989). The observation that the size of the coarsest particle in the Helley-Smith sampler increases with flow discharge is consistent with selective transport of the finer fraction (Fig. 9). In contrast to the findings of Lisle (1995), there is no indication that selective transport becomes less important in basins larger than 100 km^2. For all sites, selective transport of finer sediment is observed rather than equal mobility.

The channel beds studied are armored; the ratio of surface to subsurface D^{50} averages 3.37 for the sites. Armoring has been linked (cf. Dietrich et al., 1989) to sediment supply that is less than the ability of the stream to transport that load. Supply limitations can give rise to armoring over small areas on the bed or over the entire bed. Local armor can be due to flow and shear-stress patterns, whereas general armor of the stream bed is due to watershed processes limiting the availability of sediment delivered to the stream. Observed clockwise hysteresis in bed load at discharges above a threshold corresponding to about one-third to one-half bankfull, as described earlier, is consistent with supply limitation inferred from the armoring (Dietrich et al., 1989). The ubiquity of surface armoring found at sites as well as the consistency of clockwise hysteresis suggest that the supply limitations reflect watershed processes rather than local in-channel processes.

Cumulative Curves Relating Bed Load to Streamflow

The flows responsible for moving bed load can be described by dimensionless cumulative curves of bed load and discharge at each site, as shown in Figure 11. Slopes of these curves progressively steepen to the right (at higher discharges). The fig-

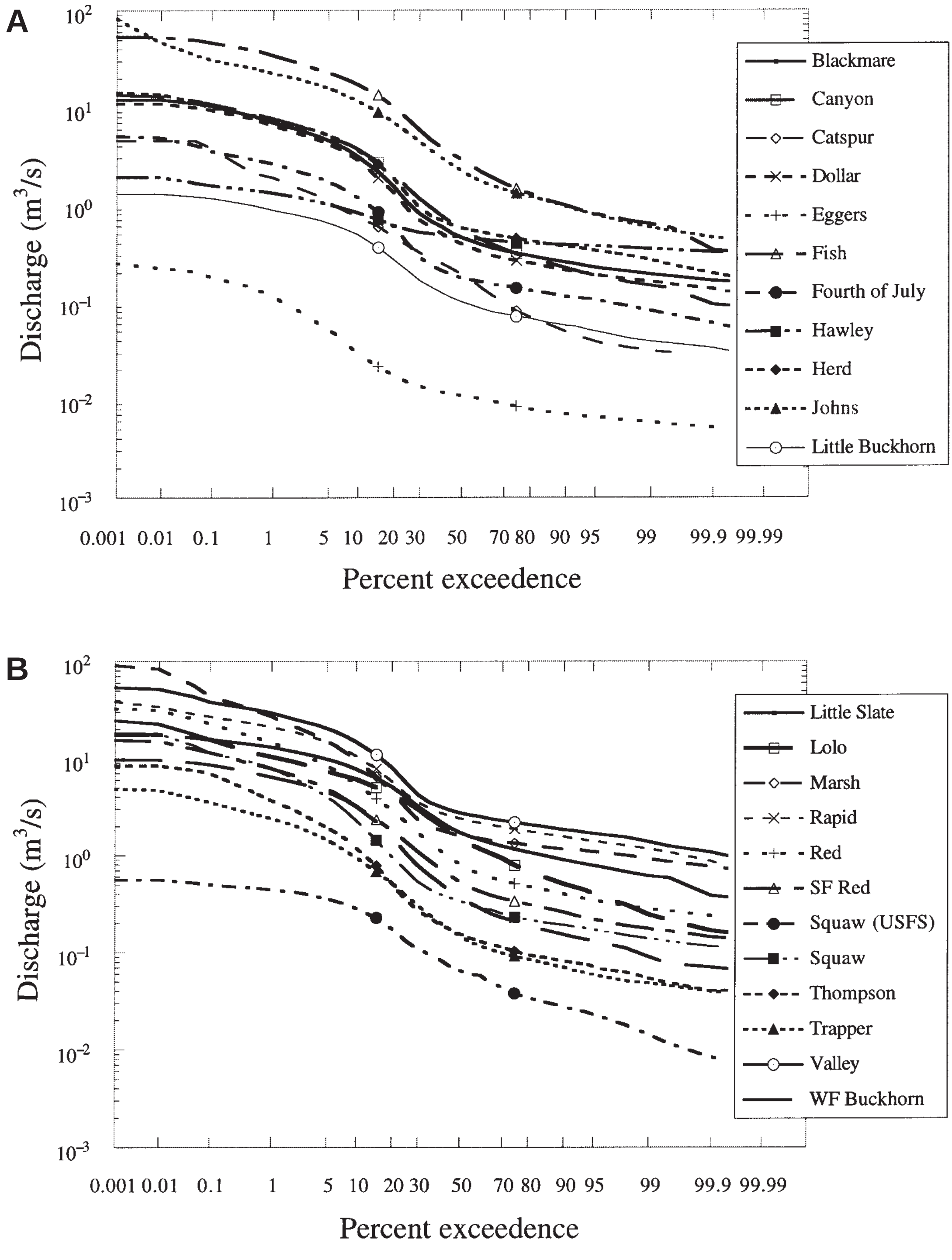


Figure 7. Probability of exceedence for flows at sites—presented alphabetically in A and B. The similarity of most curves is due to the common snowmelt-dominated hydrograph at sites. The exceptions are Hawley, Lolo, and Squaw Creeks, in which there are large spring-fed components.

ure clearly shows that the higher flows move a disproportionately large amount of the sediment. Almost one-half of the water (43%), that water associated with the lower discharges, moves only the first 10% of the bed load. In contrast, the upper 1.9% of the flow moves the upper 10% of the bed load. One-half of the bed load is moved by the highest 18% of the flow. Webb and Walling (1982) found that 50% of suspended load was moved by 9% of the total discharge. The results clearly show that the highest flows are most efficient in moving sediment. At an individual stream, the greater the exponent to the bed-load rating curve, the smaller the proportion of water used to move the highest 10% of bed load (Fig. 12).

An average of 6% of bed load is moved by flows below mean annual discharge. An average of 57% of bed load is moved between mean annual discharge and bankfull discharge, and 37% of the bed load is moved by flows above bankfull discharge. Andrews and Nankervis (1995), in their study of gravel-bed streams, reported that bed material (> 4 mm) began to move at approximately 60% of the bankfull discharge and calculated that discharges less than the bankfull transported 39% of the annual bed load. However, we noted in Idaho that about 30% of bed load is moved by

TABLE 3. HYDROLOGY AND BED-LOAD TRANSPORT AT SITES

Stream	Mean annual flow	Bed load samples	x (Eq. 9)	Annual bed load yield	Bankfull		Effective	
					Discharge	Rec. Int.	Discharge	Rec. Int.
	(m^3/s)			(t /km^2 /yr)	(m^3/s)	(yr)	(m^3/s)	(yr)
Blackmare Creek	1.11	94	2.27	4.45	4.73	1.1	5.24	1.2
Canyon Creek	1.31	17	2.57	1.57	6.48	3.7	4.93	1.3
Catspur Creek	0.334	34	2.70	2.43	2.36	1.4	1.70	1.1
Dollar Creek	0.988	87	2.37	3.47	6.43	1.1	4.64	1.1
Eggers Creek	0.0168	137	1.63	7.65	0.0478	1.2	0.0170	1.04
Fish Creek	6.20	15	3.42	0.61	20.9	1.1	23.3	1.2
Fourth of July Creek	0.425	76	3.06	1.04	3.88	1.5	2.01	1.1
Hawley Creek	0.535	63	2.79	1.16	1.33	2.1	0.470	1.00
Herd Creek	1.31	69	2.74	3.50	5.46	1.4	5.89	1.5
Johns Creek	4.47	115	2.24	0.21	49.0	4.8	16.5	1.1
Little Buckhorn Creek	0.188	78	2.30	3.14	N.A.	N.A.	0.708	2.8
Little Slate Creek	3.11	114	1.62	0.42	12.2	2.4	7.50	1.5
Lolo Creek	2.51	109	1.76	1.33	11.7	2.3	6.66	1.05
Marsh Creek	3.79	98	2.55	7.22	20.8	1.6	18.3	1.5
Rapid River	4.30	128	2.16	2.13	17.7	2.3	15.9	1.9
Red River	2.07	167	2.41	7.86	9.34	1.3	10.3	1.3
SF Red River	1.24	174	1.97	3.22	7.25	1.9	5.75	1.3
Squaw Creek	0.110	31	1.75	0.32	0.623	4.2	0.365	1.5
Squaw Creek	0.903	91	2.77	1.69	5.12	1.6	8.04	2.6
Thompson Creek	0.447	77	3.34	4.43	2.48	1.6	3.40	2.3
Trapper Creek	0.360	153	1.85	5.93	2.56	2.6	1.06	1.1
Valley Creek	5.63	88	3.93	15.66	28.6	2.0	23.8	1.5
WF Buckhorn Creek	1.03	87	1.93	3.00	5.72	1.2	4.98	1.00

Note: WF— West Fork; N.A.—Bankfull could not be identified at this site.

flows below 60% of bankfull and that 60% of bed load is moved by discharges below bankfull. Given that Andrews and Nankervis (1995) may have excluded the finer bed load in their analysis, our results may be substantially similar.

Bed-Load Effective Discharge

Effective discharge is that flow which, over the long term, transports more sediment than any other flow. Effective discharge considers both the magnitude and frequency of events transporting sediment. The estimate of effective discharge follows the general approach of Wolman and Miller (1960) and Andrews (1980). Unlike most other researchers, we have estimated the effective discharge based upon measured bed-load sediment rating curves, as opposed to bed load calculated from bed-load transport equations. Furthermore, calculation of the effective discharge is not based upon binning of flows into regular discharge increments (i.e., Andrews, 1980), but rather is based upon the product of the sediment-rating curve (magnitude) and discharge probability density function (frequency). Probability density functions (pdf) were evaluated at 100 points covering the discharge domain using the S-PLUS software function “density” with kernel estimates based on Gaussian windows. That is, for each of these 100 points, a function (the window) is centered on the point and the heights of the window at each data point are summed. This sum, after normalization, is the value of the pdf at that discharge. The window is a Gaussian curve with a standard error equal to 0.025 times the width of the discharge domain. This was found by experience to smooth out likely idiosyncrasies in the pdf and to retain the more fundamental features.

Figure 13 shows effective discharge curves for selected sites. Many of the plots show the classic pattern of a single modal effective transport rate (“Wolman hump”), but several of the plots display multiple modes. As suggested by Nash (1994), these multiple modes may be caused by records that are insufficiently long to average out infrequent events. Alternatively, these modes could be due to inherent structure in the frequency of hydrologic events. In other cases shown in Figure 13, a single modal value is strongly skewed to relatively small discharges; e.g., Hawley Creek and Eggers Creek. This may be tied to reduced variability in flow spring-fed systems such as Hawley Creek. Nash (1994) mentioned that poor estimation of the transport function at the highest flows can alter the classic form of effective discharge curves. Extrapolation of the bed-load function to higher measured discharges could influence the resulting effective discharge curve. An average of 11.3% of the total bed-load transport is associated with extrapolation above the highest flows for which there were bed-load measurements. We do not believe that either poor estimation of the transport function at high discharges or extrapolation are important factors in the results for these sites.

The effective discharge increases as the bankfull discharge increases (Fig. 14). Effective discharge at all the sites averages 80% of the bankfull discharge and ranges from 34% to 157% at sites. Several of the lowest effective discharges, relative to bankfull discharge, are at sites with comparatively small ranges in discharge, whereas some of the largest effective discharges are at sites with large ranges in discharge; the significance of these observations will be discussed later.

The recurrence interval of effective discharge averages 1.4 yr (Table 3), whereas the recurrence interval for bankfull discharge is 2.0 yr (Table 3). The median percent of time that the effective discharge is equaled or exceeded is 3.8 (range = 1.0%–47%; average = 6.5%) or 14 days per year. The median value is more consistent than the 0.35% to 6.51% reported by others studying gravel-bed streams (Webb and Walling, 1982; Andrews and Nankervis, 1995).

Others have reported similarities between bankfull and effective discharges (Andrews, 1980; Biedernharn et al., 1987; Leopold, 1992; Nolan et al., 1987; Pickup and Warner, 1976; Webb and Walling, 1982; Wolman and Miller, 1960), although all these reports except those of Leopold (1992) and Andrews and Nankervis (1995) were for total or suspended load. Ashmore and Day (1988) found that the effective discharge for suspended sediment was a rarer and more extreme event in smaller drainage basins. Pitlick (1988) found that bed-load effective discharge in northern California became an increasingly rare event as the gradient and bed-material size increased. For the channels examined in Idaho,

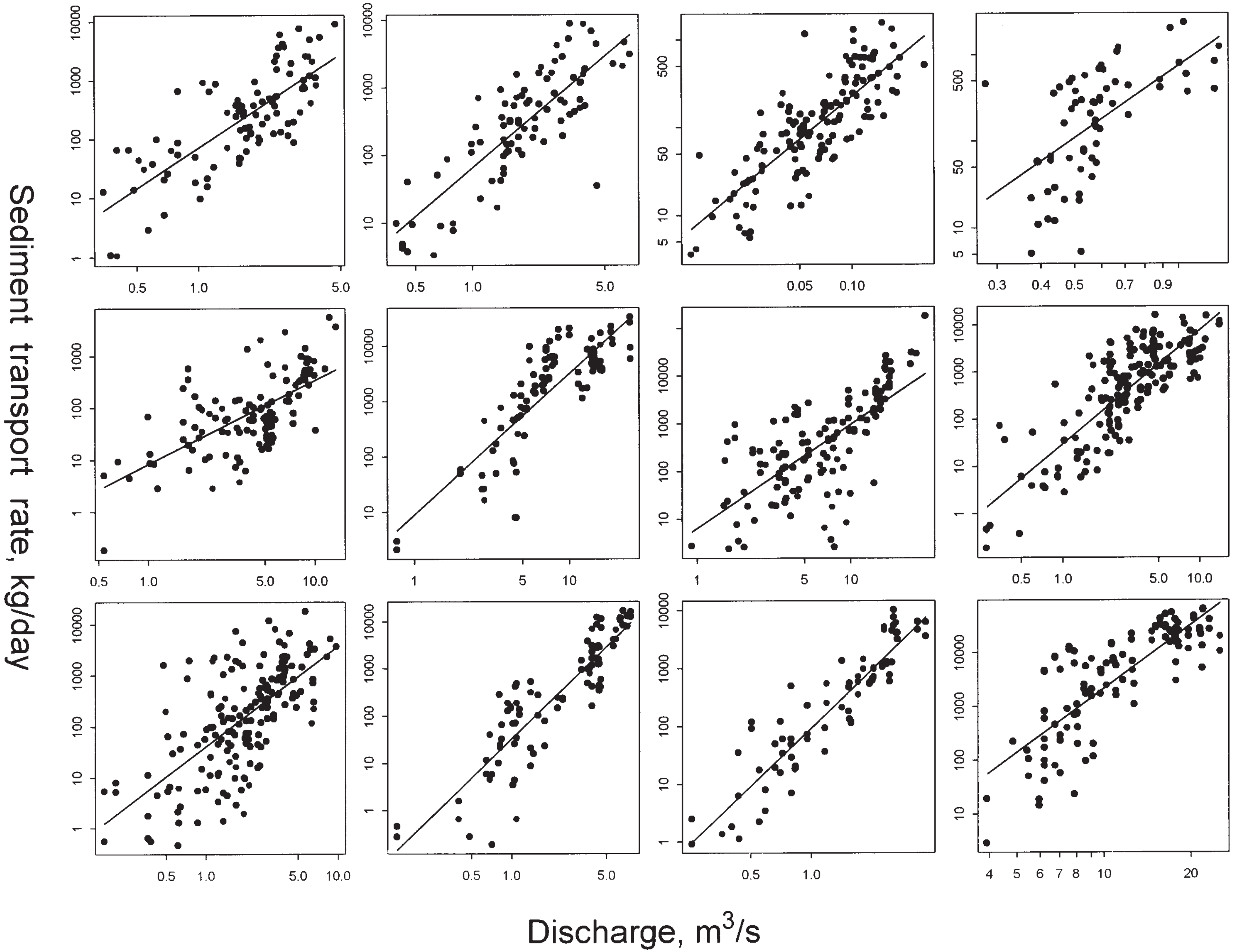


Figure 8. Bed-load sediment rating curves for select sites. From the upper left to the lower right these are: Blackmare Creek, Dollar Creek, Eggers Creek, Hawley Creek, Little Slate Creek, Marsh Creek, Rapid River, Red River, South Fork Red River, Squaw Creek, Thompson Creek, and Valley Creek. Exponents to the rating curves are listed in Table 3.

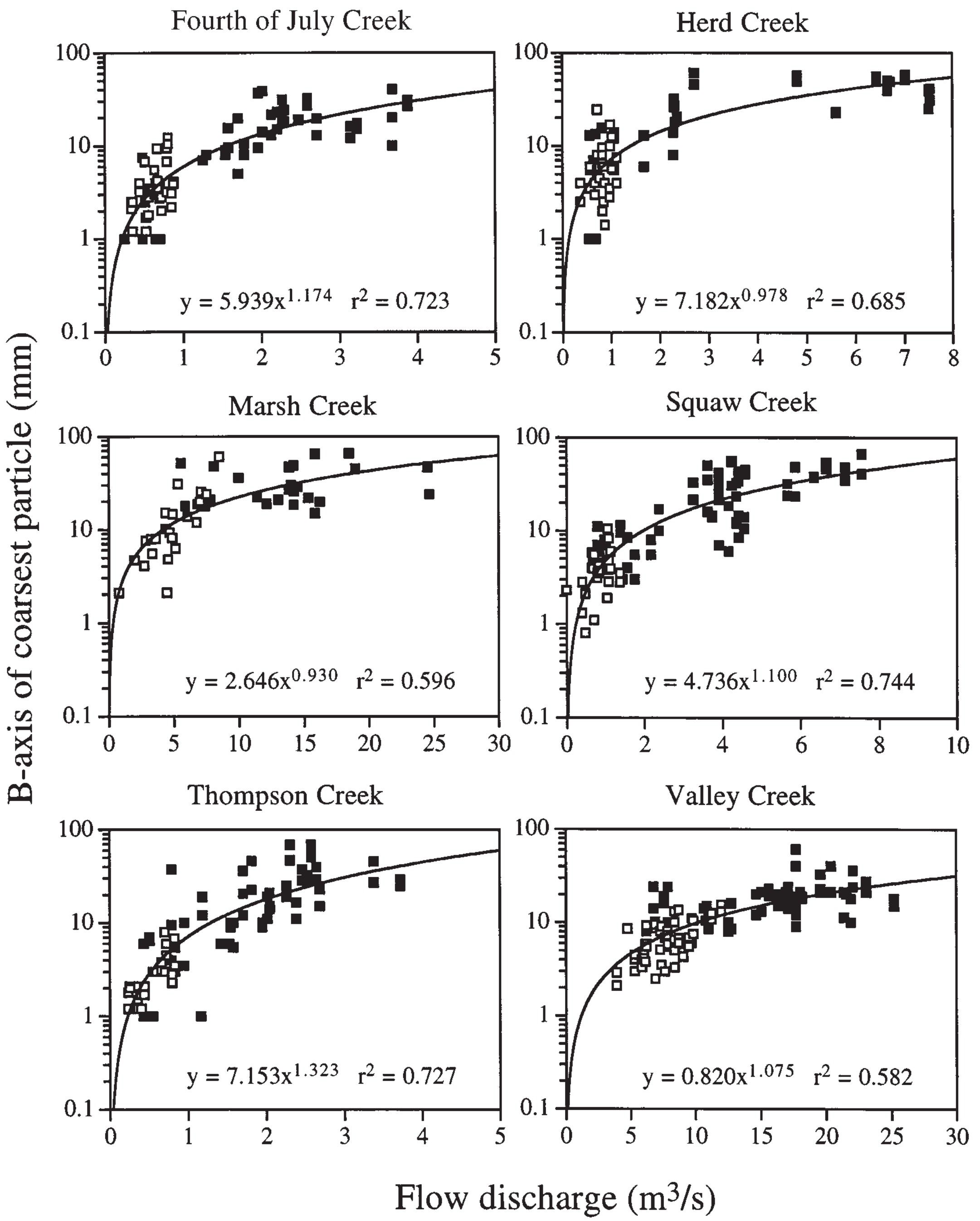


Figure 9. The size of the coarsest particle caught in the Helley-Smith sampler increases as the discharge increases. Open squares represent data collected in 1994: closed squares represent data collected in 1995. y = B-axis; X = flow discharge.

there is no evidence that effective discharge becomes a rare event, in smaller drainage basins or in streams (Fig. 15A), with larger particles (Fig. 15B) or steeper gradients (Fig. 15C).

Wolman and Miller (1960), Andrews (1980), and Nash (1994) suggested that the magnitude or recurrence interval of effective discharge increases with discharge variability. Our findings are similar; we observe that the relative magnitude of effective discharge increases as the relative magnitude of rare events (large events) increases (Fig. 16). Here, we describe variability in terms of the discharge associated with an event with an exceedence probability of 0.1% as scaled by the median discharge. Since the effective discharge is the product of magnitude and frequency, streams in which discharge is relatively larger for a given frequency of event will have larger effective discharges, all else being equal. Spring-fed streams such as Hawley Creek and Squaw Creek in the Salmon National Forest with relatively small discharge ranges (Fig. 7, A and B) have relatively small effective discharges (respectively, 35% and 59% of bankfull discharge). However, flashy streams such as Thompson Creek and Squaw Creek have relatively large effective discharges (157% and 137% of bankfull discharge).

An inverse relationship exists between drainage area and the ratio of the magnitude of a given event to the mean annual discharge (Kuiper, 1957). Insofar as smaller basins tend to be more prone to flash floods, and more-extreme, less-frequent events are thought to play a more important role in smaller basins (Costa, 1974; Grant et al., 1990), one would expect effective discharge to increase in relative magnitude and to become a less frequent

TABLE 4. SURFACE, SUBSURFACE, AND BED LOAD SIZES

Stream	Bedsurface D_{50} (mm)	Subsurface D_{50} (mm)	D_{50} of load (mm)	Sand in bed load† %	Truncated subsurface D_{50}§ (mm)	D*
Johns Creek	207	56	5.7	85	N.A.	N.A.
Little Slate Creek	102	31	1.4	90	9.2	6.6
Lolo Creek	68	20	1.4	83	10.0	7.1
Rapid River	92	31	1.9	63	9.1	4.8
Red River	39	20	2.4	80	6.9	2.9
SF Red River	107	26	4.0	80	8.0	2.0
Trapper Creek	85	17	3.1	75	8.7	2.8
Valley Creek	53	29	3.0	59	8.9	3.0
Fourth of July Creek	51	N.A.	N.A.	74	N.A.	N.A.
Herd Creek	67	N.A.	N.A.	56	N.A.	N.A.
Marsh Creek	56	N.A.	N.A.	40	N.A.	N.A.
Squaw Creek	43	N.A.	N.A.	73	N.A.	N.A.
Thompson Creek	66	N.A.	N.A.	87	N.A.	N.A.

Note: SF—South Fork; N.A.—Bankfull could not be identified at this site; D_{50}—median diameter.
*Surface size at site of subsurface sample if one taken.
†Average of samples in 1994 and 1995.
§Subsurface grain-size distribution truncated to sizes to less than 32 mm.

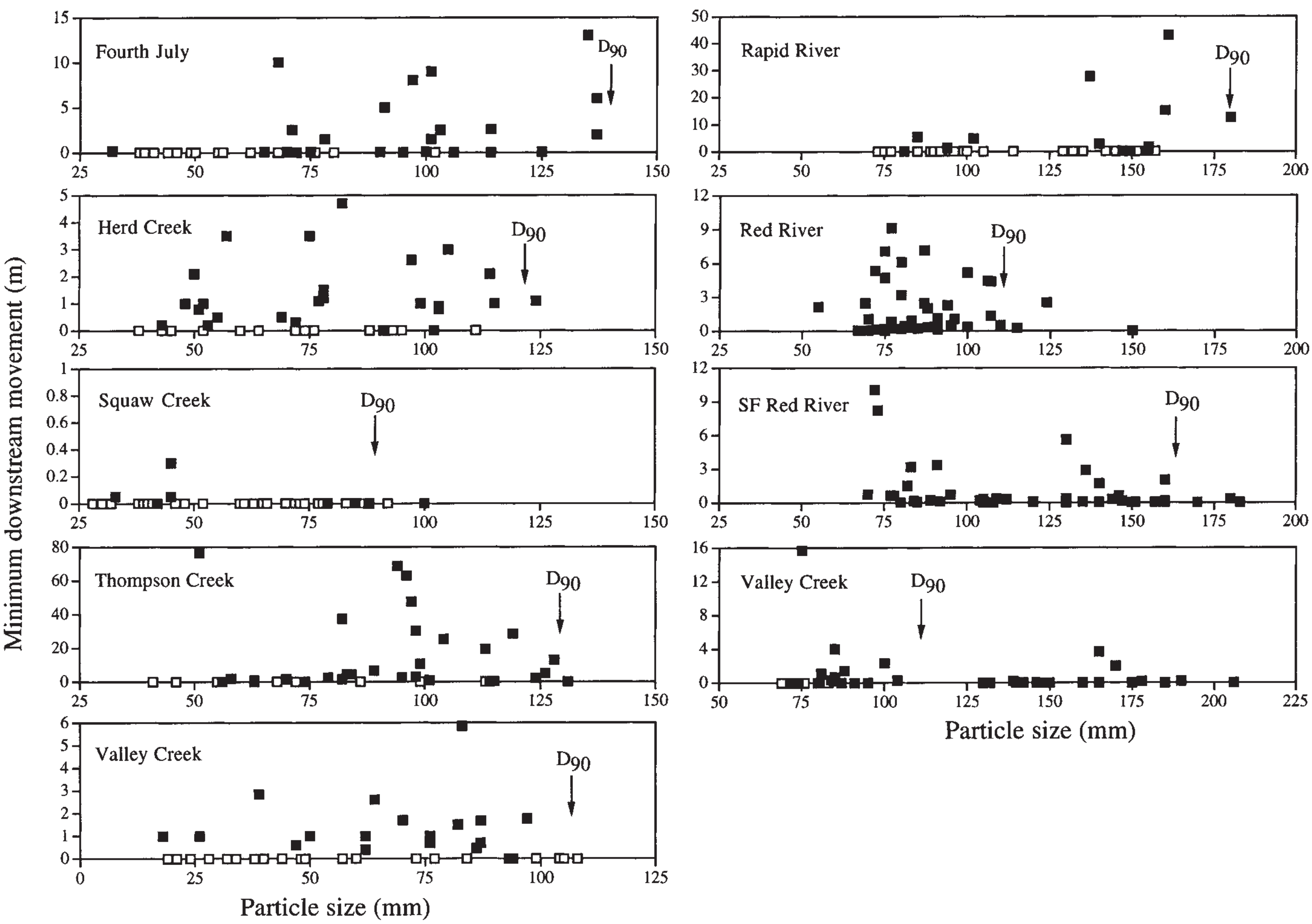


Figure 10. Downstream displacement of painted rocks placed in two 20-rock transects per site at sites. Recovered particles are shown as filled squares; unrecovered particles are shown as open squares.

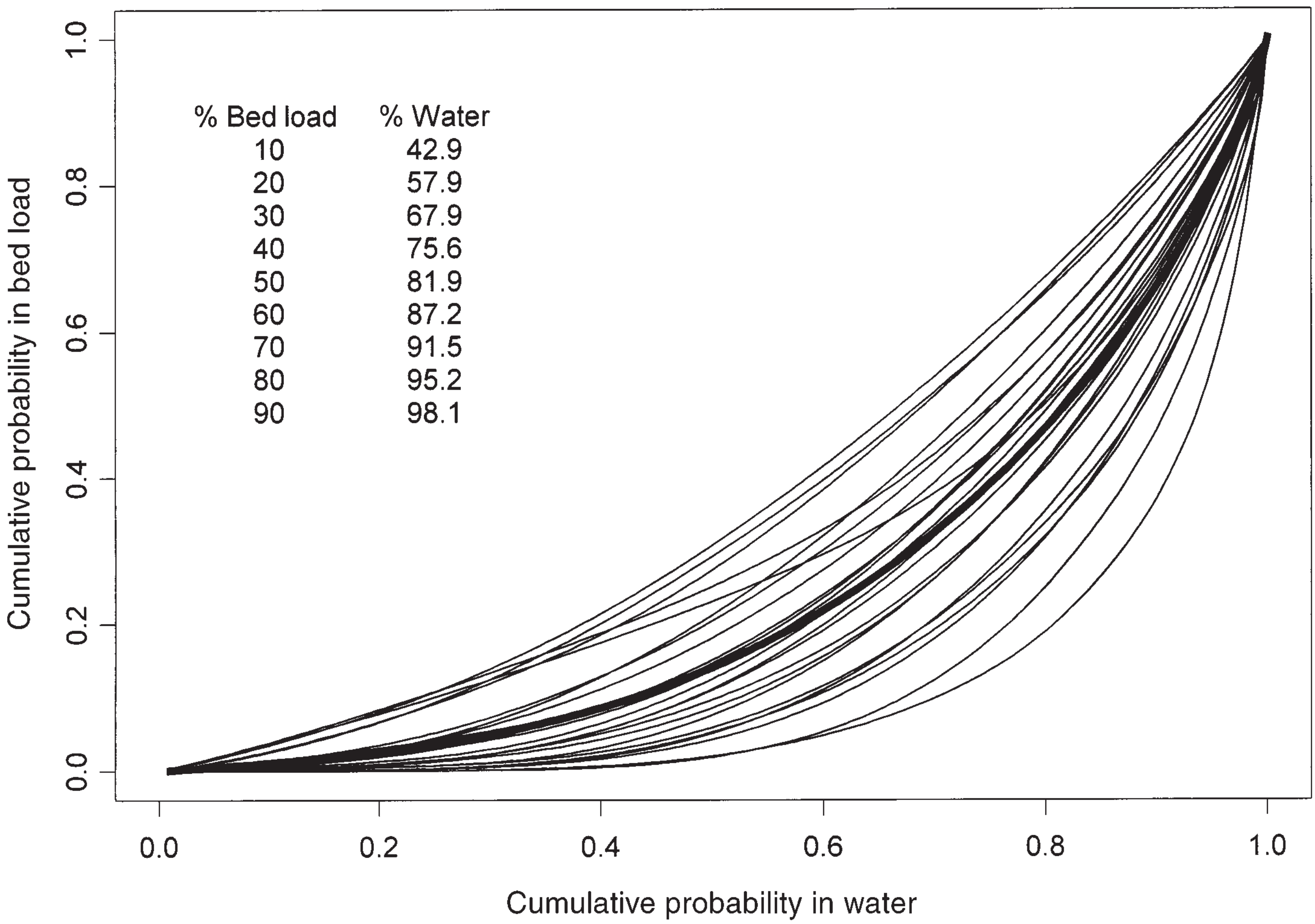


Figure 11. Cumulative curves for the percent of water used to move a given percent of bed load at sites (fine line). The average of all values at sites is the bold line. The amount of water associated with the decadal cumulative percentages of bed load is shown on the inset table.

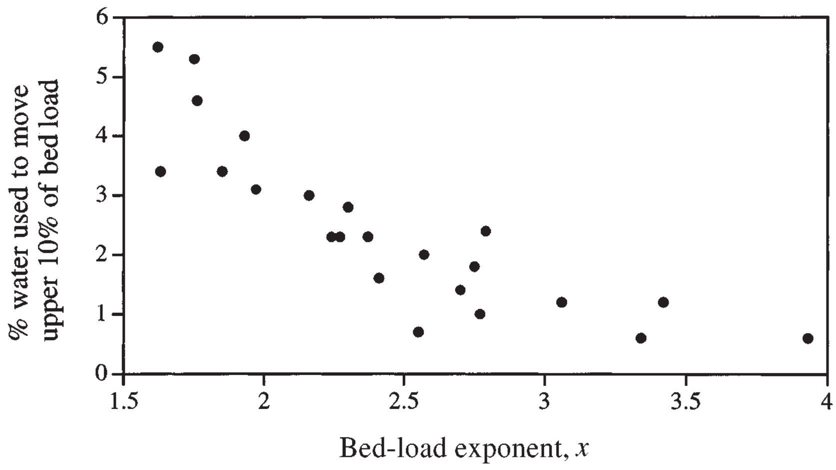


Figure 12. The greater the exponent to the bed-load rating curve, *x*, the smaller the percentage of water used to move the upper 10% of the bed load.

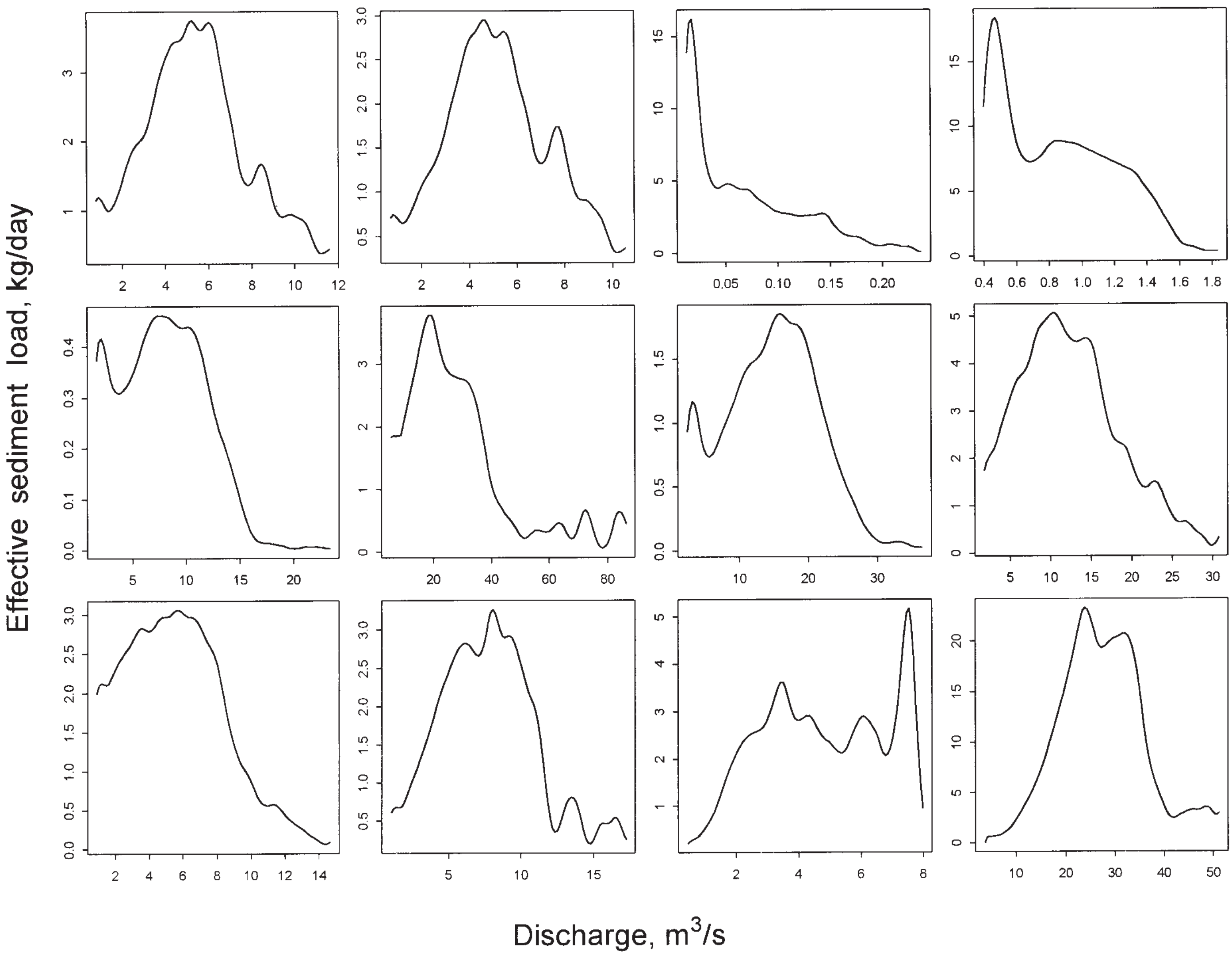
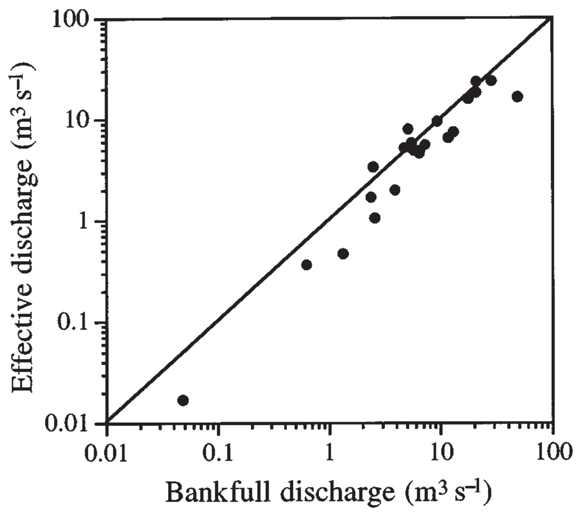


Figure 13. Effective discharge curves for selected sites. Reading from left to right across the first row and ending at the lower right these are: Blackmare Creek, Dollar Creek, Eggers Creek, Hawley Creek, Little Slate Creek, Marsh Creek, Rapid River, Red River, SF Red River, Squaw Creek (U.S.G.S.), Thompson Creek, and Valley Creek.

Figure 14. Bankfull and effective discharge at sites. The line indicates a 1:1 relationship. Effective discharge averages 80% of bankfull discharge at sites.



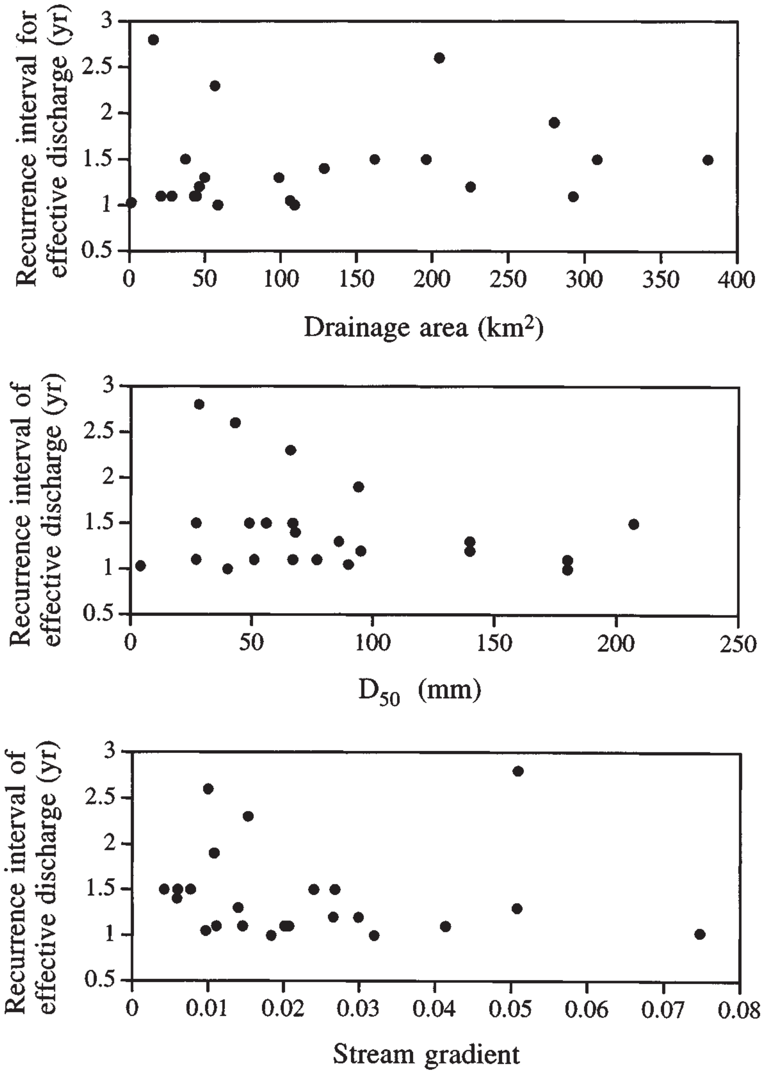


Figure 15. There is no obvious trend in effective discharge with drainage area (A), median particle size (D_{50})(B), or stream gradient (C).

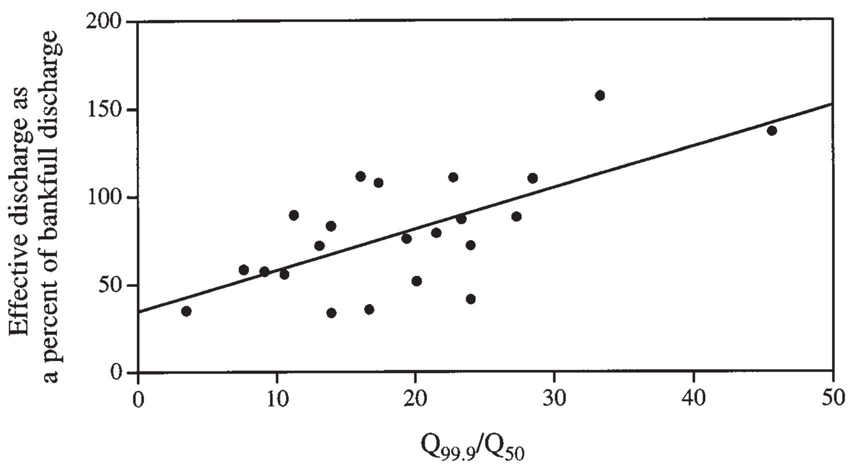


Figure 16. The relative magnitude of effective discharge increases as the relative magnitude of rare events increases: y = 2.35x + 35.23, r^2 = 0.48.

event in smaller basins. It is somewhat surprising that there is no support for this inference in the data from these gravel-bed headwater streams (Fig. 15A). Local differences in hydrology may swamp the expected trend in effective discharge associated with basin size.

Bed-load effective discharge and bankfull discharge are similar in magnitude and are both moderately frequent events. This has been taken (e.g., Andrews, 1980) to indicate that channels are adjusted to the effective discharge. The relationships between channel dimensions and drainage area (Fig. 3), bankfull discharge and drainage area (Fig. 4), stream power and surface-particle size (Fig. 6) and hydraulic geometry (Fig. 5), further link those streamflows moving the most sediment to the creation and maintenance of the channel dimensions. We emphasize that effective discharge is a conceptually powerful tool for understanding the flows moving sediment, but that effective discharge is a surrogate for describing all flows that, by moving sediment making up the bed and bank, create and maintain the geometry of the channel.

CONCLUSIONS

The 23 alluvial headwater streams studied in central and northern Idaho span a large range of channel gradient, grain size, and drainage area. To the degree that the studied streams cover some of the variation seen in headwater streams, results and conclusions drawn from such streams should be generalizable to other alluvial headwater streams with similar climates and hydrology.

The study of the geometry and transport regime of headwater channels as well as the link between them points to the following conclusions.

1. Bankfull discharge is a moderately frequent event having an average recurrence interval of 2.0 yr. Bankfull discharge is reached on average nine days a year.

2. Progressively larger sediment is in motion as discharge rises but the majority of bed load is sand at all discharges. The bed load is finer than the bed material, suggesting that there is selective transport of finer sediment.

3. Hysteresis is seen in the bed load such that more transport is observed early in the hydrograph. This and the surface armor provide strong evidence for the supply limits in these channels.

4. Effective discharge averages 80% of bankfull discharge and has a recurrence interval of 1.4 yr. Effective discharge, as a percentage of bankfull discharge, is independent of basin size, grain size, and gradient, at least in these streams, although effective discharge as a percentage of bankfull discharge increases with the relative magnitude of large infrequent events. Streams with only moderate rises in

discharge during high flows have the smallest effective discharges.

5. About half of the water (that water associated with the lowest discharges) moves 10% of the bed load. In contrast, the upper 1.9% of the water moves 10% of the bed load.

6. Little bed load (~6%) is moved by flows below mean annual discharge, while 57% is moved between mean annual and bankfull discharge: 37% of the bed load is moved by flows above bankfull discharge.

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