

Critical flow constrains flow hydraulics in mobile-bed streams: A new hypothesis

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Abstract. A new hypothesis predicts that in mobile-bed river channels, interactions between the channel hydraulics and bed configuration prevent the Froude number (Fr) from exceeding 1 for more than short distances or periods of time. Flow conditions in many steep, competent streams appear to be close to critical. Froude numbers of steep (slope ≈ 0.01) sand-bed streams with considerable freedom to adjust boundaries oscillate between 0.7 and 1.3 over 20- to 30-s cycles, with an average of 1.0 at the channel thalweg. Critical flow in these streams is maintained by the interaction between the mobile bed and free water surface at high Fr , which results in a cyclical pattern of creation and destruction of bed forms. Field observations support that a similar mechanism of flow–bed form interaction constrains $Fr \leq 1$ in active-bed braided gravel rivers, step-pool streams, laboratory rills, lahar-runout channels, and even some bedrock channels. Empirical and analytical results show that as slope increases, competent flows tend to asymptotically approach critical flow. An assumption of critical flow would dramatically simplify paleohydraulic flow reconstructions and modeling of flow hydraulics in high gradient streams.

Introduction

The morphology of alluvial channels reflects a complex interaction among flow hydraulics, channel geometry, energy dissipation, and sediment transport. This relationship is at the heart of Mackin's [1948, p. 471] famous dictum that "a graded stream is one in which, over a period of years, slope is delicately adjusted to provide, with available discharge and prevailing channel characteristics, just the velocity required for the transportation of the load supplied by the drainage basin." Ever since Mackin, fluvial geomorphologists and engineers have been searching for the principles and physical mechanisms underlying channel adjustments. This work has taken on new urgency as direct streamflow and channel modifications, land use, and changing climate alter water and sediment transport regimes, sometimes changing the channel morphology and ecological functioning of streams. The many mutually dependent variables range over wide spatial and temporal scales in fluvial systems, however, and this variability has forestalled any comprehensive or deterministic predictions of how hydraulics, sediment transport, and channel morphology adjust to changes in driving variables [Schumm and Lichty, 1965; Slingerland, 1981].

These mutual adjustments are perhaps best understood in sand-bed streams, where both field and laboratory observations demonstrate the close interaction among the hydraulic free surface, channel bed forms, and sediment transport dynamics [Kennedy, 1963; Foley and Vanoni, 1977; Bean, 1977; Schumm et al., 1982]. For example, as flow velocity increases, bed forms in sand channels typically undergo a transition from plane bed to dunes to antidunes, with concurrent changes in the free surface from flat to ripples to standing waves. Much less understood is the relation among flow hydraulics, bed forms, and channel morphology in steeper, coarser-grained

channels, such as mountain streams, which commonly display a diverse array of bed forms (e.g., transverse ribs in gravel channels, step-pools in boulder-bed streams) and complex hydraulic phenomena (e.g., breaking standing waves, hydraulic jumps). Differences in the abundance of these hydraulic and bed form features form the basis for many of the proposed channel classification schemes for steep channels [i.e., Grant et al., 1990; Richards and Clifford, 1991; Montgomery and Buffington, 1993]. But no comprehensive theory yet exists to link flow hydraulics and bed forms across a range of channel types, bed material, and slopes.

As a first step toward developing such a theory, a new hypothesis is proposed here that argues that a similar mechanism of flow–bed form interaction as observed in sand-bed streams applies to a wide range of mobile-bed channels. Support for this hypothesis is developed in three parts: first by detailed examination of the mechanism observed in sand-bed streams, then through an empirical analysis of hydraulic data from a wide range of mobile-bed channels, and finally by reference to studies on diverse channels drawn from the literature. Taken together, these results suggest that the hypothesis is reasonable and consistent with the best available data; proof of the hypothesis awaits a more rigorous theoretical analysis of bed-flow interactions.

The Hypothesis

The hypothesis states that in mobile-bed river channels, interactions between the channel hydraulics and bed configuration prevent the Froude number from exceeding one for more than short distances or periods of time. The Froude number (Fr) is defined by

$$Fr = \frac{\alpha^{0.5} v}{(gd)^{0.5}} \quad (1)$$

where v is flow velocity, d is flow depth, α is the kinetic energy correction factor, and g is gravitational acceleration. The

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Froude number equals 1 at critical flow; when $Fr > 1$, the flow is termed supercritical, and when $Fr < 1$ the flow is subcritical. The hypothesis therefore states that supercritical flow should be rather uncommon in mobile-bed channels, except over short distances (i.e., tens of meters) and timescales (i.e., seconds to minutes). For the purposes of this paper, mobile-bed channels are defined as those competent to transport their bed material, that is, where total boundary shear stress, τ_0 , is greater than or equal to τ_{cr} , the critical shear stress for entraining grains on the stream bed. While this hypothesis applies to all mobile-bed channels, it is most relevant to hydraulically steep streams (i.e., those with gradients in excess of 0.01), since most low-gradient alluvial streams, such as the Mississippi River, are quite competent but have insufficient velocities for their depths to achieve near-critical flow.

The proposition that supercritical flow in channels is rare is not in itself new, and in recent years there has been considerable debate over whether the Froude number criterion can be used to constrain indirect flood and paleoflood measurements. Much of this debate has focused on finding empirical examples of subcritical or supercritical flood flows, with mixed results. For example, *Trieste* [1992, 1994] argues that many paleodischarge estimates that report supercritical flow are due to underestimates of roughness in mountain streams; similar conclusions are reached by *Jarrett* [1984]. On the other hand, *Wahl* [1993], *Simon and Hardison* [1994], and *House and Pearthree* [1995], among others, document examples of near-critical and supercritical flow in steep channels. Little attention has been paid, however, to either the underlying physical mechanisms that determine the flow regime or to the fact that critical flow itself may represent both a threshold condition and a common flow state. As will be shown, the tendency for Fr to increase with increasing slope under conditions of active sediment transport results in many steep, mobile-bed channels having a Froude number very close to unity. Furthermore, while it is quite common for flows to exceed criticality for short distances and periods of time, average Froude numbers are typically less than 1. Hence both sides in this debate may be correct.

Observations in Sand-Bed Streams

To test this hypothesis, I measured two small, sand-bed channels flowing over the seaward dipping, planar backshore of a beach on the Oregon coast (Figure 1a). Both channels were 6–7 m wide and had gradients of 0.012–0.018; the bed material had a median grain size (D_{50}) of 0.18 mm and a sorting coefficient (σ_ϕ) of 0.2 [*Broome and Komar*, 1979]. The longitudinal profiles of the two channels are independently imposed by wave swash during storms; tidal fluctuations change the base level to which these streams are adjusted, however, so slope can vary slightly over the course of several hours. The combination of constant discharge, high gradients at which the homogeneous, noncohesive fine sand can be readily transported, and absence of external controls (e.g., bedrock) mean that these channels are neither supply-limited nor energy-limited and hence have unlimited freedom to adjust bed forms and cross-sectional dimensions, within the constraint of the imposed gradient.

Both of the measured channels and all other streams observed on the beach displayed “pulsating flow” resulting from alternate formation and destruction of standing waves and antidunes. Cross-sectional and temporal variations in the Froude number were measured through a tidal cycle. Froude

numbers were calculated by (1) using instantaneous measurements of depth and velocity and assuming $\alpha = 1$. Depth was measured with a stadia rod calibrated and read to the nearest 0.5 cm; velocity was measured at 0.6 times depth with a recently calibrated Montedoro Whitney model PVM-2A electronic current meter with digital readout to the nearest 0.01 m/s and accuracy $\pm 1\%$. Depth of the velocity probe was constantly adjusted to maintain $0.6d$. To get instantaneous measured pairs of depth and velocity, I lashed the digital readout of the current meter to the stadia rod, videotaped both, and then sampled the videotape at 5-s intervals.

All measured cross sections had the same general pattern: average Froude numbers ranged from subcritical ($Fr < 1$) in the slower flow near the channel margins to near 1.0 at the center of the channel, with the average Froude number for the entire cross section slightly less than 1 (Figure 2a). The range of Froude numbers measured at a vertical section through time typically ranged from subcritical to supercritical ($Fr > 1$). At the channel center and thalweg the Froude number ranged between 0.7 and 1.3 around a mean of 1.0, over cycles that ranged from 20–35 s (Figure 2b). These oscillations corresponded to changes in bed and surface wave configurations as surface waves built, broke, and washed out the bed forms, returning the channel to a plane-bed condition with $Fr < 1.0$. These flow patterns persisted through channel incision, bank erosion, and planform changes accompanying the tidal cycle.

These channels illustrated the physical mechanism, also noted by others [*Kennedy*, 1963; *Foley and Vanoni*, 1977; *Bean*, 1977; *Schumm et al.*, 1982], underlying the proposed hypothesis (Figure 3). Accelerating, near-critical flow deformed an initially plane bed (Figure 3a) into a series of antidunes and in-phase surface waves (Figure 3b). Increasing velocity and decreasing depths and corresponding scour in the wave troughs caused the surface waves and associated antidunes to steepen, become unstable as the flow became supercritical in the troughs, and break upstream as hydraulic jumps (Figure 3c). This transition to supercritical flow in the troughs was accentuated by the tendency of the antidunes to migrate upstream, thereby becoming out of phase with the surface waves [*Kennedy*, 1963]. The downward flux of momentum from the breaking hydraulic jumps caused intense, localized bed scour, eroded the antidunes, restored the plane bed, and abruptly reduced the velocity (Figure 3d). Lowered velocities coupled with increased depths, as water previously stored in the stationary waves was released, caused the flow to become subcritical again and the cycle to repeat itself (Figure 3e). The key point is that the high-amplitude bed configuration caused by increasing flow velocity induced flow instability at flow slightly above critical, leading to very rapid energy dissipation and erosion of bed forms. This feedback resulted in unsteady, nonuniform flow around $Fr = 1$ (Figure 2b) and a cyclical creation-destruction sequence of bed forms [*Allen*, 1976].

Release of wave-stored water as the standing waves break typically results in the formation of a downstream migrating bore [*Foley and Vanoni*, 1977; *Bean*, 1977; *Schumm et al.*, 1982]. I observed bores due to wave collapse in the coastal channels with periodicities of 30 to 60 s, bore front heights of 2–4 cm, and speeds of 1–2 m/s. The bores contributed to the unsteady, nonuniform flow dynamics. The arrival of a bore from upstream into a train of standing waves caused disruption of the flow pattern and either accelerated the regular cycle by initiating breaking waves or else washed out the bed forms, reinitiating the wave-building sequence (Figure 3).



Figure 1. Examples of standing wave trains in fine-grained channels. (a) Study site at Big Creek, Oregon, showing building standing waves. Seagulls in background for scale. (b) Braided reach of the Pasig-Potrero River, Philippines, at low flow after lahar passage the previous day. Note standing wave trains in each of the anabranches.

These observations are not limited to coastal sand-bed channels but can readily be observed in other fine-grained channels with slopes in excess of about 0.01. Dynamics virtually identical to those of the Oregon beach channels were described by Schumm *et al.* [1982] for Medano Creek, a sand-bed channel in Colorado with similar slope. Clear water and fine-grained hyperconcentrated flows in channels subject to lahars in volcanically affected landscapes also display similar behavior [Pierson and Scott, 1985; Simon, 1992; Simon and Hardison, 1994]. For example, I measured a lahar-runout reach of the Pasig-Potrero River draining Mount Pinatubo in the central Philippines (Figure 1b). Because of the 1991 eruption of Mount Pinatubo and subsequent channel changes, the Pasig-Potrero now carries an extremely high sediment load, consisting of coarse sand and gravel derived from the 1991 pyroclastic field, along with pu-

miceous cobbles and boulders up to 20 cm in diameter. During the wet season, from June through September, thunderstorms with high-precipitation intensities produce frequent lahars [Pierson *et al.*, 1992]. Following each lahar, the low-flow channel reworks lahar-derived deposits into a 50- to 100-m wide shallow, braided channel with a gradient of 0.012 (Figure 1b). Individual braids of this channel have hydraulics identical to those of the Oregon coastal channels, including wave trains of building and breaking standing waves, periodic bores, and flow oscillating between subcritical and supercritical. I measured velocities by using floats and depths by using a folding stadia rod. In a straight, uniform 30-m reach, surface velocities ranged from 1.7 to 2.6 m/s, averaging 2.0 m/s, and depths ranged from 0.16 to 0.22 m, averaging 0.20 m. Because of the methods used, I could not instantaneously measure Froude

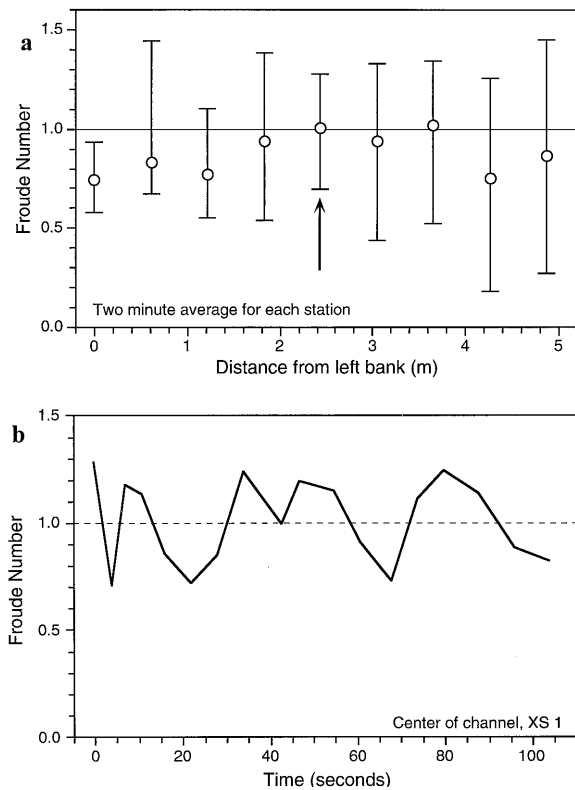


Figure 2. Variation of Froude number (a) over the cross section and (b) through time for Big Creek, Oregon. (a) Mean and range are shown for instantaneously measured values of Froude number at vertical sections sampled at 5-s intervals over 2–3 min. Average Froude number for the entire cross-section is 0.87. Arrow denotes center of channel section used in Figure 2b. (b) Variation in Froude number over approximately 2 min measured at center of channel; average Froude number is 1.00. Antidune and standing wave buildup accompany increasing Froude number; breaking waves and hydraulic jumps occur at maximum Froude number, and plane beds occur at lowest Froude numbers.

number. Based on the average depth and surface velocity, however, and assuming that the average cross-section velocity was 80% of the surface velocity [Mathes, 1956], average Froude number for the reach was 1.1.

Froude Number of Mobile-Bed Streams: An Analytical and Empirical Approach

The physical mechanism that causes hydraulics to oscillate around critical flow requires that the channel slope be great enough for the flow to approach critical when sediment transport is actively occurring. At a minimum this requires that the incipient motion threshold for the bed be exceeded. The Froude number at this threshold condition can be determined analytically by simultaneously expressing a flow resistance equation and a sediment transport relation in terms of the relative submergence (d/D_{84} , where d is flow depth and D_{84} is the 84th percentile grain size of the channel bed) and using these equations to solve (1) recast in terms of the channel slope. Because the average shear velocity v^* is equal to $(gdS)^{0.5}$, where S is the channel slope, (1) can be rewritten as

$$Fr = \frac{v}{v^*} S^{0.5} \quad (2)$$

Flume experiments by Bayazit [1983] have shown that in steep, hydraulically rough channels, the flow resistance is given by a Keulegan-type relation:

$$\frac{v}{v^*} = 2.18 \left[\ln \left(\frac{d}{D_{84}} \right) + 1.35 \right] \quad (3)$$

Also, critical dimensionless shear stress, τ_{cr}^* , is given by the Shields relation [Shields, 1936]:

$$\tau_{cr}^* = \frac{dS}{\left[\left(\frac{\gamma_s}{\gamma_w} \right) - 1 \right] D} \quad (4)$$

where γ_s and γ_w are the specific weights of water and sediment respectively, and $\tan \theta = S =$ channel slope. Assuming $\gamma_s = 2.65$, (4) can be rearranged as a relative roughness equation:

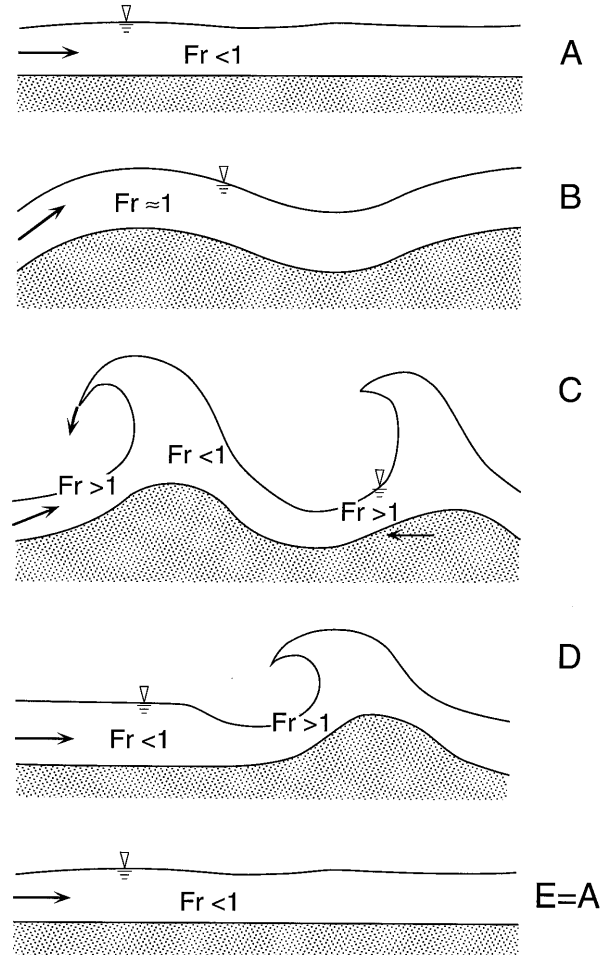


Figure 3. Cyclic sequence of surface wave and bed form deformation in sand-bed channels near critical flow. (a) Plane-bed, subcritical flow. (b) Building antidunes, approximately critical flow. (c) Breaking antidunes, flow both subcritical and supercritical. Note that upstream migration of antidune into wave trough induces hydraulic jump formation, causing standing wave to break. (d) Upstream wave has broken, downstream wave is about to break. (e) Scour of antidunes returns channel to plane-bed condition.

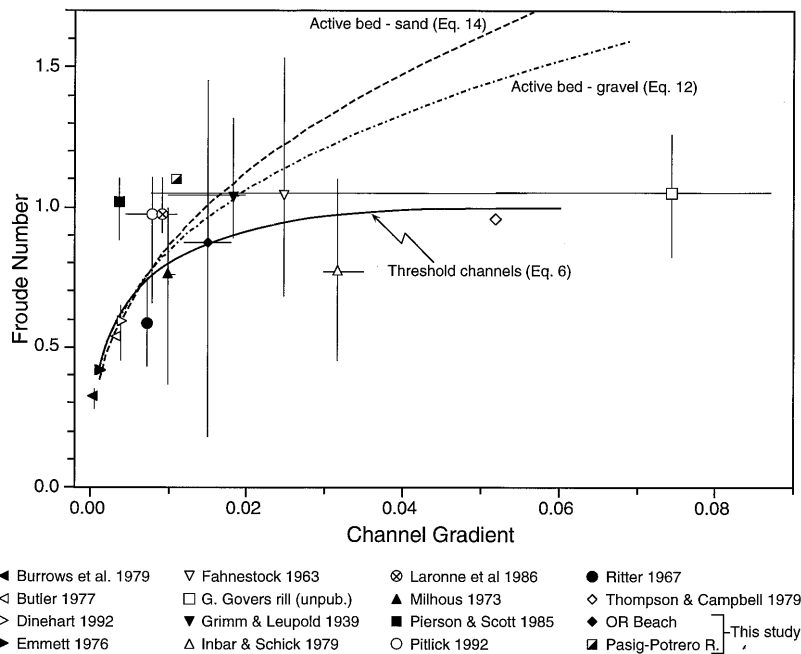


Figure 4. Plot of (6), (12), and (14) and published data for the relation between channel gradient and Froude number. Froude numbers are calculated using (1) with $\alpha = 1.0$. Means and ranges of Froude number and stream gradient for each data set are shown and, except for those of *Fahnestock* [1963], are based on individual measurements of mean cross-section velocity and depth under flow conditions where bed load was actively being transported. Points for *Fahnestock* are means and ranges of single-current meter measurements of maximum velocity and depth where bed load was being sampled.

$$\frac{d}{D} = 1.65 \frac{\tau_{cr}^*}{S} \quad (5)$$

By setting $D = D_{8.4}$ and substituting (5) into (3) and (2), we can calculate Froude number at incipient motion for a given slope as

$$Fr = 2.18 \left[\ln \left(1.65 \frac{\tau_{cr}^*}{S} \right) + 1.35 \right] S^{0.5} \quad (6)$$

Equation (6) therefore gives the Froude number at conditions of incipient motion for a specified slope and value of τ_{cr}^* . At $\tau_{cr}^* = 0.06$, which has been argued as the “channel-forming discharge” for gravel-bed streams [i.e., *Andrews*, 1983, 1984; *Parker*, 1990], (6) predicts that the Froude number will increase asymptotically toward 1.0 as friction slopes approach 0.02–0.03 (Figure 4). For lower values of τ_{cr}^* the Froude number increases but remains less than 1; higher values of τ_{cr}^* predict supercritical flows at incipient motion.

Because it is based on an assumption of steady, one-dimensional, uniform flow and relies on empirical flow resistance relations, this approach can only approximate the necessary channel slope; near-critical flow conditions are typically nonuniform and unsteady. This approach also neglects the potentially large component of flow resistance associated with bed forms as well as grain roughness [*Davies*, 1980] and does not account for the increased flow resistance created by free surface instabilities and hydraulic jumps at low values of relative submergence (i.e., $d/D_{8.4} < 1.5$) [*Ashida and Bayazit*, 1973; *Bathurst et al.*, 1979]. Both of these factors result in lower velocities for a given slope and would therefore tend to push the asymptote more to the right. An additional complicating factor is that flow resistance itself varies with Froude number

[*Bathurst et al.*, 1979; *Rosso et al.*, 1990]. Flume experiments by *Rosso et al.* [1990], however, demonstrate that in channels with gradients less than 0.05, only a 5–10% error in determining resistance is introduced by neglecting this effect.

Despite these uncertainties, the derived curve for $\tau_{cr}^* = 0.06$ is in reasonable agreement with published data for mobile-bed channels (Figure 4). This threshold curve represents a lower bound for competent flows; therefore all channels with active transport ($\tau_0/\tau_{cr} \geq 1$) should plot above the line, but incompetent flows ($\tau_0/\tau_{cr} < 1$) should plot below it. Most published hydraulic data for competent flows (i.e., where bed load transport was actually observed) for a wide range of stream types and slopes, including sand-bed, gravel-bed, and boulder-bed channels, lie on or above the predicted threshold (Figure 4). Froude numbers calculated from published values of depth and velocity may be somewhat lower than actual Froude numbers because most data report mean cross-section values of d and v , which are typically less than flow-weighted or maximum values. Most published velocity data also assume $\alpha = 1$, although it may be 10–15% larger [*Wahl*, 1993]. The highest reported and greatest range in Froude numbers are from point measurements made in glacial outwash streams where flow conditions were explicitly described as breaking antidunes [*Fahnestock*, 1963, p. A28]. As previously shown, point Froude numbers can be significantly greater than 1.0 even when the cross section as a whole is near-critical (Figure 2a).

Although there is some scatter in Figure 4, the data points are arrayed in a fairly narrow range above the lower envelope curve, but less than or equal to the critical flow line. Above a slope of about 0.01, all but one of the data sets report mean Froude numbers of 1.0 ± 0.2 . The mean Froude number for *Inbar and Schick's* [1979] data falls slightly outside this range;

however, most of their reported values are for storms with low bed load transport rates, where the large boulders that constituted the median grain sizes were not transported. Under these flow conditions the bed may not be adjusting. During storms when the largest boulders were moved, Froude numbers ranged from 1.0 to 1.1. Critical flow therefore appears to be a limiting condition for many mobile bed channels.

The analysis presented is based on consideration of threshold channels, that is, those that are near the incipient motion threshold. An upper bound for the relation between Froude number and slope can be derived from an analysis of mobile-bed channels, on the basis of Parker's [1990] work. Field data from gravel-bed channels suggest that maximum shear stresses in natural mobile-bed channels rarely exceed critical shear stress for the median grain size (D_{50}) by more than 20–50%, averaging around 40% for gravel-bed streams [Parker, 1982]. Then from (4) we have

$$1.4\tau_{cr}^* = \frac{dS}{\left[\left(\frac{\gamma_s}{\gamma_w}\right) - 1\right]D_{50}} \quad (7)$$

Use of the median grain size requires a slightly different formulation of the resistance equation. Parker [1990] uses a Manning-Strickler-type resistance equation rather than the Keulegan relation given in (3),

$$\frac{v}{v^*} = 8.1\left(\frac{d}{k_s}\right)^{1/6} \quad (8)$$

where k_s is a roughness height given by

$$k_s = \lambda D_{50} \quad (9)$$

and λ is typically near 6 (G. Parker, written communication, 1995). Combining (2), (7), (8), and (9) gives

$$Fr = \chi S^{1/3} \quad (10)$$

where

$$\chi = 8.1\left(\frac{1.4G\tau_{cr}^*}{\lambda}\right)^{1/3} \quad (11)$$

and G is the submerged specific gravity of sediment (1.65). With the aforementioned values for $\lambda = 6$ and $\tau_{cr}^* = 0.03$ for active gravel-bed streams,

$$Fr = 3.85S^{1/3} \quad (12)$$

This curve shows the same general trend of increasing Froude number with slope and provides an envelope curve for most of the plotted field data (Figure 4).

Recent work by G. Parker (written communication, 1995) demonstrates that for sand-bed streams freely forming braid anabranches and antidunes, shear stresses are typically 20 times τ_{cr}^* . For these channels the resistance equation (8) is modified:

$$\frac{v}{v^*} = 4.8(d/D_{50})^{0.11} \quad (13)$$

Then, by the same analysis and assuming that $\tau_{cr}^* = 0.06$ for sand-bed channels,

$$Fr = 5.18S^{0.11} \quad (14)$$

This equation predicts Froude numbers for active-bed sand channels that are slightly higher than those for gravel channels (Figure 4).

Because all of the curves shown in Figure 4 assume steady, quasi-equilibrium flow conditions and rely on empirical estimates of critical parameters, they should be viewed as defining the lower and upper ranges of Froude numbers for competent flows at a given slope. The assumptions and data used to derive these parameters become suspect at slopes much in excess of 0.04. Both the threshold and mobile-bed curves completely break down when slopes exceed 0.08 or so and sediment transport undergoes a transition toward debris flow [Takahashi, 1987]. The most salient feature of both data and curves presented in Figure 4 is that they all define a similar trend of increasing Froude number with slope and predict flow conditions near or slightly above critical for both threshold and active-bed, high-gradient streams. As slope increases above 0.01, flow is constrained to lie near critical for two reasons: first, the incipient motion threshold is associated with higher, near-critical Froude numbers; and second, bed form and free surface hydraulic interactions at high Froude numbers will tend to produce the oscillatory flow pattern previously described, restricting the flow regime to near-critical. The data clearly show that Froude numbers much in excess of 1.0 are uncommon in mobile-bed channels but that critical flow itself is common in streams with gradients above 0.01 (Figure 4).

Evidence for Critical Flow in Other Channel Types

Although the tendency for streams to adjust channel bed forms to maintain critical flow at competent discharges is most readily illustrated in sand-bed streams, the same phenomenon occurs in many other coarse-bed, high-gradient streams. Most workers consider transverse ribs in gravel-bed channels to be analogous to antidunes, forming by a similar process of bed deformation and hydraulic-jump development at critical flow [Boothroyd, 1972; Gustavson, 1974; Shaw and Kellerhals, 1977; Koster, 1978]. Allen [1983] gives an alternative explanation of ribs forming downstream of hydraulic jumps and creating critical overfall weirs; his model also relies on flow oscillating longitudinally around critical. Braided, gravel-bed rivers with slopes in excess of 0.01 tend to maintain critical flow in each of their braids over a range of discharges [Fahnestock, 1963; Schumm and Khan, 1972; Boothroyd and Ashley, 1975; Ergenzinger, 1987]. In fact, the pictures and descriptions of upper regime flow in gravel outwash streams provided by Fahnestock are virtually identical to the descriptions of the Oregon beach channels [Fahnestock, 1963, Figures 27 and 28].

Step pools are the dominant bed form in boulder-bed, mountain streams [Whittaker and Jaeggi, 1982; Chin, 1989; Grant et al., 1990; Ergenzinger, 1992; Abrahams et al., 1995; de Jong, 1995]. Flume experiments demonstrate that steps form when the Froude number approaches 1.0 and τ_0 barely exceeds τ_{cr} for the maximum grain sizes; step spacing varies with the wavelength of standing waves; and pools are scoured downstream of steps by hydraulic jumps (Figure 5) [Whittaker and Jaeggi, 1982; Ashida et al., 1984; Grant and Mizuyama, 1991]. In step-pool channels the narrow range of discharge values under which steps form, as well as the short time period during which those conditions typically occur, tends to restrict opportunity for repeated cycles of formation and destruction of steps; instead, steps represent disequilibrium bed forms, preserved as

flow conditions drop below those required to sustain boulder transport. The observed transition between step-pool channels with well-defined step bed forms and lower-gradient, plane-bed channels without distinct bed forms is at a slope of about 0.03 [Montgomery and Buffington, 1993]. The hypothesis presented in this paper suggests that the morphology changes because slopes are insufficient to generate critical flow in these coarse-bedded channels (Figure 4), and bed form roughness is replaced by bar roughness.

Additional support for this general hypothesis comes from a variety of published and unpublished observations or reconstructions of flows in steep channels. Recently published data for steep, gauged channels show that Froude numbers tend to converge toward 1.0 as discharge increases [Wahl, 1993]; this trend can also be observed in Milhous' [1973] data, which show Froude number increasing at a site with increasing discharge. Recent flume work on rill development show that in rills freely formed at very high slopes ($S > 0.05$) in tilled silty loam of moderate cohesion, without layering or anisotropy, Froude number remains constant at 1.0, independent of both slope and discharge (G. Govers, Leuven University, written communication, 1995). Paleohydraulic reconstructions of channel-forming floods through mountain channels also document flows at or near critical [Bowman, 1977; Grant et al., 1990; House and Pearthree, 1995] as do direct measurements of flow hydraulics [i.e., Beaumont and Oberlander, 1971; Lucchitta and Suneson, 1981].

There is also some evidence that hydraulics of hyperconcentrated flows on steep slopes are near-critical. Pierson and Scott [1985] describe the transition from lahars to hyperconcentrated streamflow in a channel draining Mt. St. Helens, Washington. During the latter phase, they note breaking standing waves and calculate Froude numbers of 0.9 to 1.1 at stream gradients of 0.003 to 0.004 (Figure 4). Their data suggest that hyperconcentrated flows may achieve critical flow at correspondingly lower slopes than for "normal" streamflows, perhaps owing to suppression of turbulence. The relatively high Froude numbers measured in the Pasig-Potrero river (Figure 4), where sediment concentrations were estimated to be 5% by volume and hence approached hyperconcentrated conditions, support this interpretation and help explain why these two data points fall above the predicted active-bed transport curve (Figure 4). In addition, assumptions regarding fluid density and resistance used to derive the envelope curves are not met for hyperconcentrated flows.

Recent work on floods in steep bedrock channels suggests that even where the boundary is nonadjustable, critical flow may still constrain flow hydraulics. Paleohydraulic reconstruction of flow hydraulics associated with rapid release of water from Lake Bonneville and associated cataclysmic flooding demonstrate near-critical flow conditions over long (tens of kilometers) reaches of channel [O'Connor, 1993, pp. 35–36; Figure 24]. Floods from the breakout of Lake Missoula were generally subcritical but locally achieved critical flow at constrictions and when channel slopes exceeded 0.01 [O'Connor and Waitt, 1995]. Constrictions of the Colorado River at alluvial fans in the Grand Canyon adjust to maintain near-critical flows through rapids because of scour by upstream migrating hydraulic jumps during competent flows [Kieffer, 1985]. Although in the case of bedrock rivers the bed is not adjusting to the flow, irregularities in the bed surface and protrusions into the flow may induce formation of hydraulic jumps, resulting in energy losses. Flow near critical is poised in the sense that

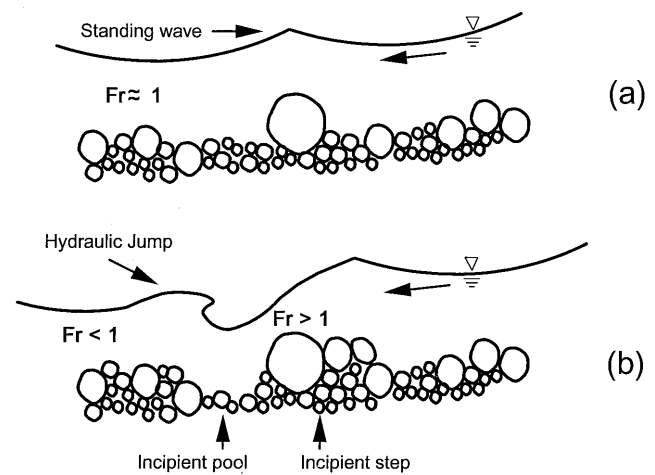


Figure 5. Model of step-pool formation in coarse-grained channels, based on flume experiments [Grant and Mizuyama, 1991]. (a) Large clasts come to rest under standing waves, trapping smaller clasts and beginning step formation. (b) Flow over clast locally becomes supercritical, returning to subcritical just downstream and forming a hydraulic jump; scour under jump creates pool.

shifts in flow regime and large amplitude changes in the water surface elevation can be triggered by small irregularities in the bed [e.g., Henderson, 1966, p. 45]. Supercritical flow can only be maintained in steep, hydraulically smooth channels, such as those lined with concrete described by Vaughn [1990], where Froude numbers typically range from 1.0 to as high as 2.5 or more over short (tens of meters) distances. Costa [1987] also reported indirect measurements of Froude numbers in excess of 1 for extremely high peak flows but noted that these were probably instantaneous values for short periods of time. Viewed from this perspective, the scatter of Froude numbers for adjustable channels in Figure 4 seems quite modest.

Discussion

Critical flow is highly efficient for routing water through channels. At critical flow, discharge per unit width is maximized for the available specific energy ($d + v^2/2g$), and specific energy is minimized for the available discharge [Henderson, 1966]. That steep channels adjust their bed forms and energy expenditures to maintain critical flow is therefore consistent with the hypothesis of minimum rate of energy expenditure and, under some constraints, maximum friction factor [Yang et al., 1981; Davies and Sutherland, 1983]. This parsimony arises not only because of thermodynamic principles, but also because of close coupling of the flow hydraulics with bed deformation and sediment transport. Convective accelerations in the flow field are balanced by development of bed forms and increased flow resistance. Critical flow represents an energy threshold beyond which flow instability (hydraulic jumps) results in rapid energy dissipation and morphologic change counteracting flow acceleration. The role played by hydraulic jumps is key; in channels ranging from sand to gravel to boulder bed, development of highly erosive and turbulent hydraulic jumps is the hydraulic mechanism that applies the "brake" to flow acceleration. The net result is a tendency for critical flow to constrain adjustment of channel hydraulics and flow resistance at competent discharges.

The relation between Froude number and flow resistance is more complex but may result in maximization of friction factor at critical flow. In general, the Froude number is inversely related to the Darcy-Weisbach friction factor, f , by

$$Fr^2 = \frac{v^2}{gd} = \frac{8}{f} S \quad (15)$$

where

$$\left(\frac{8}{f}\right)^{0.5} = \frac{v}{v^*} \quad (16)$$

Flume studies with flow over obstructions on fixed beds have shown that at low relative submergences ($y/D_{50} < 0.5$), the drag coefficient tends to be maximized at critical flow [Flammer *et al.*, 1970]. At higher relative submergences the peak drag occurs at lower Froude numbers. This effect appears to be due to the increased free surface drag resulting from hydraulic jumps and free surface distortions over submerged roughness elements at critical flow [Bathurst *et al.*, 1979]. Step-pool channels retain this highly dissipative morphology at less-than-formative flows because of the immobility of clasts forming steps. Recent work demonstrates that step spacing observed in natural channels may correspond with maximum flow resistance at formative [Ashida *et al.*, 1986] and less-than-formative [Abrahams *et al.*, 1995] discharges. By the model proposed here, maximization of friction factor must be viewed as a corollary to channel and flow adjustments rather than as a cause of them, as some extremal value theories have argued [Davies and Sutherland, 1980, 1983; Abrahams *et al.*, 1995]. Flow may still be critical at less than effective discharges; tumbling flow over static large roughness elements tends to be at or near critical [Morris, 1968]. Under these conditions, however, flow and channel geometry are not necessarily in equilibrium, so channel geometry is a relict of some more formative discharge.

In contrast to step-pool channels, which preserve bed morphologies developed at critical flow, antidune structures generated near critical flow in sand-bed systems are typically destroyed as flows wane, although an internal antidune stratification may be preserved [Middleton, 1965; Hand *et al.*, 1969]. This erasure of bed forms occurs because sand-bed streams tend to be competent over a much wider range of flows; hence bed morphology continues to adjust. In some cases the standing wave/antidune flow structure is maintained as flows recede, although the amplitude and wavelength diminish markedly. I have observed tiny sequences of standing waves in very shallow flows on the Oregon beach. Coarser gravel channels are in between sand-bed and boulder-bed streams, and a distinct relict antidune structure may be preserved as gravel lenses [Foley, 1977] or transverse ribs [Boothroyd and Ashley, 1975; Church and Gilbert, 1975; Rust and Gostin, 1981].

Jia [1990] argued that the equilibrium bed morphology for sand channels tends to minimize Froude number. This conclusion was derived from computer simulations of channels in the subcritical flow regime, mostly below the 0.01 slope threshold. In these simulations, where the slope equaled or exceeded 0.01, the minimum Froude number for the equilibrium channel approached 1.0 (cf. Figures 3 and 4 and Table 1 of Jia [1990]). The clustering of empirical points close to the threshold, as opposed to the active-bed lines, in Figure 4 similarly suggests that the Froude number will not exceed one for a given set of flow conditions.

In sand-bed streams, oscillation around critical flow occurs

in both time (i.e., Figure 2b) and space (i.e., longitudinal sequences of standing waves), but in boulder-bed and bedrock channels the oscillation occurs primarily in space, as the result of longitudinal accelerations and decelerations through the critical threshold. This is well illustrated in the paleohydraulic reconstructions of the Bonneville Floods [O'Connor, 1993]. Kieffer's [1987] detailed study of the hydraulics of rapids in the Grand Canyon also demonstrates this spatial variation, although she points out that flow may vary temporally because of surges and pulses similar to those observed in sand-bed streams. Release of water from breaking standing waves may produce unsteady flow, even if bed forms are not destroyed.

Further validation of this hypothesis poses logistical but not technical problems. Needed are instantaneous measurements of velocities and depths for cross sections representing a range of stream gradients and types, under high flow conditions when bed load transport is active. Instantaneous measurements are required because of the strongly unsteady nature of high flows. The primary logistical problems are access to high flows in steep channels and accurate measurements of velocity and depth in channels undergoing rapid bed deformation and sediment transport. Examination of stream gauge records to identify sites with high Froude number flows provides a first cut at candidate sites [Wahl, 1993].

From a practical standpoint an assumption of critical flow would dramatically simplify the many problems associated with predicting how floods behave in steep channels. Assumption of critical flow would essentially replace the standard resistance equations, such as Manning's or Chezy's, with the equation $v = (gd)^{0.5}$. Establishing the discharge of paleofloods in steep channels for hazard assessments, climate change analyses, or land-use planning could then proceed by calculating the velocity directly from the depth of flow, as determined from flood deposit stratigraphy or shear stress considerations. In fact, this assumption has already been proposed on empirical grounds alone, with good results [Allen, 1982, p. 411].

Conclusions

Interactions between the water surface and bed forms are proposed to maintain competent flows in mobile channels to $Fr \leq 1$. Competent, high-gradient streams with beds ranging from sand to boulders typically achieve an equilibrium adjustment between the flow, sediment transport, and channel morphology at or near critical flow. Under this unique hydraulic condition the maximum amount of water can be transmitted downstream for the available energy, and the tendency for flow acceleration is balanced by development of bed forms and flow structures that offset that tendency by dissipating flow energy. The adjustment to a critical flow threshold is a dynamic one, involving feedbacks between the free surface, bed configuration, flow resistance, and the sediment transport rate. In sand-bed channels this dynamic equilibrium generates pulsating flow with a concomitant nonuniform and unsteady flow field. In coarser-grained systems this oscillatory flow pattern is suppressed somewhat by the lower bed load transport rates, more chaotic structure of both the bed and the flow, and the importance of particle interactions; that is, collisions, imbrication, and lodging of very large particles influence the bed forms. The hypothesis that high-gradient streams adjust their hydraulics according to a common principle deserves further empirical and laboratory testing. If validated, it offers a simple, useful means of predicting flow hydraulics in mountain streams.

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