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Hyperconcentrated flow – transitional process between water flow and debris flow

Thomas C. Pierson

8.1 INTRODUCTION

Naturally occurring high-discharge flows of water and sediment in open channels (i.e., ‘‘floods’’), vary over a wide and continuous spectrum of sediment concentration and particle-size distribution. Water floods normally transport mostly fine sediment and in relatively small quantities (as a proportion of total flow volume), with suspended sediment having little effect on flow behaviour and concentrations generally less than 4% by volume (vol%) or 10% by weight (wt%) (e.g., Waananen et al., 1970). At the other end of the spectrum, especially in favourable geologic or geomorphic settings, high-discharge *debris flows* and *mudflows*[1] may transport more sediment than water. In these flows sediment concentrations are often in excess of 60 vol% (80 wt%)[2] (Costa, 1984, 1988; Pierson and Costa, 1987), and the sediment plays an integral role in the flow behaviour and mechanics (Wan and Wang, 1994; Coussot and Piau, 1994; Coussot, 1995; Iverson, 1997; see also Chapter 6). The term *hyperconcentrated flow* is most often applied to flows intermediate between these two end-members, although *debris flood* and *mud flood* have also been used. There is a subtle but important distinction to be considered in choosing the term, however. Debris floods and mud floods, discussed below, are extreme-magnitude, sediment-rich flow events, which may or may not

[1] Debris flows and mudflows are similar processes – complex, highly concentrated, pseudo-one-phase gravitational flows of sediment and water (Pierson and Costa, 1987; Wan and Wang, 1994). Debris flows are generally considered to contain more than 50% particles larger than sand size (Varnes, 1978), and varying physical interactions between coarse clasts and between clasts and fluid play significant roles in the flow mechanics (Iverson, 1997; see also Chapter 6). Mudflows (of the type studied by Chinese researchers and discussed in this chapter) are composed predominantly of silt, with some clay and fine sand. A combination of concentration-dependent and shear-rate-dependent, electrochemical and frictional interactions between particles largely determine flow behaviour (Coussot and Piau, 1994).

[2] A particle density of 2.65 g/cm^3 is used to compute volumetric concentrations from weight concentrations for kaolinite clay and silicate mixtures of silt, sand, and gravel; 2.3 g/cm^3 is used for smectite clay.

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involve the physical mechanisms characteristic of hyperconcentrated flows. It will be shown in this chapter that hyperconcentrated flow is a distinct flow process that can occur at low, as well as high, discharges.

Debris flood has been defined in a general sense as "a flood intermediate between the turbid flood of a mountain stream and a true mudflow" (Bates and Jackson, 1987). The approximately synonymous term *mud flood* (Gagoshidze, 1969; Committee on Methodologies for Predicting Mudflow Areas, 1982) has also appeared a few times in the literature. More specifically, a debris flood has been defined as "a very rapid, surging flow of water, heavily charged with debris, in a steep channel" (Hungr et al., 2001; see also Chapter 2), differentiated from debris flow in that it presumably maintains the characteristics of a Newtonian fluid and does not exhibit a surging or pulsating behaviour (Aulitsky, 1980). Although the term is discussed in classification schemes, it is generally not used in the English-language literature to describe actual events.

Floods that would qualify as debris floods or mud floods by the above definitions (but not necessarily labeled as such by the researchers who studied them) might include landslide-dam or log-jam breakout floods or thunderstorm-generated flash floods in relatively steep, narrow canyons. Such floods can move prodigious quantities of mud, sand, and gravel, including very large boulders, owing to high discharge, steep water-surface slopes, and valley constrictions, which keep flows deep and fast enough to produce high bed shear stress and strong turbulence. These floods are typically described by eye-witnesses (e.g., Glancy and Harmsen, 1975; Glancy and Bell, 2000) as: (a) arriving initially as a steep, fast-moving "wall" of boulders, logs, and other debris; (b) being very muddy in appearance and extremely turbulent; and (c) appearing overall as very muddy water, but occasionally having a consistency similar to fresh, wet concrete near the flow front (i.e., possibly including a debris-flow phase). Published examples of such floods include the 1964 dam-failure flood on the Rubicon River, California (Scott and Gravlee, 1968), the 1974 Eldorado Canyon flash flood, Nevada (Glancy and Harmsen, 1975), the 1976 Big Thompson River flash flood, Colorado (Balog, 1978; Costa, 1978), the 1982 Lawn Lake dam-failure flood, Colorado (Jarrett and Costa, 1986), the 1983 Ophir creek flood, Nevada (Glancy and Bell, 2000), and the 1996 Barranco de Aras flood, Spain (Alcoverro et al., 1999; Batalla et al., 1999). While debris floods or mud floods may achieve debris-flow characteristics toward the front of the flood wave or may include a submerged debris-flow phase along the channel bed (e.g., Scott and Gravlee, 1968), debris floods are primarily normal water floods or hyperconcentrated floods that are able to move large quantities of coarse sediment because of high discharges and/or steep channel slopes. Their deposits typically are composed of poorly stratified, loose mixtures of coarse sand and gravel, with the gravels commonly showing both an openwork texture throughout and an imbricated orientation of cobbles and boulders – characteristics not typical of debris-flow deposits. With the exception of the minor debris-flow phases, there are as yet no data nor compelling arguments to suggest that (a) basic bedload transport mechanics are not sufficient to account for the large-scale bedload movement in debris floods or mud floods, or (b) debris floods or mud floods should be classified as a separate process. Bedload transport is proportional to flow velocity raised to the exponent of 3–5, so dramatic increases in coarse-sediment transport can be expected in normal floods where high velocities are maintained and abundant coarse sediment is available (Mizuyama, 1981; Komar, 1988). Thus, the terms *debris flood* and *mud flood* may be descriptively useful to characterize a specific flood event, but they will not be discussed further in this chapter.

Evidence presented in this chapter from field observations and laboratory experiments strongly supports the argument that hyperconcentrated flow should be considered a separate flow process, not just a flow event. Flows with unusually high concentrations of suspended sediment were noted decades ago in streams and rivers in the semiarid and arid areas of the western USA (Lane, 1940; Bondurand, 1951; Nordin, 1963; Richardson and Hanly, 1965), in rivers draining the loess plateau of central China (Todd and Eliassen, 1940), and in freshly disturbed volcanic landscapes (Segerstrom, 1950). However, the term *hyperconcentrated flow* was first coined in 1964 (Beverage and Culbertson, 1964) to distinguish these muddy flows from normal streamflow, because of their tendency to clog irrigation canals and aggrade natural stream beds. Subsequent field and experimental studies have shown that natural hyperconcentrated flows are turbulent, two-phase, gravity-driven flows of water and sediment, intermediate in suspended-sediment concentration between normal sediment-laden streamflow and debris flow or mudflow. They appear to be fundamentally different from these other flow types on the basis of how sediment is transported, although more research is needed. Likewise, they present hazards that are somewhat different from those presented by water flows or debris flows and mudflows. This chapter reviews the various ways in which hyperconcentrated flows have been defined, explores ways in which they are different from other flow types, and synthesizes results from disparate studies to obtain a workable conceptual definition of hyperconcentrated flow.

8.2 DEFINING BOUNDARIES OF HYPERCONCENTRATED FLOW

Problems arise in trying to define hyperconcentrated flow. In water flow, the fluid mechanics are dominated by viscous fluid forces acting on channel boundaries and on individual entrained sediment grains that have little meaningful interaction with each other. In debris flow and mudflow, the flow mechanics involve complex combinations of physical particle interactions (friction and momentum transfer between coarse particles) and electrochemical particle interactions (double-layer and van der Waals attractions between fine particles), physical interactions between the sediment load and the bed, and strong but varying interactions between sediment grains and the fluid (Coussot and Piau, 1994; Coussot, 1995; Iverson, 1997; see also Chapter 6). The spectrum of physical processes at work in water flow and in debris flow or mudflow represents a continuum, and sharp, discrete demarcations between flow types probably do not exist.

Beverage and Culbertson (1964) initially defined hyperconcentrated flow as having a suspended sediment concentration of at least 20 vol% (40 wt%) and not more than 60 vol% (80 wt%). While these authors described some of the unique properties of hyperconcentrated flow, the boundary values were assigned without objective criteria that could be transferred to other settings. Since then, many other authors have simply taken these concentration limits and applied (or misapplied) them. In addition to sediment concentration, two other types of criteria have been used to distinguish hyperconcentrated flow: (1) criteria based on the bulk rheological properties of the suspensions, and (2) criteria based on how (and how much) sand is suspended and deposited in the flow. These approaches are discussed below. So far, a single, precise, comprehensive, and commonly accepted definition of hyperconcentrated flow has remained elusive.

8.2.1 Rheological criteria

Rheology, the study of the deformation and flow of materials, examines the bulk behaviour of two-phase sediment–water mixtures, within which complicated interactions between solid and fluid forces take place during flow. So while rheology may ignore the mechanistic details, rheological definitions of flow type have the advantage that tests sometimes can be performed in the laboratory (or observations made in the field) to define flow type.

Water behaves as a Newtonian fluid,[3] even when mixed with up to 35 vol% sand or gravel-size particles (Fei, 1983; Wan and Wang, 1994). However, smaller amounts of *fines*,[4] especially clay, added to a suspension will cause the onset of measurable yield strength[5] (Figure 8.1), marking the onset of non-Newtonian fluid behaviour. This transition from a Newtonian to a non-Newtonian fluid has been used by some authors to define the lower threshold of hyperconcentrated flow (Qian et al., 1981; Pierson and Costa, 1987; Rickenmann, 1991; Xu, 2002b, 2003). The upper threshold, namely the transformation to debris flow or mudflow, can also be defined in terms of yield strength – the point at which mixture yield strength, combined with buoyancy, is sufficient to fully suspend particles recognizable in the field as gravel,[6] whether or not the flow is moving (Pierson and Costa, 1987). A minimum yield strength of about 60 Pa is required, in addition to the buoyancy provided by the fluid, to suspend a 4 mm mineral grain in a static mixture of clay and water (Hampton, 1975), although turbulence in a flowing mixture can decrease the yield strength by breaking flocculent structures (Coussot and Piau, 1994; Wan and Wang, 1994). Support of coarse gravel in static debris-flow mixtures requires the additional support mechanism of grain-to-grain frictional contact (Pierson, 1981), and

[3] Newtonian fluids have a constant viscosity with respect to shear rate and will flow in response to any applied shear stress (i.e., they do not have internal strength to resist flow).
[4] "Fines" are considered by North American researchers to be particles <0.62 mm (i.e., the full range of silt and clay-size grains). In China, however, where much of the work on hyperconcentrated flow has been done, "fines" are considered to be only particles <0.01 mm (i.e., fine silt and clay). Unless otherwise stated, the North American convention will be used in this chapter.
[5] Yield strength (also termed shear strength) is the internal resistance of the sediment mixture to shear deformation; it is the result of friction between grains and cohesion.
[6] Gravel here refers to particles >4 mm diameter, because particles 2–4 mm are sometimes labeled "grus" and are hard to differentiate from sand.

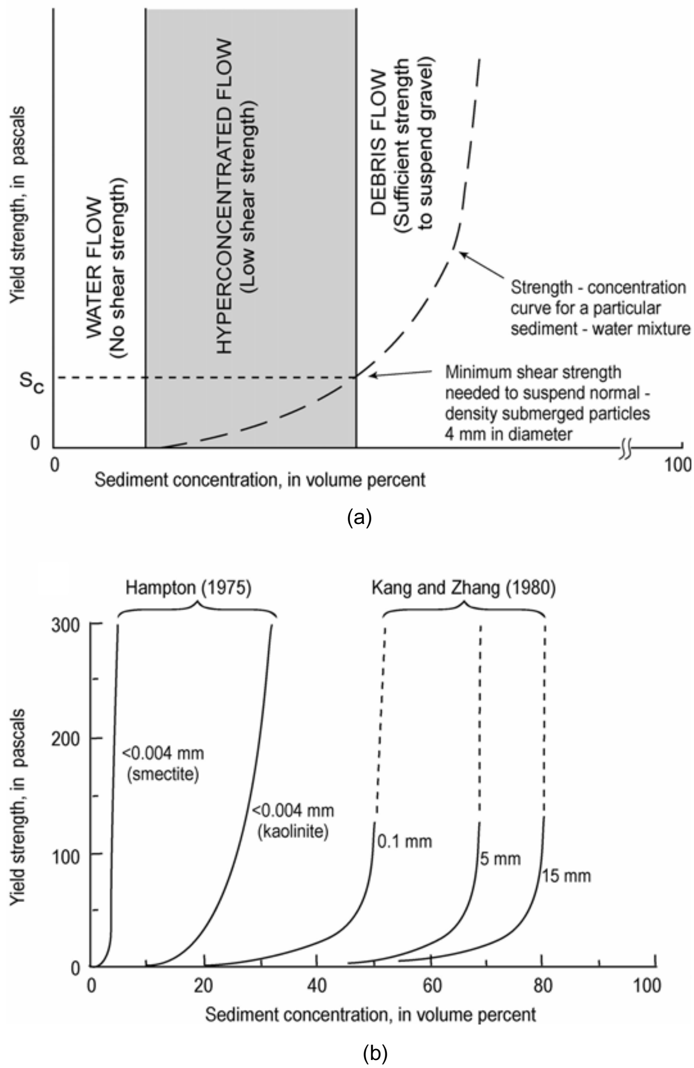

Figure 8.1. Yield strength of sediment–water mixtures as a function of suspended sediment concentration. (a) Definitions of flow type based on an idealized yield-strength curve of a poorly sorted sediment-water mixture. (b) Range in measured rheological properties of various sediment–water mixtures resulting from differences in grain-size distribution (average grain size indicated, in mm).

(b) From Pierson and Scott (1985).

increasing granular friction between more closely packed grains may explain the exponential steepening of yield strength curves (Figure 8.1) in response to increasing sediment concentration. Moving debris flows, however, involve a complex and variable interplay between solid grain forces and viscous fluid (Iverson, 1997, 2003; Iverson and Vallance, 2001, see also Chapter 6).

8.2.2 Sand suspension and settling criteria

Hyperconcentrated flows also have been characterized as flows in which large quantities of sand are transported in full dynamic suspension once minimum concentrations of fines (clay and fine silt) are achieved (Cao and Qian, 1990; Rickenmann, 1991; Dinehart, 1999). Field observations (Cronin et al., 1999) indicate that some fine gravel may be included with the sand. The point at which the proportion of sand in suspension abruptly increases relative to the suspended fines can also be used to define the lower boundary of hyperconcentrated flow. Hyperconcentrated flows characteristically have sand concentrations that greatly exceed the fines concentrations.[7] This sudden increase in the effectiveness of sand suspension is in agreement with a model for transition from low-concentration suspensions to intermediate or moderate-concentration suspensions (Druitt, 1995; Major, 2003), whereby upwelling of fluid displaced by the downward settling of some of the grains adds an additional buoyancy mechanism to the mixture, allowing the suspension to hold more sand. A critical part of this definition is that the sand and gravel still can settle out of the water column whenever energy of the flow decreases. Hyperconcentrated-flow deposits, therefore, commonly acquire some stratification and better sorting than the suspensions from which they settle out.

If sediment continues to be added to a hyperconcentrated mixture, however, a point is eventually reached whereby sand and gravel grains in the suspension begin to significantly interact with each other and frictional forces between grains hinder selective settling from the fluid suspension when the flow slows or stops (Druitt, 1995; Major, 2003). Frictional contact between grains prevents the larger and denser grains from settling faster than the surrounding finer and lighter grains. Consequently all the grains settle at the same rate and the result is an unsorted, unstratified deposit. Chinese researchers refer to this as the transition from "heterogeneous hyperconcentrated flow" to "homogeneous hyperconcentrated flow" or mudflow (Qian et al., 1981; Cheng et al., 1999; Cao and Qian, 1990). For suspensions of sediment from the Yellow River, this upper boundary is at about 19 vol% (Qian et al., 1981), but for coarser, more poorly sorted mixtures the boundary is between 50 and 55 vol% (Pierson, 1986; Cronin et al., 1999; Major, 2003).

8.3 CHARACTERISTICS OF HYPERCONCENTRATED FLOW

8.3.1 Initiation

Hyperconcentrated flows can initiate when water floods acquire added suspended sediment through erosion and entrainment or when debris flows lose coarse sediment through dilution and selective deposition. Documented initiation mechanisms include hillslope and channel erosion during intense rainstorms (Beverage and Culbertson, 1964; Major et al., 1996; Pierson et al., 1996), lake-breakout floods

[7] Some water floods may carry more sand than fines in suspension, but generally the fines fraction is significantly greater than the sand fraction (cf. Waananen et al., 1970).

(Rodolfo et al., 1996; O'Connor et al., 2001), glacier-outburst floods (Maizels, 1989), dilution and/or selective deposition at the heads and tails of debris flows (Pierson, 1986; Pierson and Scott, 1985; Cronin et al., 1999, 2000), and inputs of large sediment volumes to water floods by landslides (Kostaschuk et al., 2003). No matter what the water source is, an ample supply of easily erodible, relatively fine-grained sediment is critical. Most naturally occurring hyperconcentrated flows occur as floods (i.e., have higher than normal discharges), but in basins where sediment is extremely erodible they readily occur under low-flow conditions as well (Beverage and Culbertson, 1964; Montgomery et al., 1999). Studies have shown that there is often no direct relation between discharge and suspended-sediment concentration in natural occurrences of hyperconcentrated flow (Beverage and Culbertson, 1964; Komar, 1988; Alexandrov et al., 2003).

Hyperconcentrated flows occur commonly in semiarid and arid regions, particularly where basins are steep, hillslopes are erodible, channel banks are fragile, and channel beds are unarmored and erodible (Gerson, 1977; Laronne and Reid, 1993; Laronne et al., 1994). Although suspended sediment concentrations during floods in this terrain normally are in the concentration range of 0.5–5 vol%, much higher concentrations (in excess of 40 vol%) occur during extreme discharge events, during first runoff events of the season, or when landslides enter streams during floods (Lane, 1940; Beverage and Culbertson, 1964; Wannanen et al., 1970; Gerson, 1977). The loess plateau of central China has an especially high incidence of hyperconcentrated flows due to the combination of abundantly available fine sediment (thick deposits of eolian silt and fine sand), a subhumid to semiarid climate, and relatively steep, deeply incised channels resulting from regional tectonic uplift (Todd and Eliassen, 1940; Cao and Qian, 1990; Wang, 1990; Cheng et al., 1999; Xu, 1999). This region produces rainfall-runoff floods with maximum concentrations occasionally exceeding 50 vol%. Disturbed watersheds on volcanoes that have had recent explosive eruptions also tend to produce hyperconcentrated flows, owing to the widespread distribution of uncompacted fine tephra deposits (Segerstrom, 1950; Waldron, 1967; Scott, 1988; Major et al., 1996; Pierson et al., 1996; Rodolfo et al., 1996).

Two factors (found primarily at volcanoes) appear to be prerequisite for hyperconcentrated flows to evolve from debris flows: (1) the debris flows need to be sufficiently large to flow for long enough distances for flow transformations to occur, and (2) the debris mixtures must be relatively poor in fines (clay-rich debris flows tend not to transform). Where these conditions have been met, transformation from debris flow to hyperconcentrated flow has been inferred or documented (Janda et al., 1981; Pierson and Scott, 1985; Scott, 1988; Major and Newhall, 1989; Smith and Lowe, 1991; Scott et al., 1995; Major et al., 1996; Cronin et al., 1999, 2000; O'Connor et al., 2001).

8.3.2 Rheology

Viscosity is the property of a fluid that slows the settling of suspended particles and allows it to resist shear deformation, thus controlling the rate of shear or flow.

Viscosity increases with increasing sediment concentration in aqueous suspensions of silt and clay (Cao and Qian, 1990; Julian and Lan, 1991), and the effect can be dramatic. Suspensions of silt and clay containing between 15 and 45 vol% solids can have dynamic viscosities 1.5–4 orders of magnitude greater than the viscosity of clear water (Julian and Lan, 1991).

While Newtonian fluids such as clear water are deformable under any applied stress and have a constant viscosity that is independent of shear rate, suspensions of sediment and water commonly do not show either of these traits – viscosity can change with shear rate, and some minimum stress may be required to deform some mixtures. There is an extensive literature on the rheometry of fine-particle suspensions and a number of rheologic models have been proposed to explain their behaviour under applied stress (see Major, 1993 for a summary). Despite reported complexities in rheologic behaviour, flow of natural silt/clay suspensions is commonly considered to be reasonably well approximated by the Bingham model over naturally occurring ranges of grain-size distribution, sediment concentration, and shear rate (including rates representative of natural open-channel flows) (Fei, 1983; Wang et al., 1983; Engelund and Wan, 1984; Cao and Qian, 1990; Phillips and Davies, 1991; Major and Pierson, 1992; Wan and Wang, 1994). The Bingham model predicts a constant Bingham viscosity (μ_B), after a finite yield stress has been exceeded and can be represented as:

$$\tau = s + \mu_B \frac{du}{dy} \tag{8.1}$$

where τ is shear stress, s is yield stress, and du/dy is the velocity gradient. Barnes and Walters (1985) argue, however, that a true yield stress does not exist in fine-particle suspensions and that apparent yield stress disappears at very low shear rates ($< 10^{-4}\,\mathrm{s}^{-1}$), although such low shear rates may be irrelevant for natural flows.

In most natural fine-grained mixtures, laboratory measurement of shear stress over a wide range in shear rate (>3 orders of magnitude) results in demonstrable shear thinning (i.e., decrease in viscosity with increase in shear rate), which is probably due to the shear-induced breakdown of interconnected floc networks (Coussot, 1995). At relatively high shear rates (e.g., $100\,\mathrm{s}^{-1}$), even in mixtures containing some sand, particle interactions are minimized, and the viscosity of the interstitial fluid controls energy dissipation. At low shear rates ($<5\,\mathrm{s}^{-1}$), disrupted floc networks can reconnect (Coussot, 1995) and sand grains begin to physically interact (Major and Pierson, 1992). Both of these phenomena also cause the mixture rheology to deviate from the Bingham model by increasing apparent viscosity. This control of viscosity by shear rate can be better modeled by the Herschel–Bulkley model (Coussot and Piau, 1994; Coussot, 1995):

$$\tau = s + \mu \left(\frac{du}{dy} \right)^n \tag{8.2}$$

where μ is variable viscosity and the exponent n ($0 \leq n \leq 1$) defines the rate of shear thinning. When $n = 1$, this equation becomes the Bingham model.

8.3.3 Turbulence and velocity distribution

Much of the research into the effect of sediment loads on turbulence structure and velocity distribution in open-channel flow has been carried out on clear-water flows transporting large quantities of relatively coarse sand (no fines in suspension), and the conclusions are not always in agreement (discrepancies summarized by Wang and Qian, 1989; and Cao et al., 2003).

Limited experimental work, specifically with turbulent hyperconcentrated flows (large quantity of fines in suspension), suggests that vertical velocity distributions are essentially logarithmic in shape as they are in clear-water streamflow (Zhang and Ren, 1982; Yang and Zhao, 1983; Zhou et al., 1983), although experimental work by Bradley (1986) indicates that a power-law function may also adequately describe velocity profiles in hyperconcentrated flow. Overall, turbulence is somewhat dampened in hyperconcentrated flows by the higher fluid viscosity, resulting in less frictional loss of energy (van Rijn, 1983; Yang and Zhao, 1983), although flows transporting coarse bed load experience greater turbulence intensity near the bed and thus greater energy loss (Wang and Larsen, 1994). This effect of bedload intensifying near-bed turbulence has also been noted in clear-water flows (at normalized depths < 0.2). Causes of near-bed turbulence have been ascribed to greater bed roughness height, eddy shedding from large grains, grain inertial effects, and grain/bed form interactions (Best et al., 1997). Relative size of the bedload particles is an important variable (Cao et al., 2003). Turbulence has also been shown to increase when shallow hyperconcentrated flow moves over bed forms that have been immobilized by clay impregnation (Simons et al., 1963, p. G31). Decrease in flow velocity or increase in sediment concentration or suspended-particle size can alter the turbulence structure of flow and cause flow stratification to develop, whereby the upper part of the flow may become laminar and re-acquire a higher yield strength (floc networks re-establish) while flow near the bed remains turbulent (Wan and Wang, 1994; Wang and Larsen, 1994; Cao et al., 2003). Turbulence induced by coarse bed load can disrupt this stratified structure, however, and re-establish turbulence throughout the fluid column (Wang and Larsen, 1994). Rigid plugs (zones with zero velocity gradient) may develop in the upper, laminar part of the flow if concentrations approach the debris-flow/mudflow concentration threshold or the flow slows (Zhang and Ren, 1982; Yang and Zhao, 1983; Wilson, 1985) (Figure 8.2). Velocity distributions beneath such plugs remain logarithmic but with an increased velocity gradient (Yang and Zhao, 1983; Wan and Wang, 1994).

Field observations support the conclusion that many natural hyperconcentrated flows move principally as fully turbulent flows but with velocities near the river bed deviating from the logarithmic distribution (Zhou et al., 1983). Some damping of turbulence is noted, however (Pierson and Scott, 1985; Dinehart, 1999; Pringle and Cameron, 1999; Cronin et al., 1997, 1999, 2000). Large-scale boils, eddies, hydraulic jumps, and breaking antidune waves can be seen, but the small-scale choppy waves and splashing common waves in clear-water flows of equivalent discharge are commonly diminished or absent in relatively deep flows, and flow surfaces are often described as having an oily sheen (Figure 8.3(a)). One early observer (Pierce,

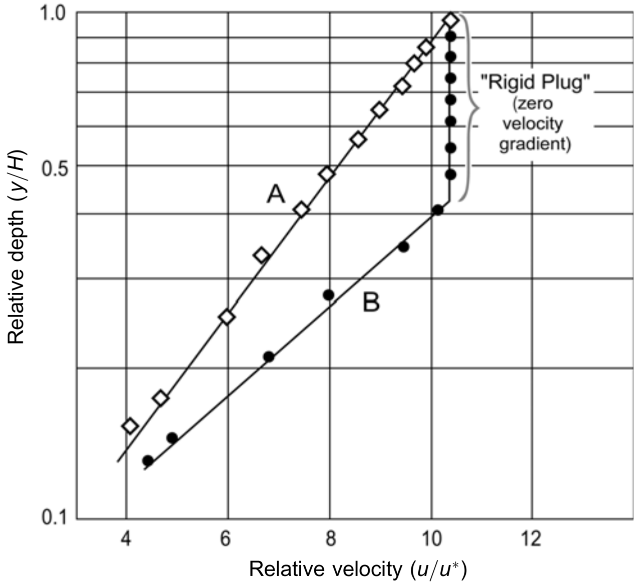

Figure 8.2. Vertical velocity profiles of experimental hyperconcentrated flow showing both a fully logarithmic profile (a) and a profile with a rigid plug above and a logarithmic profile below (b).

From Yang and Zhao (1983).

1917, p. 41) described one such flow at $Q = 370\,\mathrm{m}^3/\mathrm{s}$ and having close to 50 vol% suspended sediment: "... the river ran with a smooth, oily movement and presented the peculiar appearance of a stream of molten red metal instead of its usual rough, choppy surface." Rapid shallow flows do not exhibit this oily smooth surface, probably due to turbulence induced by coarse bedload or clay-encrusted bed forms (Figure 8.3(b)). Stratified flows probably also occur in the field, particularly at high concentrations, where turbulence near the surface vanishes while a turbulent layer is maintained near the bed. In Chinese rivers rigid plugs have been observed to form and thicken sufficiently to interact with the bed and bring flow in the channel to a standstill (Qian et al., 1981).

8.3.4 Sediment transport

Suspended load and the role of fines

There are two types of suspended load in hyperconcentrated flow, as in water flow: (1) fines (*wash load*) which form a stable suspension relative to the duration of flow and remain in suspension independent of concentration, flow velocity, or flow discharge (Cao and Qian, 1990; Xu, 1999), and (2) coarser particles (*intermittently suspended bed-material load* or simply *intermittently suspended load*) which remain temporarily in dynamic suspension for long periods of time relative to

Figure 8.3. Examples of smooth and rough flow surfaces of hyperconcentrated flows in Pasig-Potrero River, Mount Pinatubo, Philippines. Both were identified as hyperconcentrated by the high concentrations of suspended sand, but concentrations were not measured. (a) Hyperconcentrated flow (lahar) spilling over highway and creating a hydraulic jump on downstream side; (b) shallow hyperconcentrated flow on a sand bed with randomly choppy surface, possibly due to fines-crusted bed forms resisting transformation to a higher order form as noted in experiments by Simons et al. (1963, p. G31).

their size during flow (Komar, 1988; Bridge, 2003). Sand grains may only occasionally contact the bed while saltating cobbles may stay suspended for only seconds. However, all of the coarser particles that make up the intermittently suspended load will settle out of suspension as flow slows or stops. Although these two types of suspended load are present in both hyperconcentrated flow and water flow, the mean grain size of the intermittently suspended load is coarser in hyperconcentrated flow, much higher concentrations of intermittently suspended load are achieved in hyperconcentrated flow, and the relative concentration of intermittently suspended sand with distance from the bed is typically uniform in hyperconcentrated flow but strongly stratified in water flow. Intermittently suspended particles may be as large as small boulders in some hyperconcentrated flows (Cronin et al., 1999).

Concentration of fines appears to play a dominant role in the mechanics of transport of the intermittently suspended load. Field and experimental evidence suggest that high concentrations of sand cannot be transported in suspension in hyperconcentrated flows *unless* a minimum fines concentration is first achieved (Beverage and Culbertson, 1964; Gerson, 1977; Wan and Song, 1987). In this sense, the water plus wash load mixture can be considered to be the "carrier fluid" for the coarser suspended load (Wilson, 1985; Cao and Qian, 1990). Minimum wash-load concentrations appear to vary with the grain-size distribution of both types of suspended load (Table 8.1). At Mount St. Helens for example, a

Table 8.1. Minimum "fines" concentrations required to permit hyperconcentrated flows to suspend large quantities of sand. It should be noted that Cronin et al. (1999) measured 5 vol% suspended fines in measured flow LH7, which did not make the transition to hyperconcentrated flow (i.e., the relative amount of suspended sand remained low).

Location	Minimum fines required[a] (in wt% solids)	Minimum fines required[b] (in wt% solids)	Minimum fines required[c] (in vol% solids)	Reference
New Mexico and Arizona (Rio Puerco, Rio Salado, Paria River, Little Colorado River)		~15	~6	Beverage and Culberson (1964)
Toutle River, Mount St. Helens, Washington		~18*	~8	Dinehart (1999)
Yellow River basin, China	3–5*	Estimated 7–23 from range of grain-size curves (Qian et al., 1981)	3–10	Xu (2002b)

[a] "Fines" = particles $\leq$0.01 mm.
[b] "Fines" = particles $\leq$0.062 mm.
[c] "Fines" = particles $\leq$0.062 mm.
* Concentration values read from graphs of fines concentration with corresponding sand concentration, at the point at which sand concentration begins to increase significantly.

fines concentration of about 8 vol% is required before significant increases in suspended sand transport will occur, while at fines concentrations below 6 vol%, relatively little sand is suspended (Dinehart, 1999). During the passage of hyperconcentrated flow flood waves, fines concentrations tend to reach a maximum and fairly constant "plateau level" (even though sand concentrations may vary greatly), which may be due to supply limitations for the fines (Dinehart, 1999; Xu, 2002b). These maximum levels vary for different flows in different regions.

The roles played by the fines in helping keep sand in suspension during hyperconcentrated flow are complex. For decades it has been observed that high concentrations of suspended fines decrease fall velocity[8] of sand grains by increasing both fluid density (i.e., increasing buoyancy) and fluid viscosity (Simons et al., 1963; Nordin, 1963; Beverage and Culbertson, 1964; Xu, 1999). Dynamic viscosity can increase by as much as four orders of magnitude in going from clear water to a fine-sediment suspension of 45 vol% solids (Julian and Lan, 1991). Fall velocities for fine sand (0.125 to 0.25 mm) are reduced by an order of magnitude, whereas those for coarse sand (0.5 to 1.0 mm) are cut in half by fines concentrations typically found in hyperconcentrated flows in Arizona and New Mexico (Beverage and Culbertson, 1964). Bentonite suspensions of 4 vol% solids reduced fall velocity of various natural sands (0.1 to 1 mm diameter) 2.5 to 4 times (Simons et al., 1963). Thus, both yield strength and viscosity play important roles. Yield strength maintains the integrity of the carrier fluid by keeping fines permanently in suspension (Wan and Wang, 1994), and increased viscosity slows the settling of sand and coarser particles (Wang et al., 1983; Wan, 1985). Not coincidentally, yield strength is acquired at about the same point that sand concentrations begin to increase dramatically (Wang et al., 1983).

Suspension of coarse sand and gravel must rely on dynamic processes in addition to the fluid properties that hinder grain settling. These have not been studied in detail, but probably include increased drag provided by turbulent eddies in the carrier fluid, and upward dispersive stress provided by collisions between saltating grains and the bed and between grains themselves (Bridge, 2003). Also, experiments have shown that the quantity of suspended coarse load increases with increasing discharge (i.e., with increasing velocity and turbulence) (Wang, 1990), suggesting that the dynamic mechanisms of turbulence and grain collisions must play very important roles.

Coarse particles in hyperconcentrated flows will stay dynamically suspended so long as the sum of fluctuating upward-directed fluid flow and momentum transfer from grain collisions is greater than the particle fall velocity. Some of the upward flow results from turbulence and some from convection. Upward-directed fluid turbulence in hyperconcentrated flows, especially the action of macroturbulent vortices, observed in natural flows (Pierson and Scott, 1985; Dinehart, 1999), probably is fundamental in getting sand into suspension and helping to keep it there (Wilson, 1985; Komar, 1988). Convection occurs as a direct result of particle

[8] Constant fall velocity is reached when the submerged weight of a particle is balanced by the drag force imposed by the fluid (Wan and Wang, 1994).

Figure 8.4. Surface of hyperconcentrated flow revealing reticulate pattern apparently resulting from elutriation of fines at margins of convection cells caused by downward sinking coarse suspended load. Flow was the LH5 lahar in Whangaehu River (Ruapehu volcano, New Zealand), 27 September, 1995, 0945 hr, 42 km from source. Sediment concentration was 46 vol% (see Cronin et al., 1999).

Photo by Shane J. Cronin (Massey University).

settling – irregular cylindrical or planar zones of fluid are displaced upward by downward settling of larger, denser particles (Major, 2003). Such upward counter-flow can elutriate fines from within the suspension and carry them to the flow surface (Cronin et al., 1999; Major, 2003), sometimes resulting in the formation of swirling reticulate patterns on the flow surface (Figure 8.4).

Bedload

Bedload (or contact load) is the sum of all sliding or rolling particles that stay in more or less continuous contact with the bed, as well as the saltating particles that move close to the bed and are frequently in contact with it. In hyperconcentrated flow, the boundary between bedload and intermittently suspended load cannot be sharply defined. Little is known about bedload transport during natural hyperconcentrated flows, because standard bedload samplers cannot be deployed in flows with such high drag forces. It has been noted that the sound of boulders moving on the bed increases as the total suspended load increases (Pierson and Scott, 1985;

Dinehart, 1999). Cronin et al. (1997) observed boulders up to 2 m in diameter being rolled by hyperconcentrated flow having a sediment concentration of 37 vol%.

Recent theoretical and experimental work on high-concentration particle transport suggests that bedload sediment in hyperconcentrated flow may be transported in a concentrated zone of intense bed shear that has been referred as a *traction carpet* (Hanes and Bowen, 1985; Todd, 1989; Sohn, 1997). Observers who have reached hands into shallow ($\leq$1 m) hyperconcentrated flows have verified that an increasingly dense zone of moving bed material can be felt near the bed (Dinehart, 1999). A dense basal underflow layer (akin to a submerged debris flow beneath a more dilute flow) has also been inferred from field relationships (Scott and Gravlee, 1968; Cronin et al., 2000; Manville et al., 2000). In water flows, this dense zone of moving bed material has been subdivided into two zones: an upper *saltation* zone and a lower *collisional grain flow* zone (Hanes and Bowen, 1985). A somewhat similar model is used to describe motion within dry grain flows (e.g., Drake, 1990); it includes (1) an upper *collisional zone*, which is characterized by large gradients in particle concentration and velocity, active grain collision, high granular temperature, and generation of dispersive pressure; and (2) a lower *frictional* zone, which is a compact layer of slowly moving grains that are entirely in frictional contact with each other. Both models have been combined by Sohn (1997) in a conceptual model for bedload movement beneath turbulent overlying flows (Figure 8.5). It has been noted that variation in thickness of the total bedload shear layer, regardless of its internal mechanics, affects flow resistance and may lead to flow instability for hydraulic reasons (Wilson, 1985).

Vertical distribution of sediment

The vertical distribution of suspended sediment concentration is relatively uniform in turbulent hyperconcentrated flow – much more so than in normal streamflow (Qian et al., 1981; Yang and Zhao, 1983; McCutcheon and Bradley, 1984; Bradley, 1986; Rickenmann, 1991), and this uniformity of distribution depends on turbulence (Durand and Condolios, 1952). The concentration distribution is size-dependent (Figure 8.6) and can be predicted by the Rouse equation for distribution of relative concentration (Nordin, 1963, p. C10). Nevertheless, concentration will increase near the bed, and there may be no real demarcation between bedload and suspended load in the near-bed region, where there should be a continuous exchange of particles between suspended load and bedload (Wilson, 1985). As flow velocity decreases: (a) particles settle out of suspension according to size, (b) the average size of the intermittently suspended load becomes finer, (c) suspended sediment becomes progressively more concentrated near the bed, and (d) flow becomes more stratified (Wilson, 1985).

Transport rates

It has been well demonstrated that suspended-load transport rates for turbulent water flow and hyperconcentrated flow increase with increasing fine-material concentration (Simons et al., 1963; Kikkawa and Fukuoka, 1969; Wan, 1982; Bradley,

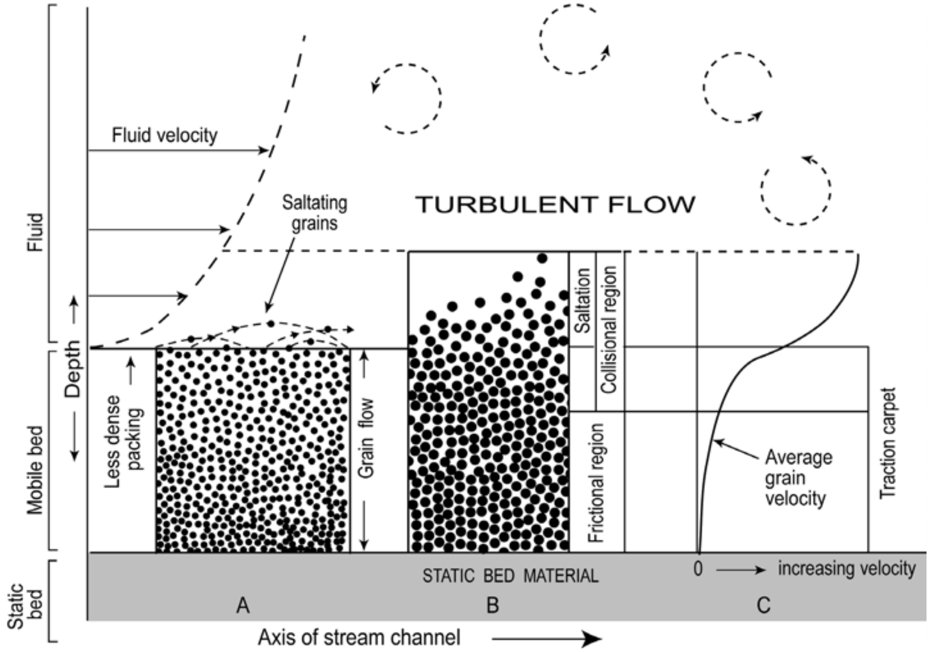

Figure 8.5. Generalized models for movement of bedload beneath overlying turbulent, shearing flow, which are presumed to describe mechanisms operative in a traction carpet in hyperconcentrated flow. (a) Model of Hanes and Bowen (1985) that postulates an internally shearing, collision-dominated granular fluid region beneath a turbulent fluid shear region. (b) Model of Drake (1990), based on high-speed photography of dry grain-flow experiments, that identifies (1) a basal frictional region with enduring frictional contacts between grains, which has a lower quasistatic zone and an upper block-gliding zone with groups of grains moving as coherent blocks, and (2) an overlying collisional region that is subdivided into a lower grain-layer gliding zone, a middle chaotic zone, and an upper saltational zone. (c) The conceptual synthesis of models A and B (Sohn, 1997) that predicts non-linearly increasing average velocities of grains with height above the static bed, up to the top of the traction carpet.

1986; Wan and Song, 1987; Dinehart, 1999). The average particle size of transported sand also increases with increasing fines concentration (Cao and Qian, 1990; Xu, 1999). Beverage and Culbertson (1964) predicted overall transport rate increases of about 500% for hyperconcentrated flow above water-flow transport rates. For over an hour during the 1982 hyperconcentrated flow at Mount St. Helens, the quantity of sediment in suspension was between 15 and 40 times more than during normal floods of similar magnitude (Pierson and Scott, 1985; Dinehart, 1999). Wan and Song (1987) predict overall increases in sediment transport rate of up to nearly 100 times more than normal streamflow.

Bedload transport rates are higher for hyperconcentrated flow than normal streamflow, as well. Bradley and McCutcheon (1985) calculated that bedload transport rate should increase between 50–70% for suspended sediment concentrations of 20 vol%, using well-known sediment transport formulas. Rickenmann (1991)

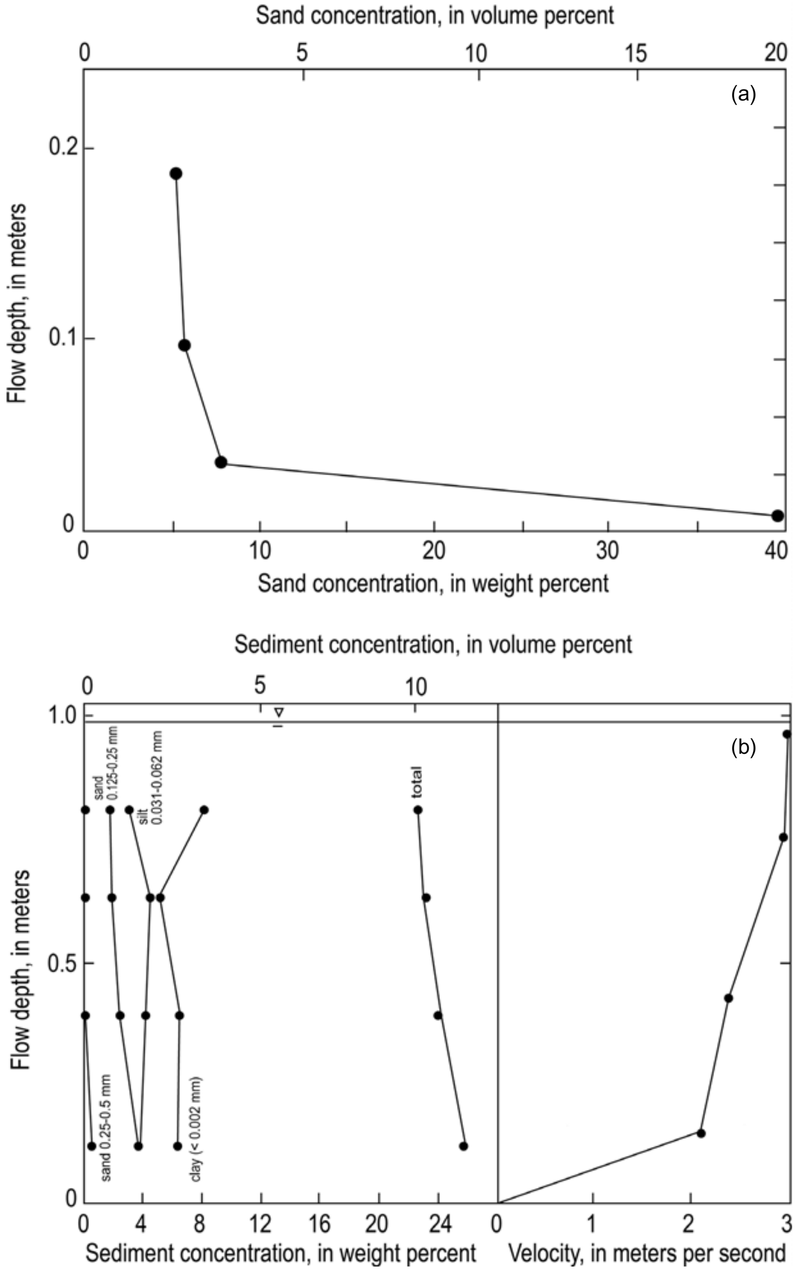

Figure 8.6. Vertical distribution of suspended sediment concentration in hyperconcentrated flows. (a) Experimental flow using sand ($D_{50} = 0.18\,\text{mm}$) added to bentonite clay suspension of 3.8 vol% concentration. (b) Apparent natural hyperconcentrated flow on Rio Puerco near Bernardo, New Mexico, station 185, 20 September, 1961.

(a) From Bradley (1986). (b) From Nordin (1963).

experimentally achieved bedload transport rates up to about 300% above clear-water rates, as fines concentration was increased up to a limiting value of 17 vol%. At concentrations above this limiting value, the fluid was less effective in transporting bedload. In turbulent flow, Rickenmann attributed this increase in bedload transport capacity to the efficacy of increased fluid density. Bradley's (1986) experiments showed an increase of two orders of magnitude in bed-material transport rates.

8.3.5 Bed material, bed forms, and flow resistance

Resistance to flow in an open channel is affected both by the internal properties of the fluid itself (fluid density and viscosity), by the shape and sinuosity of the channel, and by the drag imposed by the roughness of the bed. In alluvial channels, bed roughness can be further subdivided between particle roughness (size and shape of individual particles on the bed, mobile or stationary) and form roughness (shape and size of bed forms such as ripples, dunes, and antidunes).

Flume experiments have indicated that the roughness of the stationary bed and the presence or absence of relatively coarse bedload are important to flow resistance in hyperconcentrated flows. Hyperconcentrated flows flowing over rough beds experience less drag than clear-water flows at the same flow rate on the same beds, because the fines reduce turbulence intensity and suppress small eddies, therefore less energy is consumed and flow velocities are higher (Wan and Wang, 1994). Increasing the sediment concentration (in the range 3–9 vol%) can reduce flow resistance further. Even on smooth (sand) beds, flow resistance of the hyperconcentrated flow without coarse bedload can be less than that of the clear-water flow by about a factor of two (Shu et al., 1999). Also, it was found that flow resistance decreased with increasing suspended sediment concentration, but only until a minimum flow-resistance value was reached at a concentration of about 11–15 vol%; at concentrations above this threshold, flow resistance increased with increasing concentration (Figure 8.7). Yang and Zhao (1983) also concluded that flow resistance was commonly less in hyperconcentrated flow than in clear-water flow. With coarse bedload present that requires energy to transport, however, flow resistance is greater than in a similar flow without the bedload. The presence of coarse bed load increases turbulence intensity throughout the flow depth, transfers energy from mean motion to turbulence, and slows mean velocity (Wang and Larsen, 1994).

The effect of bed forms (form roughness) on flow resistance has been investigated for flows of clay–water suspensions at various concentration ranges: 0.4–1.0 vol% (Wan, 1985), 3–4 vol% (Simons et al., 1963), and 1–16 vol% (Wan and Song, 1987). In the lower flow regime (subcritical flow), a higher discharge is required to initiate particle motion on the bed for clay–water suspensions than for clear water, but the transformation from dunes to plane bed occurs at progressively smaller discharges as fines concentration is increased (Simons et al., 1963; Wan, 1985). In addition, the dunes that formed in flows with high concentrations of suspended clay were lower, smoother, more spaced out, and more symmetric, thus providing less

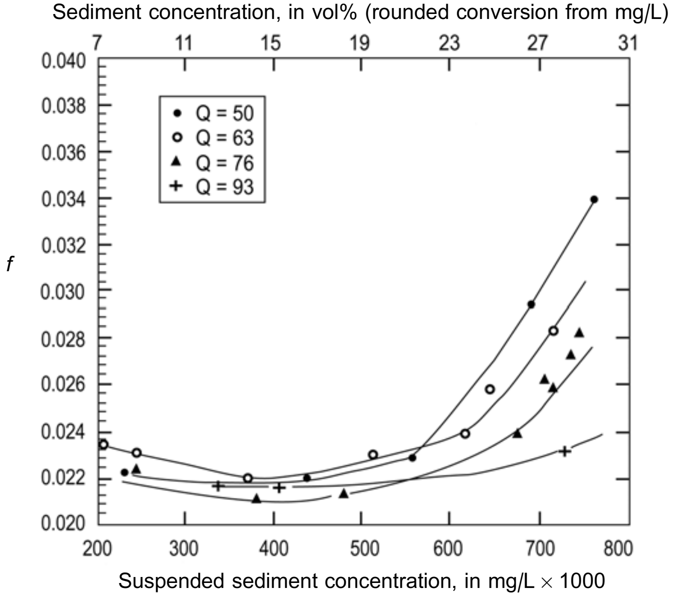

Figure 8.7. Relation between friction factor (f) and suspended sediment concentration for experimental flume runs at flow discharges between 50 and 93 L/s.
From Shu et al. (1999).

form resistance than those formed under similar flow conditions by clear water (Simons et al., 1963; Wan, 1985; Wan and Song, 1987). Thus in the lower flow regime, form roughness decreased with increasing suspended sediment concentration. However, the opposite effect was observed in supercritical flow – form roughness and flow resistance increased with increasing suspended sediment concentration. At constant discharge, the addition of fines caused standing waves to transform to breaking antidunes (with accompanying increase in turbulence and sediment transport) (Simons et al., 1963). This transition can increase flow resistance by a factor of 2–3 (Simons and Richardson, 1966). It was noted that the high fines concentrations could cause stabilization (crust formation) of ripple and dune bed surfaces due to bed impregnation by the fines. These clay-encrusted bed forms had to be broken up by higher flow discharges before higher order bed forms could be constructed (Simons et al., 1963).

Transverse and downstream-pointing V-shaped standing waves (non-breaking and breaking) have been observed in natural hyperconcentrated flows (Pierson and Scott, 1985; Dinehart 1999; Cronin et al., 1999, 2000). A progression in bed-form development was noted during flow (Dinehart, 1999), where apparent plane-bed conditions changed to standing waves and then to breaking antidunes while suspended sediment concentration increased and flow discharge decreased, supporting the experimental findings noted in the previous paragraph. Development of chutes and pools (bed forms providing still greater flow resistance) has also been observed in natural hyperconcentrated flows (cf. Pringle and Cameron, 1999).

The experimental work cited above predicts that hyperconcentrated flow should flow faster than clear-water flow for the same depth and slope, so long as flow remains predominantly in the lower flow regime and transported bedload is not excessively coarse. Furthermore, other experiments have shown that mean flow velocity in open-channel flow increases with increasing suspended-sediment concentration even before hyperconcentration levels are reached (Vanoni, 1946; Vanoni and Nomicos, 1960). These experimental conclusions are supported by field measurements at Mount St. Helens, where a high-discharge hyperconcentrated flow in 1982 had a mean velocity approximately 10% faster than several previous normal-concentration floods of approximately equivalent depth and discharge at the same gauging station (Pierson and Scott, 1985).

8.4 EXAMPLES OF OBSERVED HYPERCONCENTRATED FLOWS

Hyperconcentrated flows are relatively common in volcanic terrains recently impacted by explosive eruptions, and the best documented examples have been observed in volcanic terrains. In such settings rapid releases of water from a variety of sources will typically erode and incorporate exceptionally large volumes of unconsolidated deposits on hillslopes and in channels, and then flow downstream as *lahars*.[9] The following are three examples of lahars that were, for much of their flow paths, hyperconcentrated flows (Table 8.2), identified as such by their dynamic suspension of sand and fine gravel in large quantities. Equivalent data for non-volcanic hyperconcentrated flows have not been published.

8.4.1 1982 Hyperconcentrated lahar, Toutle River, Mount St. Helens, USA

The best documented example of a hyperconcentrated flow in North America is an eruption-triggered lahar that occurred on 19–20 March, 1982, in the Toutle River downstream from Mount St. Helens (Pierson and Scott, 1985; Scott, 1988; Dinehart, 1999; Pierson, 1999). It was caused by an explosive eruption that rapidly produced about $4 \times 10^6\,m^3$ of meltwater, forming a temporary lake in the crater. Breakout of the lake produced a pumice-charged water flood that eroded deeply into crater-floor volcaniclastic sediment and bulked to debris-flow concentration as it flowed out of the crater. Dip samples collected from the surface of the flow at three US Geologic Survey gauging stations 49, 73, and 81 km downstream from the source verified that it had transformed from debris flow to hyperconcentrated flow by the time it reached these stations. Estimated and measured peak discharges were about 960, 650, and $450\,m^3/s$, respectively – a flood peak attenuation of more than 50% over this 32-km reach. Average peak-flow depths ranged from 2.0 to 3.5 m. This example differs significantly from examples in central China in that the input of fines occurred primarily at or near source, and that the initial sediment had an extremely wide

[9] Lahars are "rapidly flowing mixtures of rock debris and water (other than normal streamflow) from volcano" (Smith and Fritz, 1989; Smith and Lowe, 1991). They can be hyperconcentrated flows or debris flows, and they commonly transform from one to the other. This definition is now gaining wide acceptance over previous usage that equated "lahar" with "volcanic debris flow".

range of particle sizes. High discharge hyperconcentrated flows in central China involve primarily silt and fine sand, and sediment inputs occur throughout the drainage basin wherever rainfall is contributing runoff.

Flood hydrographs and sediment sampling from the three stations (Figure 8.8) and field relationships show that: (1) the flood wave peaked quickly at each gauging station but peak magnitude attenuated as the flood progressed downstream and deposited large volumes of sand and gravel; (2) sediment concentration peaks lagged the flow discharge peaks with lag times increasing in the downstream direction; (3) fines concentrations plateaued at 250,000–350,000 mg/L (9–13 vol%); (4) suspension of large volumes of sand did not begin occurring until fines concentrations reached about 200,000 mg/L (8 vol%); (5) at peak concentration, the amount of sand in suspension far exceeded the quantity of suspended fines; and (6) peak total sediment concentration steadily decreased with distance downstream from at least 46 vol% (actual peak value not recorded at first station) to 41 vol% at the second station to 36 vol% at the third. Over this 32-km-long reach (average slope of 0.006 5 m/m), the flood deposited 37% of its sediment load. Of the total sediment in suspension at the time of peak sediment concentration, 70–80% was sand and fine gravel, although the suspended gravel was primarily lower density vesicular dacite and pumice (Dinehart, 1999). The coarsest suspended particles sampled at the flow surface were 16–32 mm in diameter. Plots comparing suspended sediment concentrations by size class with time (Figure 8.9) show that the coarser the grain size, the more quickly it settled out of suspension.

8.4.2 1993 Hyperconcentrated lahar, Sacobia River, Mount Pinatubo, Philippines

Monsoonal rain triggered a small (of the order of $10\,m^3/s$ peak discharge) hyperconcentrated lahar lasting about 2 hours in a broad sand and pumice-gravel-bedded braided alluvial channel on the lower east flank of Mount Pinatubo on 26 September, 1993 (Figures 8.10 and 8.11). It arrived at the observation point (20–25 km downstream from estimated initiation point) as a single low bore, 5–10 cm high, moving slightly faster than the ambient, already highly concentrated streamflow. Peak flow depth (in the distributary channel closest to bank), about 0.7 m, was reached in about 10 min, and flow discharge appeared to gradually recede as the thalweg moved to a mid-valley channel. The photographs in Figure 8.10 show both apparent plane-bed conditions accompanied by bank-parallel small waves and development of breaking antidune waves only 2 min later (slight increase in depth and possibly velocity). Total active floodway width was estimated at 30 m, although flow was concentrated in 3 or 4 distributary channels.

Crude depth-integrated sampling was carried out at intervals along the right bank using heavy-duty Ziplock®-type[10] plastic bags. The highest sampled suspended sediment concentration was 30 vol% (average particle density = $1.84\,g/cm^3$ from pumice content), which lagged about 10 min behind the apparent flow peak. When flow was at or near peak depth, sand and pumice gravel could be felt in

[10] For foreign readers, these plastic bags have a zipper-type closure that permits them to be sealed watertight; samples can be carefully transported to a laboratory for analysis without loss of fluid.

Table 8.2. Directly observed or measured characteristics of natural hyperconcentrated flows at approximate times of peak sediment concentration.

	Eruption-triggered lahar (3/19–20/82), Toutle River near Mount St. Helens (USA)	Rainfall-triggered lahar (9/26/93), Sacobia River on flank of Mount Pinatubo (Philippines)	Eruption-triggered lahar, LH6 (9/29/95), Whangaehu River on flank of Ruapehu volcano (New Zealand)
Observation location	73 km downstream from source	20–25 km downstream from source	42 km downstream from source
Flow discharge (m^3/s)	450	10 (estimate)	42 (peak, measured 14 km farther downstream)
Flow depth (m)	2.1 (mean estimated)	0.5–0.7 m (where sampled)	2–3 (estimated)
Flow velocity (m/s)	3.5–4 (estimated)	2.6–3.1 (surface)	2.7–3.0 (estimated)
Concentration peak lag behind flow peak (min)	38	Not measured	50
Suspended sediment concentration (vol%)	43	30	52
Silt + clay (fraction of total suspended sediment)	0.22	0.34	0.19
Maximum clast size in suspension	30–40 mm observed; 16–32 mm in dip samples (mostly low-density rock)	16–32 mm (low-density rock)	Up to 40 mm (normal-density rock)
Bedload movement	Boulder movement heard; observers described bed being "in motion"	Pebbles (coarser clasts not available)	Boulders; 0.5 m boulders observed saltating (surfacing momentarily)

Flow surface appearance	Oily, glassy-smooth; floating debris and pumice clasts	Slight smoothing of surface	Sandy-textured; had appearance of flowing quicksand; swirling reticulate pattern of fines; damped turbulence
Vertical sediment distribution	Sand and fine gravel in suspension near surface	Sand and fine gravel in suspension near surface; no sharp contact between bedload and suspended load	Fine gravel in full dynamic suspension; small boulders rising occasionally to surface
Bed forms	Oblique standing waves (antidunes) up to 2.5 m (estimate) high, generally breaking; alternating with plane-bed conditions	Transverse antidunes, non-breaking (0.3–0.5 m high); alternating with plane-bed conditions	Oblique standing waves up to 0.5 m
Sediment deposition	Channel bed aggraded 0.5–0.7 m; benches ~1 m thick formed at channel margins as stage dropped; deposits firm within minutes of deposition	Estimated ~10 cm bed aggradation	Benches with silty tops formed at channel margins
References	Pierson and Scott (1985); Dinehart (1999)	Pierson et al. (1996); T.C. Pierson, unpublished data	Cronin et al. (1999); written commun. S.J. Cronin (2004)

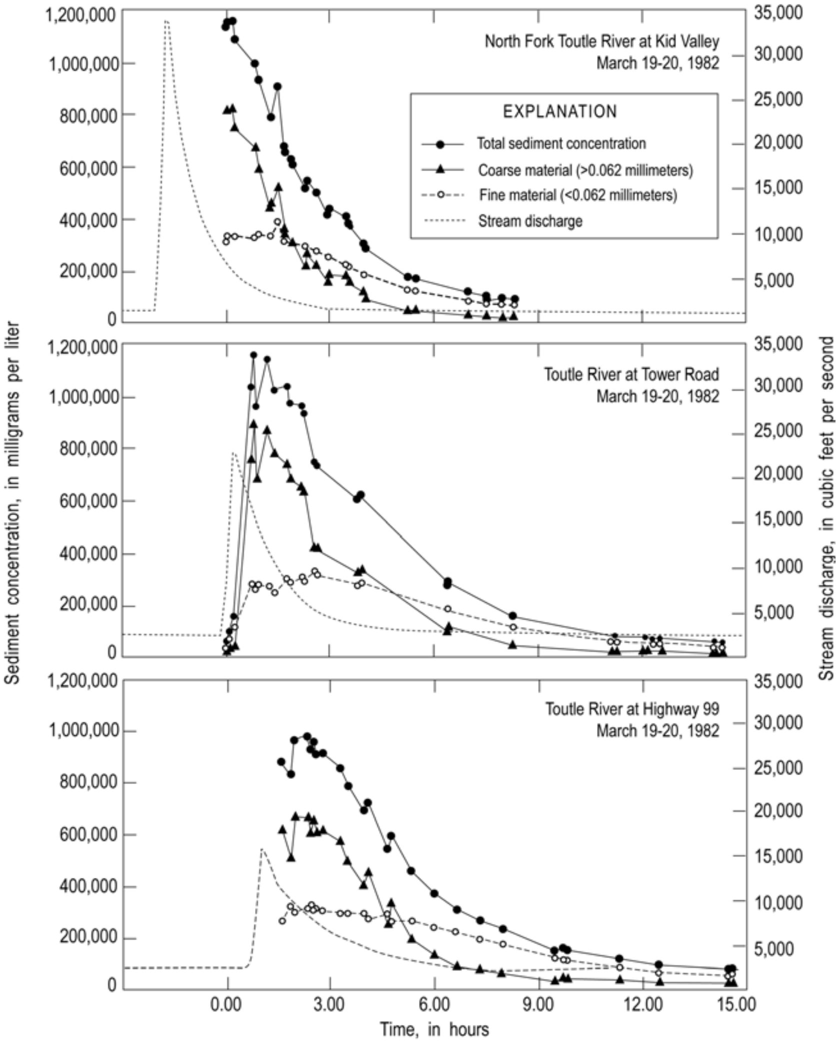

Figure 8.8. Sediment concentration and flow discharge for the 1982 hyperconcentrated lahar on the Toutle River downstream from Mount St. Helens, resulting from direct measurements of the flow at three USGS gauging stations.

From Dinehart (1999).

suspension throughout the vertical water column, and a denser layer of moving sand and gravel could be felt near the bed. In fact, it was difficult to identify by feel the boundary between the moving bed material and the static bed. Interestingly, fines concentrations in this hyperconcentrated flow were very low – all under 1 vol%. However, it is likely the sand-size and some gravel-size pumice grains would have fall

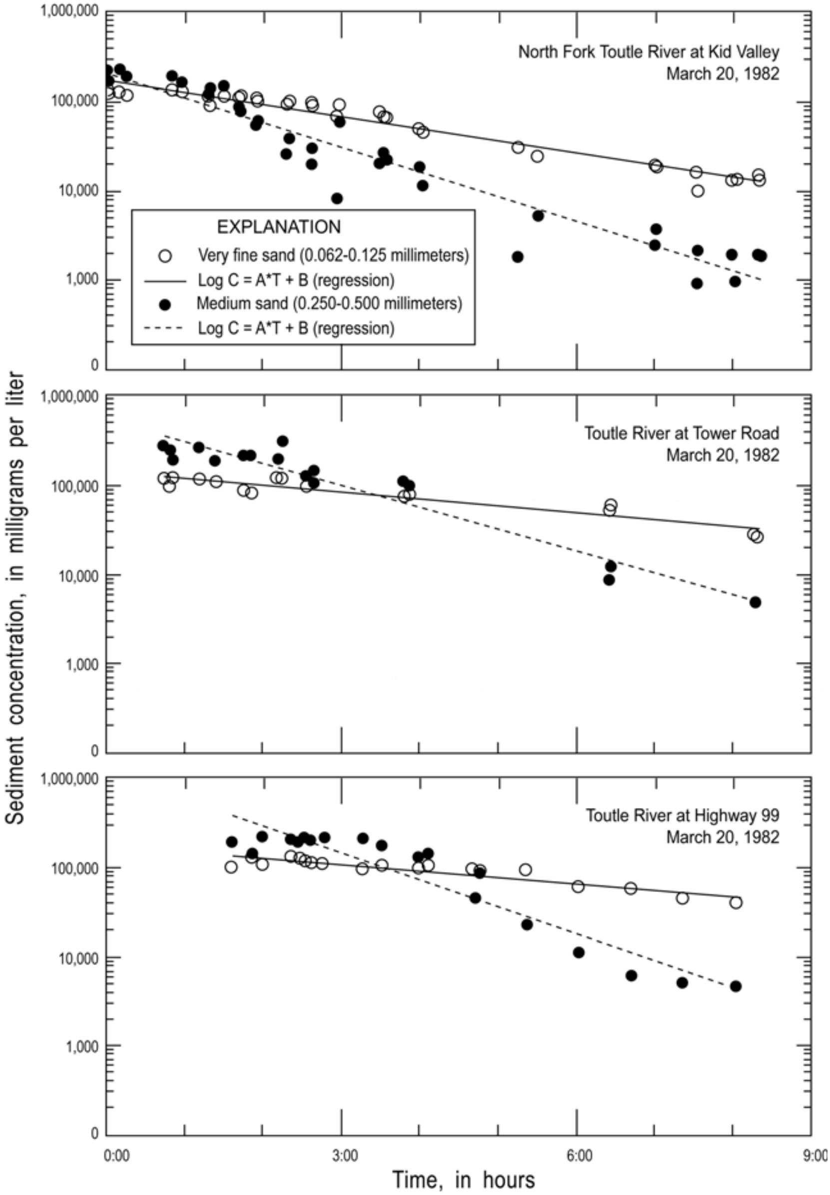

Figure 8.9. Sediment concentration vs. time during passage of the 1982 hyperconcentrated lahar on the Toutle River downstream from Mount St. Helens at different gauging stations, showing the more rapid settling out of the coarser size fraction as total concentration is decreasing in the tail of the flow.

From Dinehart (1999).

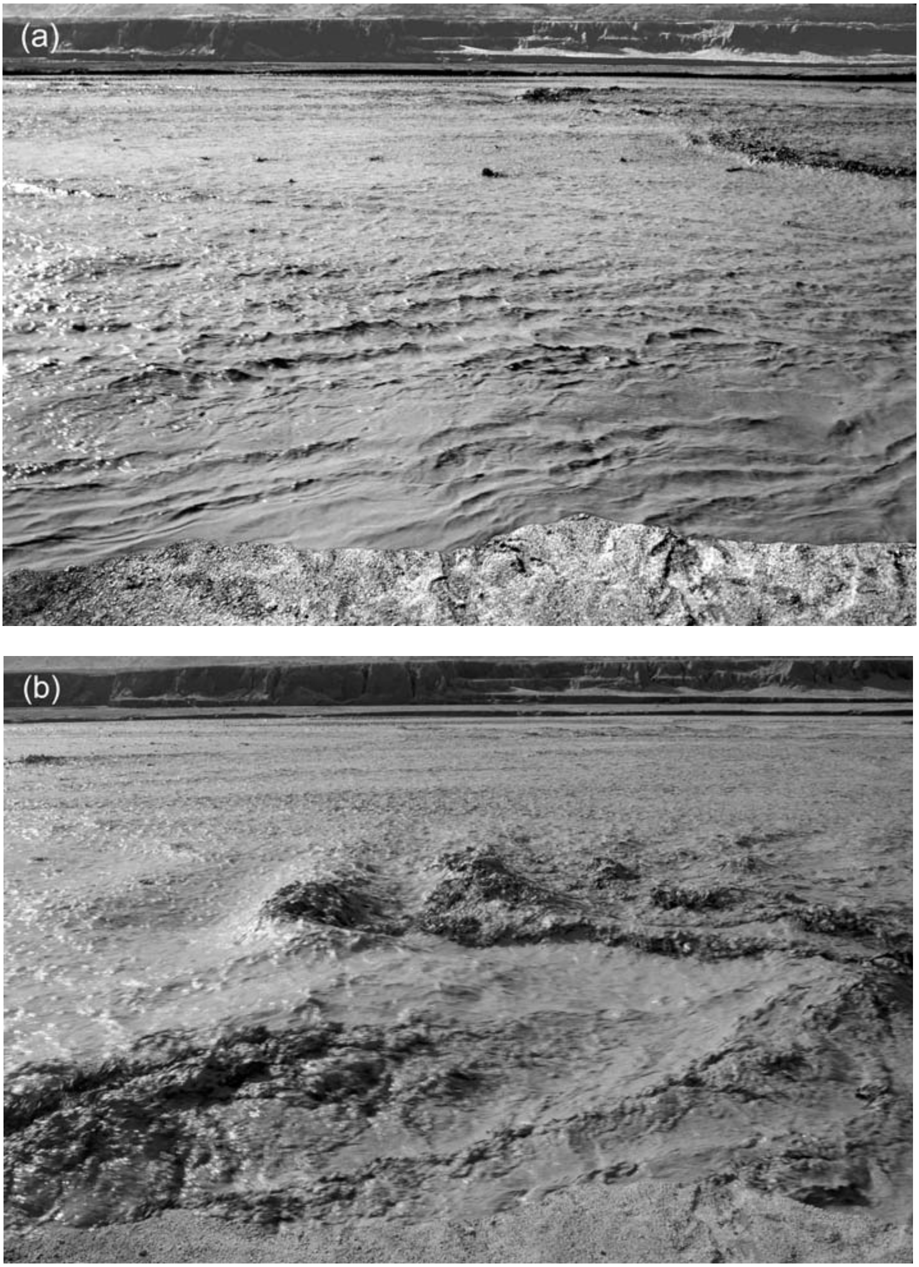

Figure 8.10. Hyperconcentrated lahar on Sacobia River, Mount Pinatubo, Philippines, 26 September, 1993, showing transition from plane bed to antidunes. Flow is left to right, and both photographs were taken from same position; time shown in upper right of each. For both, sediment concentration was about 23 vol%, flow velocity was about 2.5 m/s, flow depth was 30–50 cm; the transition to antidunes accompanied a slight increase in stage.

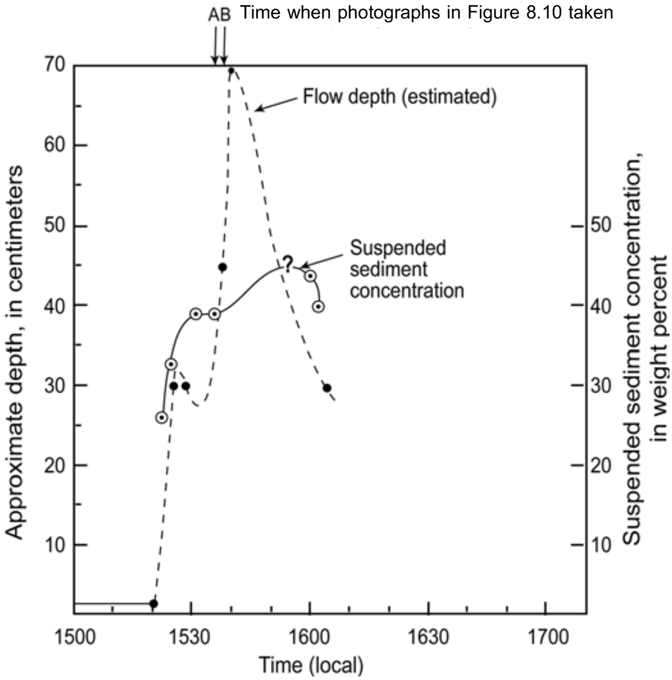

Figure 8.11. Approximate stage hydrograph and suspended sediment concentration variation with time for Sacobia River lahar (hyperconcentrated flow), 26 September, 1993, Mount Pinatubo (Philippines).

diameters in the silt range (particle densities as low as 0.7 g/cm^3) and thus would be acting like fines and supplementing the silt-size particles in forming the wash load.

8.4.3 1995 Hyperconcentrated lahar, Ruapehu volcano, New Zealand

Numerous lahars were generated on the eastern flank of Ruapehu volcano in late 1995 by pyroclastic debris erupting explosively up through the crater lake (Cronin et al., 1997, 1999, 2000). Several single-peaked and relatively sustained flows were generated on 29 September that were observed, measured, photographed, and sampled in detail (Cronin et al., 1999). The LH6 flow (Table 8.2) was identified as fully hyperconcentrated (by the criterion of dynamic suspension of sand and fine gravel) at observation stations 42 and 56.5 km from source, having transformed from a debris flow farther upstream.

This single-peaked lahar moved downstream in four phases, as did the smaller and more dilute (probably not hyperconcentrated) LH7 flow. The phases could be distinguished by sampling the relative mix of chemically distinctive (highly acidic) lake water and the stream water as the flows passed by the observation points (Cronin, 1999):

1. *Rising limb of the flood wave*. At 42 km from source, LH6 stage rose relatively rapidly – 0.5 m in 50 min, and the rise to peak stage/discharge was accompanied by only a slight increase in suspended sediment comprising mostly silt and clay (up to 4 vol%), by an increase in floating woody debris, but by essentially no change in water chemistry. This last observation suggests that this part of the flood wave was only stream water that was "piling up" and being pushed ahead by the actual lahar that was moving faster than the stream.
2. *Arrival of lahar front as stage recedes*. Within the next hour stage dropped to about 60% of peak, but a large and fairly abrupt jump in dissolved sulfate, chloride, and magnesium, and a drop in pH, indicated that the packet of ejected water from the crater lake had arrived, accompanied by an increase in suspended sediment concentration that reached a peak of 52 vol%. Increases in both sediment concentration and ion concentration were more gradual in another lahar (LH7), indicating that lahar fronts can mix with stream water as well as push it ahead. Flow turbulence was dampened but nonetheless active. Slowly saltating boulders surfaced for a few seconds and then sank back down, and the flow surface had the appearance of flowing quicksand.
3. *Lahar recessional limb*. Suspended sediment concentration decreased to 33 vol% during this phase of the flow, but dissolved ions and acidity continued to increase, indicating that the flood wave of ejected lake water was being diluted by stream water at the front of the lahar. Gravel and sand (normal-density and scoriaceous) remained in dynamic suspension, and small boulders continued to saltate up to the flow surface.
4. *Highly erosive lahar tail*. Both water acidity and sediment concentration decreased during this phase, and flow transitioned back to normal streamflow. However, vigorous turbulence while sediment concentration remained above 10 vol% apparently triggered an acceleration of bank erosion in the channel.

The observations of Cronin et al. (1999) are important in clarifying the difference between debris flows and hyperconcentrated flows in coarse sediment mixtures. At their upper-concentration range, hyperconcentrated flows superficially can look like debris flows, with gravel in suspension and even having boulders bobbing at the surface. The key, however, is that the suspension of sand and gravel depends on dynamic processes that occur during flow. Once the fluid motion stops, sand and gravel readily settle out of suspension.

8.5 SEDIMENT DEPOSITION

Sediment is deposited in two ways in hyperconcentrated flows – by settling out of suspension (suspension fallout) and by traction-carpet accretion. Although a variety of depositional features have been attributed to hyperconcentrated flows and described in the sedimentological literature, the brief discussion here is limited to observations of deposits immediately after hyperconcentrated flows have passed.

Deposition by suspension fallout should occur (1) where velocity, and hence turbulence, are decreasing, and (2) where dilution by addition of stream water at flow-front mixing interfaces or at tributary inflows result in loss of suspension competence (Pierson and Scott, 1985; Cronin et al., 2000). At energy-drop locations (such as along channel margins, where flow depth decreases, in eddies, or below hydraulic jumps), the fallout from suspension should be a function of flow energy at that point. A relatively narrow range of grain sizes should rain down if discharge remains fairly constant and stratification should be only weakly developed or absent. Relatively massive and well-sorted deposits have been observed along some channel margins, for example (Pierson and Scott, 1985; Cronin et al., 2000). In suspension fallout zones where flow discharge fluctuates or where turbulence intensity varies (e.g., at mixing fronts), deposits can show weak to strong horizontal bedding (Cronin et al., 2000) (Figure 8.12(a)). Mud layers commonly form on the tops of these depositional units after subaerial exposure, due to vertical dewatering of the deposits with transport of fines to the surface, and upper portions of the deposits are commonly normally graded (Cronin et al., 2000).

Deposition by traction-carpet accretion should be greatest where flow velocities, and hence bed shear stresses, are high. Therefore, deposition along channel thalwegs should comprise a high proportion of traction-carpet accretion deposits. Here, bed material is deposited largely as layers or sheets of grains accreted from the base of the mobile traction carpet, and deposits are coarser than at channel margins (Cronin et al., 2000). These deposits commonly show more pronounced horizontal stratification than channel margin-deposits, particularly when contrasting grain types are available, but laminations typically occur without high-angle cross-bedding (Figure 8.12(b), lower part). Coarse bedload (lenses or single outsized clasts of cobbles or boulders) is commonly enveloped by finer accretionary strata or is left stranded on surfaces of berms or terraces. In distal depositional areas where flow sediment concentrations are decreasing significantly and progressively more sediment is moving as bedload, deposits become more distinctly stratified and high-angle cross-bedding can be formed and preserved (Cronin et al., 2000).

A third general type of deposit is also sometimes observed in channels after the passage of highly concentrated hyperconcentrated flows: massive, very poorly sorted, but only moderately compacted diamicts of sand and gravel, which are more friable than typical debris-flow deposits (Pierson and Scott, 1985; Cronin et al., 2000; see also Figure 8.12(b), upper part). It is uncertain whether these deposits represent a submerged debris-flow phase that lags behind peak discharge and flows beneath more dilute surface flow (i.e., a stratified lahar (the explanation favoured by Cronin et al., 2000)), or whether these deposits result simply from very rapid and chaotic suspension fallout in the rapidly fluctuating, high-energy environment found in the deepest, fastest, most concentrated part of the flow.

Overall, median grain size in all types of deposits typically decreases progressively with distance downstream (Pierson and Scott, 1985), and deposits are better sorted than the fluid mixtures that deposit them (Pierson and Scott, 1985; Cronin et al., 1997, 2000; see also Figure 8.13) – most of the silt and clay remains in suspension while the sand and gravel are deposited. In addition, hyperconcentrated flow deposits are relatively similar over hundreds of metres along a channel (Cronin

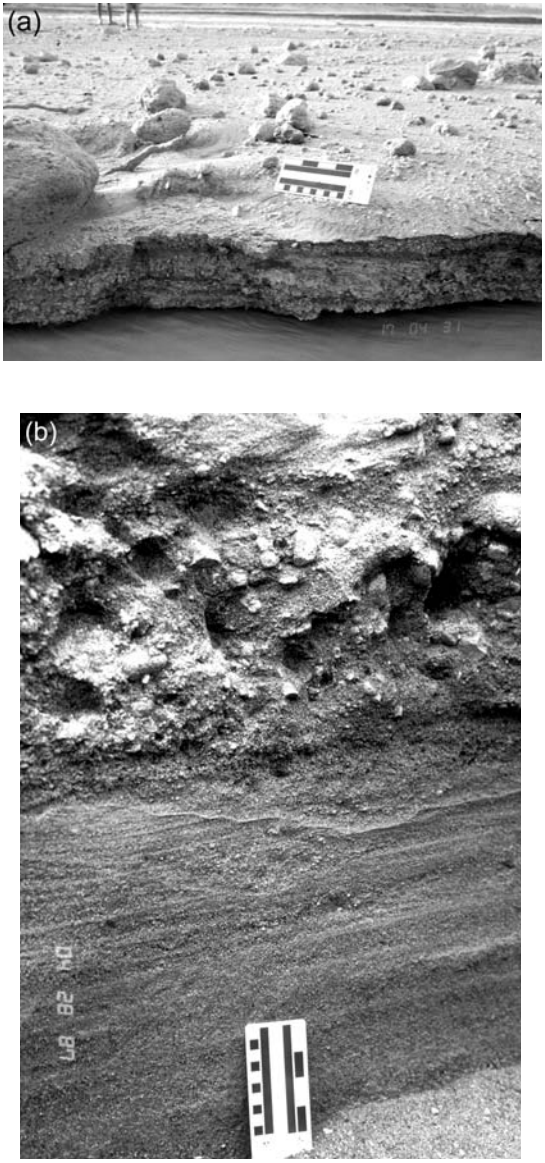

Figure 8.12. Examples of hyperconcentrated flow deposits. (a) Laminated deposit from 26 September, 1993 Sacobia River lahar, Mount Pinatubo (see Figure 8.11). Laminations emphasized by contrasting sediment grains – dark lithic medium sand and light pumiceous very coarse sand/fine gravel. (b) Section near main axis of channel showing two contrasting deposit types from the 1982 Toutle River lahar at Mount St. Helens. Lower part of section – faintly, horizontally stratified medium to coarse lithic sand; upper part of section – massive, poorly sorted diamict (resembling a debris-flow deposit but not as indurated because few fines are deposited with the sand and gravel).

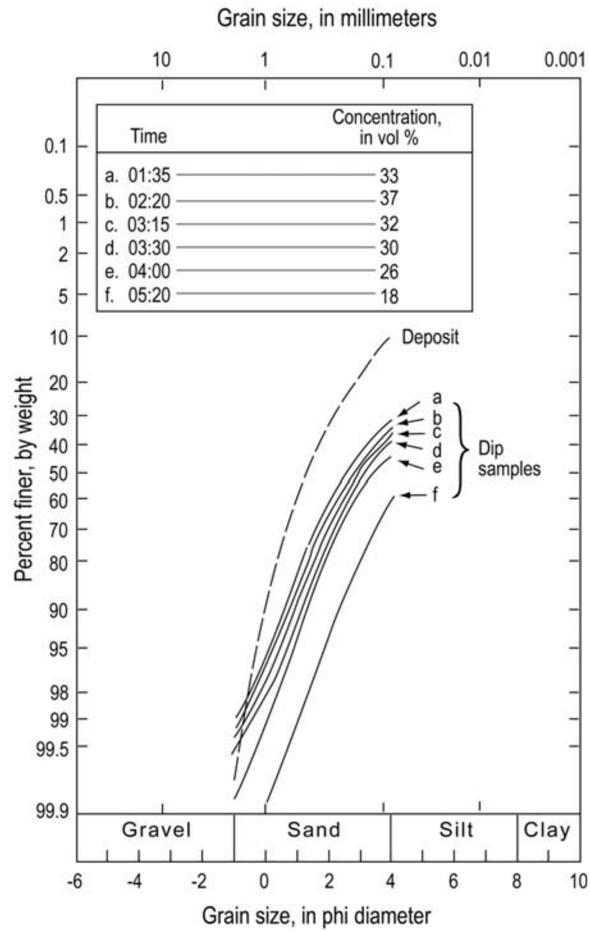

Figure 8.13. Differences in grain-size distribution between samples of hyperconcentrated flow (dip samples) collected 81 km downstream from source and the sediment deposited at the same site during the 19–20 March, 1982 hyperconcentrated lahar on the Toutle River. Dip samples were collected at different times during passage of the flow; corresponding sediment concentrations are shown.

From Pierson and Scott (1985).

et al., 2000), whereas fluvial deposits laid down by low-concentration water floods typically exhibit abrupt changes in mean grain size and stratification over relatively short distances.

Thalwegs tend to straighten during hyperconcentrated flow, moving away from the outsides of bends in sinuous reaches as a result of the formation of massive lateral berms (Zhou et al., 1983; Wan and Wang, 1994). Such berm formation tends to narrow the flow cross section, which may lead to bed degradation in the thalweg (Wan and Wang, 1994). Dinehart (1999) described such berm formation on the outside of a sweeping channel bend, where the berm formed a shallow beach (deposition occurring just below the water surface) that extended as far as 30 m out from the original bank. During more dilute flow recession, large parts of this berm were re-eroded.

8.6 HAZARDS

Hyperconcentrated flows at high discharge present significant hazards in addition to those of normal water floods and different from those of debris flows or mudflows. These hazards are commonly exhibited in watersheds where prodigious quantities of loose, erodible sediment become available for transport – in active volcanic areas, where eruptions can deposit large volumes of erodible material over broad areas and destroy vegetation cover (Major et al., 1996; Pierson et al., 1996; Rodolfo et al., 1996; Scott et al., 1996), and in mountainous areas subjected to wildfire, where burning of vegetation cover leaves soils loose and extremely vulnerable to sheetwash and rill erosion (Meyer and Wells, 1997; Cannon, 2001; see also Chapter 15).

Hyperconcentrated flows can be highly erosive, especially where channels are relatively steep (Waldron, 1967; Wang, 1990; Rickenmann, 1991; Xu, 1999, 2002a), but degree of scour is also a function of sediment concentration (Xu, 2002a). Streamflow tends to become more erosive when it transitions into the hyperconcentrated range (Xu, 2002a), but after the concentration limit (a function of grain-size distribution) has been reached for a flow, erosivity begins to decline (Wang, 1990). Debris flows and mudflows are comparatively less erosive, although the low-discharge "tails" of debris flows commonly transform to hyperconcentrated flow and can accomplish more erosion than the main bodies of the debris flows themselves (Pierson, 1986). Tens of metres of vertical scour have been observed in hyperconcentrated flow in China within a period of 10 hours (Kuang et al., 1999).

Hyperconcentrated flows are not always erosive, however. They can cause rapid deposition and river bed aggradation at places where channel gradients decrease or channels widen (Pierson et al., 1996; Rodolfo, 1996; Scott et al., 1996). With time and repeated flood events, deposition by hyperconcentrated flows can lead to incremental infilling of channels and channel shifting, reduction of flood-conveyance capacity, and burial of low-lying areas and structures in sediment (Figure 8.14). Hazardous river bed aggradation was well demonstrated along the rivers draining Mount Pinatubo (Philippines) following the large 1991 eruption of that volcano. Up to 25 m of channel aggradation occurred in rivers within the first three months following the June 1991 eruptions, and more has occurred since, due to deposition by thousands of rainfall-generated lahars (some debris flows but mostly hyperconcentrated flows) (Major et al., 1996; Pierson et al., 1996; Scott et al., 1996; Rodolfo et al., 1996). This activity has led to widespread burial of towns, roads, farms, and prime agricultural land by sediment.

Rapid lateral migration of channels is also common in low-gradient channels during aggradation by hyperconcentrated flows. This can lead to extreme rates and amounts of lateral erosion of river banks and destruction of buildings, roads, and bridges located on floodplains or alluvial fans (Major et al., 1996; Rodolfo et al.,

Figure 8.14. Burial of house and trees downstream of Mount Pinatubo, Philippines, due to river bed aggradation resulting primarily from repeated, relatively small hyperconcentrated flows that spread out on the alluvial fan after the stream channel was filled with sediment.

1996; see also Figure 8.15). Laterally eroded channels are typically rectangular in cross section, due to undercutting and the formation of near-vertical banks in unconsolidated alluvial fill. Width/depth ratios of channels cut in alluvium can be correlated to suspended sediment concentration; the higher the concentration, the deeper the channel relative to its width (Xu, 1999). Vigorous lateral erosion of unconsolidated stream banks with rates as high as 3 m/min (perpendicular to channel) has been documented at Mount Pinatubo accompanying bed aggradation (Rodolfo et al., 1996). About 40 km downstream from flow source, hundreds of buildings in Angeles City and long sections of flood-control levees were lost due to undercutting and collapse by bank erosion in the first few years following the eruption.

8.7 DISCUSSION

From the variety of experimental and field studies that have been carried out with hyperconcentrated mixtures of sediment and water, hyperconcentrated flow should be considered a distinct flow process on the basis of the following criteria:

1. Concentration of suspended fines is sufficient to impart yield strength to the fluid and maintain high fluid viscosity.

Figure 8.15. Destruction of buildings along a river bank downstream from Mount Pinatubo, Philippines, caused by lateral erosion of channel banks by hyperconcentrated flows.
Photo by Jon Major (USGS).

2. Sand and fine gravel, its settling hindered by the fluid viscosity, is kept in prolonged suspension by turbulence and dynamic grain interactions.
3. Significant bedload transport occurs as a traction carpet.
4. Mean flow velocity is greater and quantity of sediment transported is much greater than for water flow at a similar depth and slope.

These flow characteristics are different from those of either water flow or debris flow and mudflow. Yet many authors use only suspended sediment concentration alone as the defining criterion to identify hyperconcentrated flow, based on the concentration limits set by Beverage and Culbertson (1964). The fatal flaw in this approach is that grain-size distribution and grain density also play extremely important roles in determining the properties of sediment–water suspensions (Cao and Qian, 1990; Wan and Wang, 1994; Xu, 2002b, 2003). The widely used concentration thresholds of 20 vol% and 60 vol% (Beverage and Culbertson, 1964) were somewhat arbitrarily defined and can only be valid for sediment mixtures similar to the ones studied by Beverage and Culbertson.

The importance of grain-size distribution and grain density has been demonstrated experimentally with the use of artificial sediment mixtures and neutrally buoyant particles. For example, fines-free mixtures have Newtonian fluid properties (i.e., develop no yield strength) up to concentrations as high as 35 vol% for poorly sorted mixtures (Fei, 1983) and up to 50 vol% for uniformly sized coarse particles (Howard, 1965). A pure smectite clay suspension, on the other hand, acquires yield strength at concentrations of only about 1 vol% (Hampton, 1975). These values deviate greatly from Beverage and Culbertson's (1964) 20 vol% threshold, as do values for some natural flows. For example, poorly sorted (well-graded), clay-poor sediment mixtures from fresh volcanic terrains will transition from normal streamflow to hyperconcentrated flow when suspended sediment concentrations reach 10–11 vol% and will remain hyperconcentrated up to 53–54 vol%, whereupon they transition to debris flow (Pierson, 1986; Cronin et al., 1997, 1999, 2000; Dinehart, 1999). Flows in the Yellow River of China that have finer, better sorted sediment will transition from water flow to hyperconcentrated flow at a similar total suspended sediment concentration of 8–11 vol%, but they take on mudflow characteristics at much lower concentrations (19–37 vol%) (Qian et al., 1981; Wan and Wang, 1994; Xu, 1999, 2002b). Thus, if total suspended sediment concentration is used to characterize a flow, the sediment size distribution and density of particles in the flow must be stipulated and flow properties must be noted. The ranges in limiting thresholds for experimental and natural hyperconcentrated flows (Figure 8.16) show that the ranges involving mostly fine sediments can be entirely below Beverage and Culbertson's lower limit, while all of them start lower and end lower than the commonly used boundaries.

Hyperconcentrated flow should therefore be defined as a two-phase flow of water and sediment, intermediate in concentration between normal streamflow and debris flow (or mudflow), in which a viscous and yield-strength-maintained suspension of fines in water (the carrier fluid) enables the intermittent, dynamic suspension of large quantities of coarser sediment. Hyperconcentrated flow should not be defined on total suspended sediment concentration alone. One diagnostic test to make in the field would be to take a dip sample at the flow surface (well above the bed) with a bucket of similar container. The flow is hyperconcetrated if sand (and possibly fine gravel) was in suspension at the flow surface but settles out of suspension in the container within seconds. This would indicate that the coarse sediment was in dynamic suspension. The deposit in the bottom of the container should be normally graded, indicating that the coarses grains moved independently of the finer grains and settled out first.

8.8 CONCLUSIONS

Hyperconcentrated flow is a type of two-phase, non-Newtonian flow of sediment and water that operates between normal streamflow (water

flow) and debris flow (or mudflow). It is distinctive in terms of processes acting to transport the sediment. Laboratory and field evidence indicate that the transition to hyperconcentrated flow occurs when the concentration of suspended fines achieves a minimum volumetric concentration of 3–10%, depending on grain-size

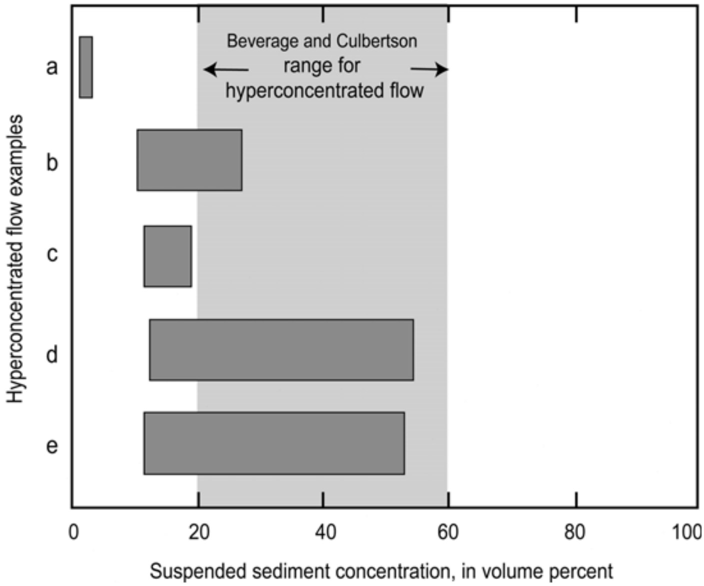

Figure 8.16. Approximate ranges in total suspended sediment concentrations for natural hyperconcentrated flows and laboratory-mixed hyperconcentrated suspensions of varying grain-size distributions and sediment compositions. Limits based on objective criteria discussed below. The shaded area is the widely quoted, arbitrarily defined concentration range assigned to hyperconcentrated flow by Beverage and Culbertson (1964). (a) Sheared experimental mixture of smectite clay and distilled water – concentration range 1–3 vol% (Hampton, 1975), where lower threshold is defined by the onset of yield strength and upper threshold by computation of competence (buoyancy and yield strength) to suspend a 4-mm mineral grain in a mixture. (b) Sheared experimental mixture of kaolinite clay and distilled water – concentration range 10–27 vol% (Hampton, 1975), with the same defining criteria as (a). (c) Clay, silt, and fine sand mixtures from the Yellow River basin, China – concentration range 11–19 vol%. Lower threshold is the concentration value widely accepted by Chinese authors defining the onset of hyperconcentrated flow (criteria not specified). The upper threshold is concentration at which sand grains can no longer settle out and mixture begins moving "as a whole" (Qian et al., 1981). (d) Poorly sorted volcaniclastic sediment (Ruapehu volcano, New Zealand) – concentration range 12–54 vol%, having fractions of 20–50 wt% fines, 50–70 wt% sand, and 0–15 wt% fine gravel (Cronin et al., 1997, 1999, 2000). Threshold criteria not defined. (e) Poorly sorted volcaniclastic sediment (Mount St. Helens, USA) – concentration range 11–53 vol% having fractions of 30–60 wt% fines, 30–70 wt% sand, and 0–10 wt% fine gravel (Pierson and Scott, 1985). Lower threshold based on concentration at which sand starts being suspended in significant quantities (Dinehard, 1999). Upper threshold based on observation of sustained suspension of pebbles and cobbles in moving flows (Pierson, 1986).

distribution, and begins to acquire yield strength. The fines mixture (sometimes referred to as the carrier fluid) then becomes able to transport prodigious quantities of coarser solid particles (sand and some gravel) in suspension. This coarse suspended load is held in prolonged dynamic suspension by turbulence, grain collisions, increased buoyancy, and increased viscosity (decreased fall velocity). However, the coarse suspended load will be selectively deposited during flow as flow velocity decreases.

Both water flow and debris flow/mudflow transport sediment differently from hyperconcentrated flow. In water flow, water has sufficiently low suspended sediment concentrations to behave as a Newtonian fluid, and sand particles are transported primarily as bed load. Large water floods carry some sand in intermittent suspension, but generally the suspended sand is mostly fine-grained, its vertical concentration profile is largely non-uniform, and the sand concentration in the flow is generally less than the fines concentration. Debris flows and mudflows, at the other end of the spectrum, are highly concentrated, relatively homogeneous slurries of sediment and water that behave as non-Newtonian, pseudo-one-phase flows. Because of dense grain packing, debris flows or mudflows cannot selectively deposit transported solid particles by size when velocity decreases or flow stops. This results in massive (non-stratified) and poorly sorted sediment textures that are characteristic of debris-flow and mudflow deposits.

It is not possible to determine whether a flow is hyperconcentrated from concentration values alone. This is because grain-size distributions and grain densities of the transported sediment control the physical properties of sediment–water mixtures, and thus also control the threshold concentrations at which flow transformations occur from water flow to hyperconcentrated flow and from hyperconcentrated flow to debris flow or mudflow. The fundamental characteristic that defines hyperconcentrated flow is the transport of large quantities of coarse sediment (sand and possibly some gravel) at high concentrations *in intermittent dynamic suspension*.

Hyperconcentrated flows generally are not as hazardous as debris flows, because their velocities are usually lower and they tend not to transport the large boulders that are responsible for impact damage in debris flows. However, high-discharge hyperconcentrated flows present a greater hazard to riverside communities than normal floods of similar magnitude, because of their greater potential for doing geomorphic work. In relatively steep channels, hyperconcentrated flows can rapidly incise their channels for tens of metres, undermining bridge piers and other channel structures, such as erosion-control dams. In channels with gentle slopes, deposition of their high sediment loads can (a) cause lateral shifting of active channels and vigorous bank erosion, causing collapse of riverside buildings, bridges, and flood-protection levees, and (b) incrementally fill river valleys with deposits of sand and gravel, burying low-lying areas in sediment and removing channel capacity that can exacerbate later flooding.

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