5 SEDIMENT CONCENTRATION

'The concentration is so high that eroded sediment can be easily carried away by the flow' Zhaohui Wan & Zhaoyin Wang, 1994

5.1 Introduction

Large sediment concentrations in runoff might significantly alter fluid properties and flow behaviour. Fluid density, settling velocity, viscosity, flow velocity and transport capacity might all change. Such changes are generally not considered in present day soil erosion models. Sediment concentrations in runoff on the Loess Plateau are among the highest on earth. The Yellow River even derives its name from the transported loess and is rightly called the world's muddiest river (Douglas, 1989). Therefore, if erosion models are to be applied to Loess Plateau conditions the effects of high concentrations must be considered. Sediment concentrations on the Loess Plateau increase with increasing discharge to a certain limit and remain constant after that limit has been reached (Gong Shiyang & Jiang Deqi, 1979). According to their data, the 'stable concentration' is about 800 g/l. They studied catchments ranging in size from 0.49 to 3,890 km^2 and found that in small catchments the stable concentration is reached at lower discharge than in large catchments. Other authors, however, report concentrations of 1000 g/l (Jiang Deqi et al, 1981, Zhang et al, 1990, Zhaohui Wan & Zhaoyin Wang, 1994) and even 1600 g/l (Long Yuqian & Xiong Guishu, 1981) and 1700 g/l have been reported (Zhaohui Wan & Zhaoyin Wang, 1994) for river flow in Yellow River tributaries.

Bradley & McCutcheon (1987) gave an overview of the effects of high suspended sediment concentrations in rivers. They showed that different authors have classified flow in different ways as a function of sediment content. A useful classification is that used by Scott (1988) and Costa (1988). They distinguished normal streamflow, hyperconcentrated streamflow and debris flow. Table 5.1 shows some characteristics of these different types of flow. In nature, a continuum of flow conditions and concentrations occurs, so that changes from one type of flow to another can be gradual. Each flow type, however, has its own specific characteristics and processes.

Normal stream-flow is a Newtonian fluid. In a Newtonian fluid the shear stress is given by:

$$\tau = \mu \cdot \frac{du}{dy} \tag{5.1}$$

Where: τ = shear stress
μ = dynamic viscosity
u = velocity
y = level above bed
du/dy = shear rate

Hence, for Newtonian fluids a chart of shear stress as function of shear rate will be a straight line passing through the origin. Turbulence is probably the most important process in supporting the sediment in the flow.

Table 5.1 Characteristics of different types of flow (based on Costa, 1988)

	Normal Streamflow	Hyperconcentrated Flow	Debris Flow
Fluid density (kg/m^3)	1010-1330	1330-1800	1800-2300
Dirty water concentration (g/l)	16-530	530-1285	1285-2088
Fluid type	Newtonian	non-Newtonian? (likely Bingham)	Visco-plastic?
Flow type	turbulent	turbulent/laminar	laminar
Sediment support mechanism	electrostatic forces turbulence	buoyancy dispersive stress turbulence	cohesion buoyancy dispersive stress structural support

For flow that contains large amounts of sediment the flow might transform in a Bingham fluid. For Bingham fluids the shear stress can be given by (Costa, 1988; Selby, 1993; Zhaohui wan & Zhaoyin Wang, 1994):

$$\tau = \tau_b + \mu \cdot \frac{du}{dy} \tag{5.2}$$

Where: τ_b = yield stress

Hence, for Bingham fluids a chart of shear stress as function of shear rate will be a straight line with intercept τ_b on the shear stress axis. In other words, a certain amount of stress can be exerted without any resulting strain rate. The existence of yield stress is one of the factors that can help explain why the behaviour of hyperconcentrated flows is different from that of normal streamflow. Yield stress increases with increasing sediment concentrations. Sediment in the flow is mainly supported by buoyancy, dispersive stress and turbulence. Hyperconcentrated flows are turbulent, solid-liquid two-phase flows (Xu Jiongxin, 1999a,b). The fluid phase is formed by water with the sediment particles below 0.01 mm uniformly distributed within it. The solid phase is formed by large (larger than 0.05 mm) suspended particles.

At very high sediment concentrations, flows might transform into debris flows. At such concentrations the flow has large yield stress (or cohesion) and also internal friction.

According to Costa (1988) the shear stress for such flow may be calculated with a Coulomb-viscous model:

$$\tau = c + \sigma \cdot \tan\varphi + \mu \cdot \frac{du}{dy} \qquad (5.3)$$

Where: c = cohesion
σ = normal stress
φ = angle of internal friction

For debris flows turbulence is usually greatly suppressed and the most important sediment supporting processes are buoyancy, dispersive stress, structural support and cohesion. Solids and water move together as a single viscoplastic body from which there is hardly any sedimentation (Selby, 1993).

Scott (1988) placed the boundaries between these three types of flow at dirty water concentrations of 530 and 1590 g/l respectively. Other authors (e.g. Xu Jiongxin, 1999b) placed the boundary between ‘normal’ flow and hyperconcentrated flow at the transition from Newtonian fluid to Non-Newtonian (usually Bingham) fluid. According to Xu Jiongxin this boundary is at about 300 to 400 g/l. Many of the floods on the Loess Plateau have concentrations above 400 g/l and can therefore be called hyperconcentrated flows.

Despite the different concentrations that different authors used to distinguish between the different flow types it is clear that hyperconcentrated flow occurs regularly on the Chinese Loess Plateau. Debris flows, however, are rare. Nevertheless, the high concentrations encountered in hyperconcentrated flow can have large influence on fluid properties, flow behaviour and transport capacity of the flow. The aims of this chapter are:

- To find out what the effects of very high sediment concentrations are on fluid properties and flow behaviour.
- To determine if these effects require adaptations in process based erosion models, and if so, what kinds of changes are needed.
- To find out what concentration-related corrections are necessary to compare simulation results with measured values of discharge and sediment concentration in the Danangou catchment.

5.2 Causes of high concentrations

5.2.1 Steep slopes with loose materials

The concentrations in runoff on the Loess Plateau are exceptionally high. Such high concentrations have not been reported from other loess areas in the world. Instead, these concentrations are comparable to those reported from some badland areas and from lahars in volcanic regions. During extreme rainfall events in mountainous regions high concentrations can also be reached (e.g. Batalla et al., 1999).

Olivier & Pebay Peyroula (1995) and Mathys (1995) reported sediment concentrations of about 500 g/l for the Terres Noires marles near Draix, southern France. These concentrations were measured after the flow passed a sedimentation pool, so that concentrations before the pool must have been higher. Mudflows with concentrations of 1500 g/l where observed in the same region. Cantón et al. (2001) reported maximum concentrations of 800 g/l for the Tabernas badlands in southern Spain.

Scott (1988) reported concentrations in lahar-runout flows of over 1000 g/l. He showed that lahars (volcanic debris flows) can be formed rapidly from normal streamflow on the steep slopes of Mount St. Helens. These steep slopes are underlain by fragmental pyroclastic debris. Further downstream, such lahars can transform to lahar-runout flows (hyperconcentrated streamflow) because of dilution by clearer water. He also found that fine-material load in hyperconcentrated flows can be highly persistent.

In the case of both badlands and lahars, erodible materials are present on steep slopes. The presence of erodible loess on the steep slopes of the Loess Plateau might therefore be one of the most important causes for the high concentrations. Slope angles in other loess regions in the world are generally less. Steep slope angles mean that the water will have high energy, since the flow of water is driven by the potential energy. Further, for loose material, the slope angle might be close to the angle of internal friction, so that such material will already almost move under the influence of gravity alone. There are indications that though steep slopes might be needed to initiate hyperconcentrated flow they are not needed to maintain this type of flow. This is due to certain feedback mechanisms that operate in these kinds of flow. These mechanisms will be discussed in chapter 5.3.

5.2.2 Loess characteristics

Another explanation for the very high concentrations observed on the Chinese Loess Plateau could be that the loess of the Loess Plateau differs from loess elsewhere. The Plateau is located in an area with a pronounced semi-arid climate. As a result there is not much water available for weathering of the loess. Table 4.10 showed that even the upper loess layers in the Danangou catchment are still very calcareous, which demonstrates that the loess is hardly weathered. Unweathered loess has a very open structure in which the silt particles are bonded to each other by calcium, soluble salts and clay minerals (Tan, 1988). These soluble salts and clay minerals are very sensitive to changes in water content, so that wetting might result in collapse of the loess structure. Furthermore, Billard et al. (2000) reported that the dissolution of soluble salts can give rise to very basic pH values, which promote the dispersion of aggregates. They reported maximum pH values of 9.1-9.3 from the western part of the Loess Plateau (Gansu Province). Messing et al. (in press a) showed that soil pH in the Danangou catchment (Shaanxi Province) is generally above 8.5. The major loess deposits of Europe and North America, however, are mostly located in temperate climates. As a result, the loess in those regions is usually weathered and decalcified in the upper part (Table 4.10), while pH is also much lower. This could explain differences in behaviour between Chinese loess and e.g. European Loess. For ploughed soils, such structural differences are probably less important than for undisturbed soils. Still, for ploughed soils these differences in structure might cause a more rapid disintegration of aggregates in the case of the Chinese Loess Plateau.

5.2.3 Climate

The harsh climatic conditions on the Loess Plateau result in poor vegetation covers, so that the soil is not well protected. The occurrence of heavy rainstorms in summer might therefore also be an important factor in causing the development of hyperconcentrated flow. Horton (1945) mentioned two factors that might explain why this can be especially important in semiarid areas. First, in semiarid areas the soils are likely to be very dry when a storm occurs. Such soils are more susceptible to erosion because capillary forces in the soil are weak and because very dry aggregates might explode when suddenly wetted. Second, high intensity storms (characteristic of semiarid areas) tend to produce the highest rainfall intensities early on in the storm, so that the soil is still dry when this happens. According to Horton, the soil might be beaten into a semifluid mass because of this.

From these three possible causes for very high sediment concentrations in runoff, the presence of steep slopes is probably the most important one, but loess characteristics and climate are likely to play a role too. A combination of these factors therefore seems the most likely cause of the very high sediment concentrations on the Chinese Loess Plateau.

5.3 Consequences of high concentrations

High sediment concentrations can have multiple effects on the behaviour of flow and its sediment transport capacity. These effects cannot really be separated since they occur simultaneously, but for the sake of clarity they will be discussed one by one.

5.3.1 Fluid density

Fluid density increases markedly with increasing sediment concentrations. The density of fluids with different concentrations can be calculated with:

$$\rho_f = \left(1 - \frac{C_f}{\rho_s}\right) \cdot \rho_w + C_f \qquad (5.4)$$

Where: ρ_f = density of fluid (kg/m^3)
ρ_s = density of solid (kg/m^3)
ρ_w = density of clear water (kg/m^3)
C_f = dirty water concentration (g/l)

Assuming that the density of water is 1000 kg/m^3 and the density of sediment is 2650 kg/m^3 a concentration of 1000g/l will result in density of 1623 kg/m^3. Such high-density flows have larger potential energy and larger momentum than clear water flow. If all other properties of the fluid would remain the same, this should result in an increase in flow velocity in comparison to clear water flow. In addition, the shear stress exerted by the flow will be larger.

5.3.2 Viscosity

For clear water, viscosity is a function of temperature only. According to Van Rijn (1993) dynamic viscosity can be approximated by:

$$\mu_0 = \frac{\rho_w \cdot (1.14 - 0.031 \cdot (T - 15) + 0.00068 \cdot (T - 15)^2)}{10^6} \tag{5.5}$$

Where: μ_0 = clear water dynamic viscosity (Ns/m^2)
ρ_w = clear water density (kg/m^3)
T = temperature (^{o}C)

According to equation 5.5 the viscosity of clear water of 15^{o}C will be 1.14*10^{-3} Ns m^{-2}. Viscosity of a fluid will increase with increasing sediment concentration. Many authors have developed equations to calculate viscosity from volumetric sediment concentration. Several equations calculate viscosity from volumetric sediment content alone, but some authors (e.g. Zhaohui Wan & Zhaoyin Wang, 1994) showed that clay particles have more influence than other particles, so that both grain-size distribution and clay mineralogy should be taken into account as well. Van Rijn (1993), Hsieh Wen Shen & Julien (1993) and Zhaohui Wan & Zhaoyin Wang (1994) all reported empirical equations to calculate viscosity for sediment-laden flows. Some of the equations are reproduced here:

Bagnold, 1954

$$\frac{\mu}{\mu_0} = (1 + p) \cdot (1 + 0.5p)$$

$$p = \frac{1}{\left(0.74 / C_{vf}\right)^{0.33} - 1} \tag{5.6}$$

Do Ik Lee, 1969

$$\frac{\mu}{\mu_0} = (1 - C_{vf})^{-(2.5 + 1.9 \cdot C_{vf} + 7.7 \cdot C_{vf}^2)} \tag{5.7}$$

Krone, 1963

$$\frac{\mu}{\mu_0} = e^{(2.5 \cdot C_{vf})} \tag{5.8}$$

Fei Xiangjun, 1982

$$\frac{\mu}{\mu_0} = \left(1 - 1.35 \cdot C_{vf}\right)^{-2.5} \tag{5.9}$$

Moliboxino, 1956

$$\frac{\mu}{\mu_0} = 1 + \frac{3}{\frac{1}{C_{vf}} - \frac{1}{0.52}} \tag{5.10}$$

Where: μ = dynamic fluid viscosity (Ns/m^2)
μ_0 = dynamic viscosity of clear water (Ns/m^2)
C_{vf} = volumetric dirty water concentration (-)
p = concentration parameter Bagnold equation

Figure 5.1 shows viscosities calculated with these different equations for different sediment concentrations. The viscosity is expressed as the fluid viscosity divided by the clear water viscosity. Viscosities calculated with the Bagnold equation are much higher than those calculated with the other equations. The other equations give more or less the same result, except for very high concentrations, where the Krone equation starts to deviate. For concentrations of 1000 g/l viscosity is about 5 times higher than for clear water. For concentrations of 400 g/l, which on the Loess Plateau would be called moderate, the increase in viscosity is about 60%. If all other fluid properties were to remain constant an increase in viscosity should result in a decrease of flow velocity. Zhaohui Wan & Zhaoyin Wang (1994) even reported that flow in some of the Yellow River tributaries actually stops sometimes because of the increase in viscosity.

5.3.3 Settling velocity

Traditionally settling velocity is calculated with the Stokes equation, which can be written as:

$$\omega = \frac{2gr^2\left(\rho_s - \rho_f\right)}{9\mu} \tag{5.11}$$

Where: ω = settling velocity (m/s)
g = gravitational acceleration (m/s^2)
r = grain radius (m)
ρ_f = density of fluid (kg/m^3)
ρ_s = density of solid (kg/m^3)
μ = dynamic fluid viscosity (Ns/m^2)

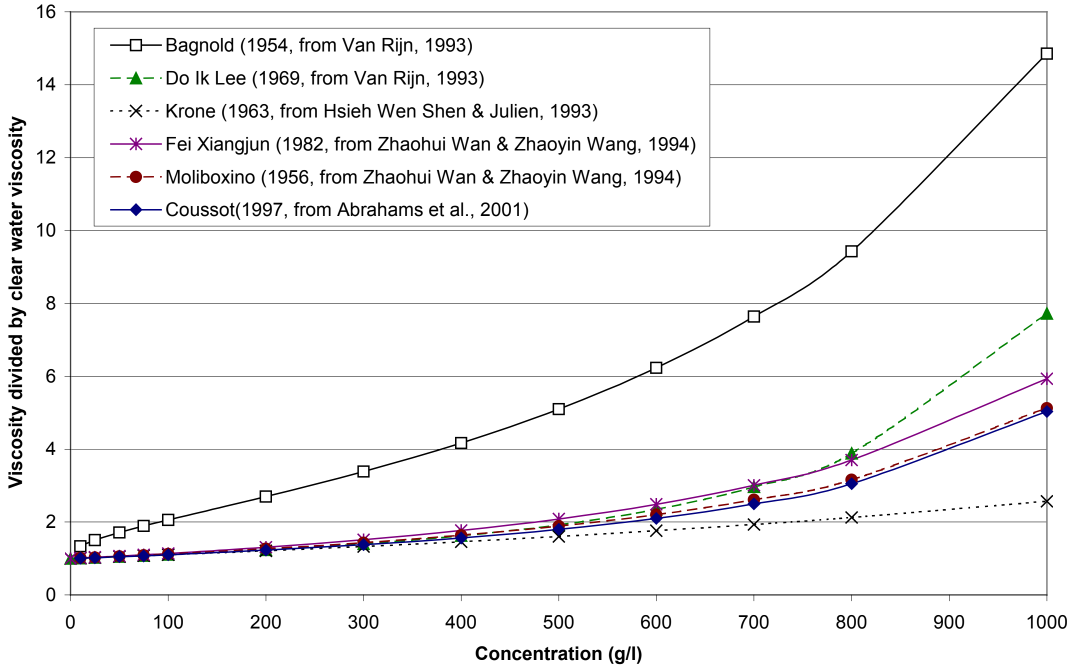

Figure 5.1 Dynamic viscosity as function of sediment concentration. Viscosity is expressed as the fluid viscosity divided by clear water viscosity. Values are for 15 degree centigrade fluid. Note that for the Bagnold and Moliboxino equations no solution is possible for zero concentration.

A for the Loess Plateau typical grainsize of 35 mu than has a settling velocity in clear water of about 1 mm/s. For water containing sediment ρ_f can be calculated with equation 5.4 and μ with equations 5.6 – 5.10. In this way the effects of decreased submerged weight and increased viscosity can be incorporated in the Stokes equation. Increasing sediment concentrations, however, have more effects on settling velocity. With an increase in concentration settling velocity will decrease due to several effects (Zhaohui Wan & Zhaoyin Wang, 1994):

- The downward movement of particles will induce an upward movement of water, which causes a drag force on the particles
- The submerged weight of the particle decreases since the density of the fluid increases.
- The viscosity increases.
- If the fluid has become a Bingham fluid there will be yield stress.
- There is interference between the settling particles
- When there is enough clay in suspension flocculation occurs. In extremis the clay particles can form a flocculent structure that prevents the coarser particles from settling as well. Instead the settling proceeds at an extremely low pace and should be regarded as a consolidation process. Turbulence might (partially) destroy the flocculent structure, so that some particles might not settle in standing water, but will settle in flowing water.

The overall result is that for hyperconcentrated flow there is practically no settling of sediment (Gong Shiyang & Jiang Deqi, 1979, Long Yuqian & Xiong Guishu, 1981, Xu Jiongxin, 1999a,b).

Many authors have developed equations to calculate settling velocity from volumetric sediment concentration. Van Rijn (1993), Hsieh Wen Shen & Julien (1993) and Zhaohui Wan & Zhaoyin Wang (1994) all reported equations to calculate settling velocity. Some of the equations are reproduced here:

Wan & Sheng, 1978

$$\frac{\omega}{\omega_0} = \frac{\left(1 - C_{vf}\right)^2}{1 + \frac{3}{\frac{1}{C_{vf}} - \frac{1}{0.52}}} \tag{5.12}$$

Hawksley, 1951

$$\frac{\omega}{\omega_0} = \xi(1 - C_{vf})^2 e(\frac{-k_1 C_{vf}}{1 - k_2 C_{vf}}) \tag{5.13}$$

Oliver, 1961

$$\frac{\omega}{\omega_0} = (1 - 2.15 \cdot C_{vf}) \cdot (1 - 0.75 \cdot C_{vf}^{0.33}) \tag{5.14}$$

Chien & Wan, 1983

$$\frac{\omega}{\omega_0} = (1 - C_{vf})^b \tag{5.15}$$

Where:
- ω = settling velocity in fluid (m/s)
- ω_0 = settling velocity in clear water (m/s)
- C_{vf} = volumetric dirty water concentration
- ξ = 1 without flocculation, 2/3 with flocculation
- k_1 = 5/2 for spheres
- k_2 = 39/64
- b = coefficient between 2.35 and 4.65

The Wan & Sheng equation makes use of the Moliboxino equation for viscosity.

Figure 5.2 shows settling velocities calculated with different equations for different sediment concentrations. The settling velocity is expressed as fraction of what the settling velocity would be in clear water. For the Hawksley equation, it was assumed that there was no flocculation, while for the Chien & Wan equation 4.65 was used as exponent. The figure shows that even for moderate concentrations (in Loess Plateau terms) of 400 g/l, and under the assumption that there is not enough clay to form a significant flocculation structure, settling velocity already decreases to some 40%-50% of its clear water value. All equations give comparable results, only the Oliver equation deviates somewhat for low concentrations.

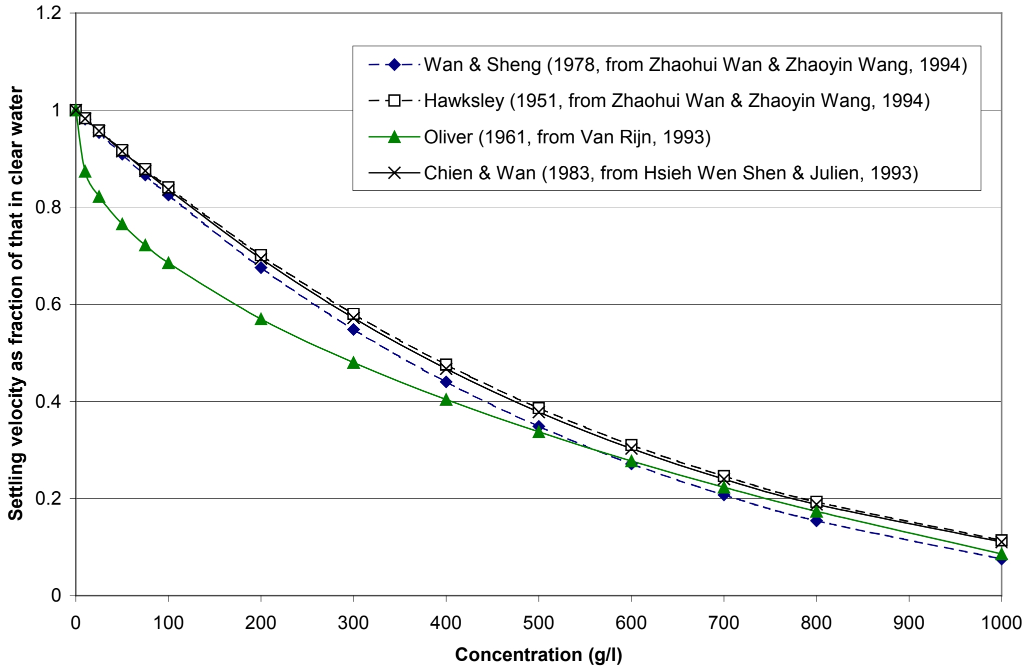

Figure 5.2 Settling velocity as a function of sediment concentration. Settling velocity is expressed as fraction of the clear water settling velocity. For the Wan & Sheng equation no solution is possible for zero concentration.

The effect of very low settling velocities should be that once sediment has been entrained by the flow there would be hardly any sedimentation out of the flow. This can be expected to result in an increase of transport rate and sediment yield. The energy needed to support the suspended sediment load is provided by turbulence. This means that the turbulence will decrease with increasing sediment load. More energy is thus used for sediment transport and less for turbulence. The net energy loss is small.

When the fluid is a Bingham fluid it has yield stress. Suspended particles exert a stress on the fluid because of gravity. If this stress is below the yield stress of the fluid the sediment will not settle at all. This load is called the neutrally buoyant load.

5.3.4 Transport capacity

Xu Jiongxin (1999a) showed how, for hyperconcentrated flows, the transport capacity increases with increasing concentration. As the sediment concentration increases the fluid density increases. This results in a lower submerged density of the particles. Less energy is therefore needed to maintain this concentration and energy will be available to entrain more sediment. Zhaohui Wan & Zhaoyin Wang (1994) combined data from different Chinese sources and found that for high concentrations (above about 200 g/l) more sediment can be carried by flows of weaker intensity. This can be attributed to a decrease in the settling velocity.

This shows that a positive-feedback mechanism is operating. Because of this feedback, there is a positive relationship between sediment concentration and suspended sediment size (expressed as D50, Xu Jiongxin 1999b, Gong Shiyang & Jiang Deqi, 1979). Another result of this feedback is that concentrations are likely to increase in the downstream direction. Zhaohui Wan & Zhaoyin Wang (1994) showed for the Chaba ravine, Dali catchment, how maximum sediment concentrations in runoff increase in the downstream direction from about 700 g/l at plot level to about 1200 g/l for the main river. In the Yellow River itself the suspended sediment concentrations can be about 600-700 g/l during the flood season. These slightly lower values might be the result of mixing with clearer waters, for example with baseflow. Hyperconcentrated flows on the Loess Plateau only occur during the flood season.

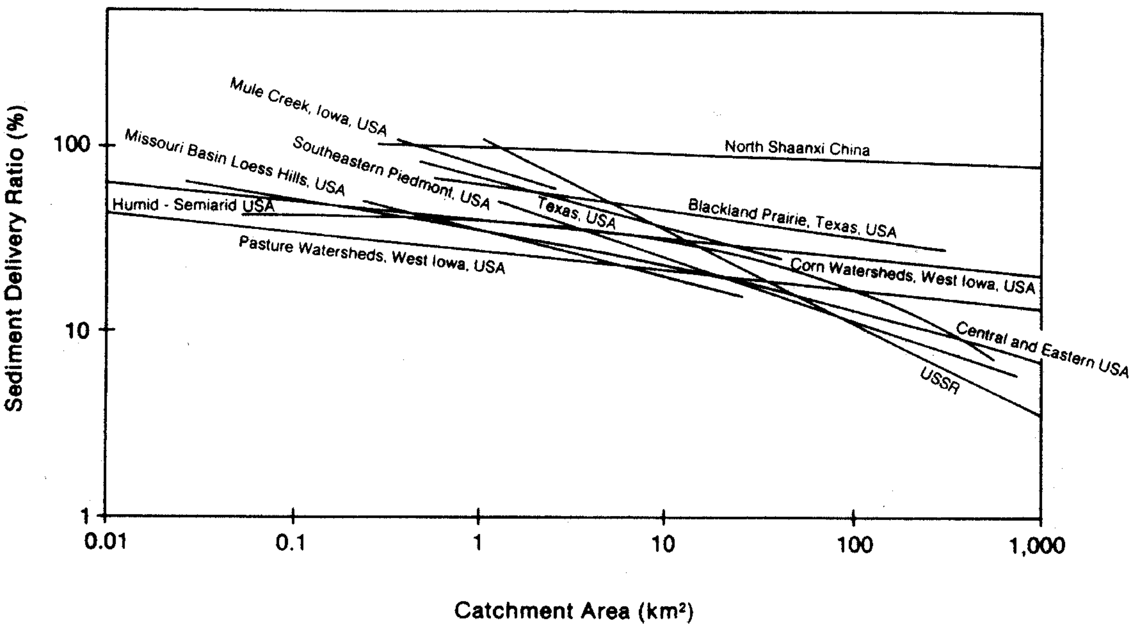

Figure 5.3 Sediment delivery ratio as a function of catchment size. Adapted from Graf (1988)

These high transport capacities result in very high sediment delivery ratios for Loess Plateau catchments. Figure 5.3 shows sediment delivery ratios for different regions as a function of catchment size. It shows that for the larger sized catchments the sediment delivery ratio for northern Shaanxi is much larger than for the other regions. According to Xu Jiongxin (1999a) the sediment delivery ratio is still almost 100% for areas as large as 10000 km^2. The lower order channels on the Loess Plateau are essentially sediment-transporting channels and under natural conditions there is very little opportunity for sedimentation.

5.3.5 Flow velocity and flow resistance

High sediment contents should also have an influence on water velocity. On the one hand one would expect velocity to decrease because of increased viscosity. On the other hand the added sediment will add momentum to the flow.

Govers (1990) found a significant increase in flow velocity with an increase in sediment content for overland flow for sands with d50 of 218 and 1098 μm. Flow with a volumetric sediment content of 0.32 had a velocity 40% higher than the clear water velocity. He attributed this to momentum added to the fluid by the sediment and to changes in the turbulence structure of the flow. According to Govers (1990) this phenomenon has also been long known to occur in rivers as well.

Einstein & Chien (1955) conducted a series of flume experiments with sands (median grainsize between 0.274 and 1.3 mm) and also found that average velocity increased with increasing sediment content. They suggested that this is the result of dampening of turbulence caused by the high concentrations. In their experiments, however, the sediment was concentrated in the lower part of the flow, while an increase in velocity was only found in the upper part of the flow. Their explanation, however, is probably valid, because turbulence will also be dampened in the upper part of the flow. Such dampened turbulence in the clear upper part of the flow might well result in higher velocity, but it might not give information about flows where high concentrations occur throughout the flow instead of just in the lower part.

Torri & Borselli (1991) showed that such an increase in velocity with increasing sediment content is only possible if less energy is dissipated in turbulence and friction. This means that flow resistance should decrease. Wan Zhaohui & Wang Zhaoyin (1994) discussed flow resistance and flow velocity for sediment-laden flows. According to them the flow resistance of sediment-laden flows consists of 3 parts:

- viscous resistance
- turbulent resistance
- resistance caused by bedload movement and bed configuration

They also distinguished between flows that carry only fines (pseudo one phase flow) and flows that carry fines as well as coarse material (sediment laden flow). For the flow carrying only fines, bedload is not important and resistance only consists of viscous resistance and turbulent resistance. As sediment content increases, viscous resistance increases, while turbulent resistance decreases. The net effect seems to depend on whether the bed is rough or smooth and whether the flow is laminar or turbulent. As a result, there need not be a decrease in resistance with increasing sediment content. According to Wan Zhaohui & Wang Zhaoyin (1994), there even usually is an increase in resistance with increasing concentration. In the case of sediment laden flow resistance caused by bedload transport and bed configuration can decrease with increasing sediment content if the higher concentration causes the flow to transport more coarse material as suspended load instead of bedload. The bed should then become smoother and the resistance would be less. Wan Zhaohui & Wang Zhaoyin (1994), however, did not discuss what happens if bed material is so large that it cannot be transported as suspended load. In that case the bed would not become smoother and resistance might not decrease. Wan Zhaohui & Wang Zhaoyin (1994) also stated that as long as the flow remains fully turbulent the resistance to flow will be the same for Newtonian and Bingham fluids.

Thus, the effect of high concentrations on resistance remains unclear. The effect on flow velocity is therefore likewise unclear.

According to Bradley & McCutcheon (1987) the Manning and Chezy equations are only applicable when it can be assumed that the velocity distribution over depth is log-linear. According to them available data on the velocity distribution in hyperconcentrated flows contradict each other and the use of Manning's equation under such circumstances is therefore doubtful. Wan Zhaohui & Wang Zhaoyin (1994) showed how the velocity profile in hyperconcentrated flows depends on the flow being laminar or turbulent. In laminar flow of a Bingham fluid the shear stress will be lower than the yield stress for the upper part of the flow. There is therefore no velocity gradient and plug flow is developed. On the other hand, in turbulent flow the velocity profile generally remains logarithmic, even though the Von Karman constant might be different than for clear water.

5.3.6 Discussion

From the preceding sections it is clear that large sediment concentrations in rivers may have considerable influence on a whole range of flow characteristics.

The effects can sometimes be unexpected and contradictory to accepted concepts in erosion modelling. For example, the observation that sometimes particles will not settle in standing water while they do settle in flowing water is unusual. Likewise, it was shown that transport capacity might increase with increasing sediment concentration. This obviously undermines the concept of transport capacity as normally used in erosion modelling. There, transport capacity is assumed to depend only on flow characteristics, while entrainment is usually modelled as a function of the difference between transport capacity and concentration.

The effects are also complex and hard to separate from each other. In addition, it seems likely that some effects will partly cancel each other out. For example, higher viscosity should decrease flow velocity, while higher density should increase it. Obviously, erosion models that want to deal with high concentrations should at least consider these effects.

5.4 Streamflow in the Danangou catchment

The maximum dirty water concentrations measured at the dam in the Danangou catchment were about 500 g/l (table 4.10), in the Yan River they were around 600 g/l. The concentrations are thus high, but in Loess Plateau terms not extremely high. This section will discuss how these high sediment contents were taken into account during processing of measurement data and what the consequences are for modelling soil erosion in the catchment.

5.4.1 Velocity and discharge

The high sediment concentrations in the Danangou catchment could influence the discharge coefficient for the weir. As described in chapter 4 discharge at the weir in the Danangou catchment was measured with an equation based on the law of Bernoulli. The resulting equation (equation 4.7) also contains a correction factor that, among other things, depends on viscosity. The correction factor supposedly does not depend on fluid density. In the Danangou catchment viscosity is likely to be the most important factor to make the use of a correction factor necessary as high sediment contents will increase viscosity. Since bed material is so coarse that it cannot be transported as suspended material there is no reason to suppose that resistance caused by bed material will decrease with increasing sediment concentration. Sediment content in this region can be as high as 1000g/l and this can change viscosity considerably. Figure 5.1 shows that for such sediment concentrations viscosity could be about 5 times higher than for clear water. Data about the relationship between viscosity and discharge coefficient are however hard to find.

It seems, nevertheless, prudent to take viscosity into consideration, as the sediment contents encountered on the Loess Plateau could well be outside the range normally considered in the determination of the discharge coefficient. Increasing viscosity should decrease velocity. This would result in a lower discharge coefficient. A discharge coefficient of 0.9 is therefore used instead of the 0.95 that was suggested in chapter 4. Introducing a sediment content (hence viscosity) dependent coefficient instead of a constant (0.9) would be preferable, but insufficient data about the relationship between viscosity and discharge coefficient were available for this.

During the event of July 20th, 1999, surface velocity at the weir was measured. Plastic bottles partially filled with sediment were thrown into the stream upstream of the weir. The sediment was needed to be able to throw the bottles far enough, and also it was hoped that by using the sediment the bottles would flow more upright and be better visible. It was measured how long it took the bottles to travel a distance of 40 metres. Observations showed that most of the bottles were held up for some time along the way. Therefore, the fastest bottles were assumed to be most representative of flow velocity. The measurements gave a surface flow velocity of about 2 m/s. Since water levels were known from the ultrasonic sensor, discharge can be estimated as the product of wetted area and average velocity. The method is not very accurate, and moreover, average velocity is not equal to surface velocity, but the data nevertheless indicated that the discharges obtained from equation 4.7 with a correction factor of 0.9 were about right. Hence, a further viscosity correction was apparently not needed for velocity.

5.4.2 Sediment content

As mentioned before sediment content of the discharge can become very high in the Danangou catchment. This not only influences viscosity, but also water level itself. Sediment contents up to 1000 g/l of fluid have been measured in the region. If a particle density of 2650 g/l is assumed this gives a sediment volume of 38%. Maximum concentrations measured at the dam are about 500 g/l. As this is suspended load, it can be expected that sediment velocity equals water velocity. In that case sediment volume can just be subtracted from fluid volume to give water volume. Note that this would not be possible if sediment velocity and water velocity are not equal (Govers, 1992a). Govers also summarized some results from studies on this subject, which where carried out for sheetflow. The quoted results were partially contradicting, which reflects the scantiness of our knowledge on this subject. Because of this Govers decided not to use a correction at all, so how to correct remains a question. For the discharge calculation in the Danangou catchment the discharge coefficient was first adapted (as described above) to calculate the fluid velocity and discharge. Calculated discharge was then corrected to clear water discharge by subtracting the sediment discharge. An alternative would be to first correct water level and then calculate discharge, as described by Steegen & Govers (2001). This approach was not used here because the discharge equation is highly sensitive to fluid level, so that using a corrected clear water level instead of the actual fluid level might well result in an underestimate of discharge. The following equation was used to correct discharge:

$$Q_w = \left(1 - \frac{C_f}{\rho_s}\right) \cdot Q_f \qquad (5.16)$$

Where: Q_w = clear water discharge (m^3/s)
Q_f = fluid discharge (m^3/s)
C_f = dirty water concentration (g/l)
ρ_s = particle density (2650 kg/m^3)

The effects of this correction are shown in figure 5.4 for the event of July 20th, 1999. This correction is necessary to be able to evaluate the relationship between precipitation and discharge for areas with high sediment contents and also because the results from soil erosion models are expressed as clear water discharges. These erosion models also express concentration as gram per litre of clear water. Since measured concentrations are expressed as gram per litre dirty water a correction is necessary. Sediment discharge can be calculated with:

$$Q_{sed} = C_f \cdot Q_f = C_w \cdot Q_w \qquad (5.17)$$

Where: Q_{sed} = sediment discharge (kg/s)
Corrected concentration can then be calculated with:

$$C_w = \frac{Q_{sed}}{Q_w} \tag{5.18}$$

or:

$$C_w = \frac{C_f}{(1 - \frac{C_f}{\rho_s})} \tag{5.19}$$

When the correction proposed in equations 5.16 and 5.19 are applied simulation results can be compared to field measurements.

When the ultrasonic sensor did not function the data from the pressure transducer had to be used (see chapter 4). In that case equation 5.16 should also be applied, but before this is possible the water level should be calculated from the pressure transducer signal. The output of the pressure transducer is a level that is based on the assumption that the density of the fluid is ρ_w (density of clear water), while in fact it is ρ_f (density of the fluid with sediment). To correct for this the pressure has to be calculated from the level given by the sensor using ρ_w. Then the 'fluid level' can be calculated with ρ_f. The procedure is as follows:

$$P = \rho_w g H_w \tag{5.20}$$

And also:

$$P = \rho_f g H_f \tag{5.21}$$

Hence,

$$H_f = \frac{\rho_w}{\rho_f} \cdot H_w \tag{5.22}$$

Where: P = pressure (N/m^2)
H_w = water level from sensor (m)
H_f = corrected level fluid (m)
ρ_w = density of clear water (kg/m^3)
ρ_f = density of fluid (kg/m^3, can be calculated with 5.4)

Then discharge can be calculated using $h_c = H_f$ in equation 4.7. Finally, equation 5.16 can be applied.

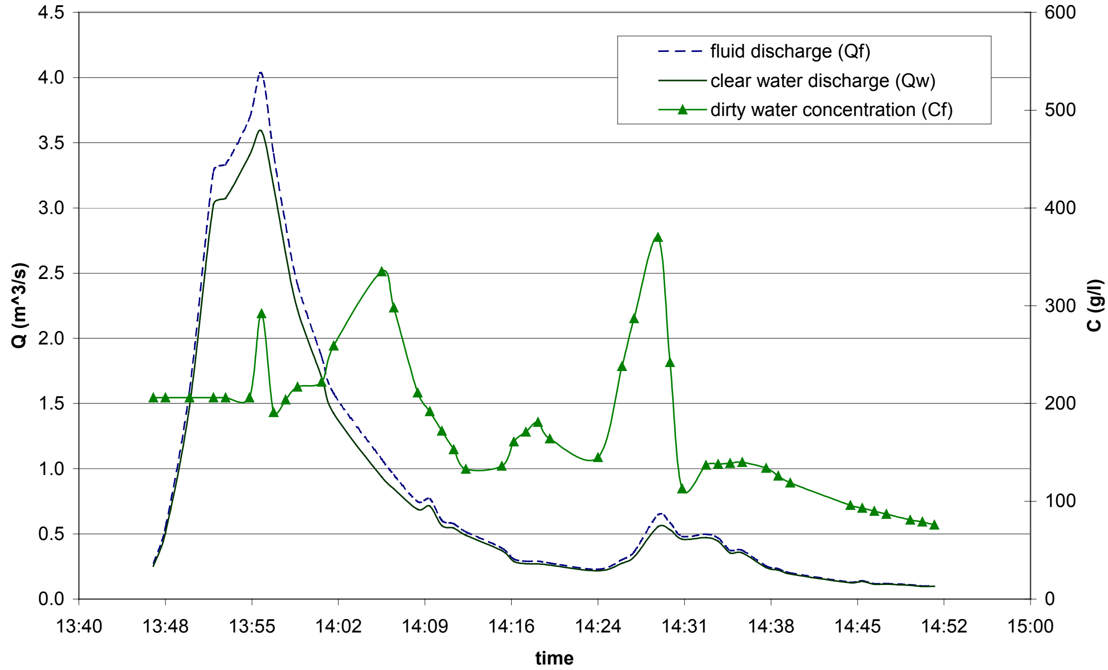

Figure 5.4 Measured discharge and concentration & sediment corrected discharge for the event of July 20th, 1999

5.4.3 Settling velocity

Settling velocity as a function of concentration could not be measured. Since all reported equations (figure 5.2) yielded similar results it seems likely that these equations could also be applied for the Danangou catchment. Grainsize analysis of sediment samples taken at the dam indicated that the amount of clay-sized particles was not much more than 10%. It was therefore assumed that no flocculation structures developed, so that a settling velocity correction based on D50 sufficed. Literature data show that for the sediment concentrations measured in the catchment the settling velocity is significantly decreased. Concentrations of around 400 g/l have been measured repeatedly. Figure 5.2 shows that for such concentrations settling velocity is already half its clear water value. Therefore an equation relating settling velocity to sediment content should be implemented in erosion models.

5.5 Overland flow in the Danangou catchment

Both at the gully-flume and at the sediment plot high dirty water concentrations were measured, at the gully-flume 600 g/l (table 4.11), at the plot about 750 g/l (table 4.12). Both flumes are H-flumes that were constructed according to literature instructions (Bos, 1989).

5.5.1 Velocity and discharge

The discharge equations of H-flumes are based on previous calibrations. This should mean that if the flume is constructed according to the literature instructions the discharge equation is the same as given in the literature. Like for the weir, however, high viscosity might decrease the discharge in comparison to the calibration conditions, while on the other hand higher density and momentum might increase it. The net effect for the discharge equation could in principle be evaluated by comparing the calculated total discharge (from the sensor data) with the total discharge amounts that have been collected using the divisor system (see chapter 4). Assuming that all water was collected in the barrels a difference in total discharge as calculated from the barrel-data and from the water level data could be ascribed to the effect of viscosity. In practice, however, this will not be possible because of uncertainty about measured water level data, sediment levels in the flumes and concentration in the barrels. The discharge equations were therefore not changed.

5.5.2 Sediment content

After an event there is usually a layer of sediment present in the flumes, as shown in figure 5.5. Cantón et al. (2001) tried to solve similar problems by using tilted false floors in their flumes in the Tabernas badlands of southern Spain. This, however, was only partially successful and they were forced to correct the falling limbs of the measured hydrographs. In the Danangou catchment this was also necessary. To be able to determine discharge from the sensor signal one needs to know when the sediment layer developed. Because of the assumption that the discharge equations are correct the total discharge from the barrels can be used to guess at the build-up of sediment. The procedure is to estimate sediment levels during the event based on the levels observed in the flume after the event. By changing the timing of sediment-buildup the total amount of discharge changes and can be made to fit the observed barrel-totals as closely as possible. In this process two assumptions were adopted:

1) That sediment build-up always started after the runoff-peak
2) That the hydrograph should maintain a probable shape, i.e. that it will show an approximately exponential decrease after the peak.

Figure 5.6 shows the results of this procedure for the event of 990710. As can be seen from the figure the measured water level levelled off at about 0.018 m (1.8 cm). This was assumed to be due to a sediment layer of that thickness in the flume. Measurements of the sediment level on the day after the event gave an average sediment level of 1.35 cm in the flume, while at the sensor the sediment thickness was above average. To assume a sediment level of 1.8 cm was therefore acceptable.

Discharge was calculated by applying the discharge equation of the flume (equation 4.9) to the uncorrected measured water level and to the estimated sediment level. Discharge was then calculated as discharge from the uncorrected measured water level minus the hypothetical sediment discharge. This is necessary because of the v-shaped aperture of an

Figure 5.5 Thick sediment layer (about 10 cm) in the gully-flume after the event of 980712. At this time the barrels below the flume had not yet been installed. Picture by E. van de Giessen and J. Snepvangers

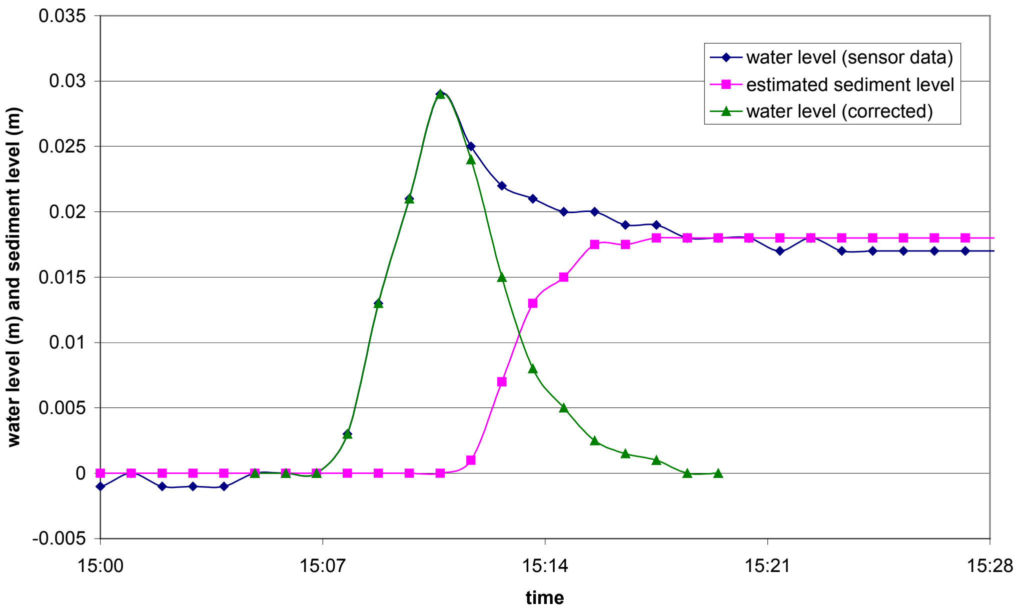

Figure 5.6 Measured water level, estimated sediment level and corrected water level for the event of 990710, sediment plot

H-flume (see figure 5.5). If the discharge equation were applied to the corrected water level discharge would be too low because the water is not flowing over the bottom of the flume, but over the sediment that deposited inside the flume. After the discharge was calculated with the discharge equation, it was corrected to clear water discharge using equation 5.16. Because no timeseries of sediment concentrations were available for the sediment plot the average concentration as determined from the barrels was used. For the gully-flume the data collected with the turbidity sensor could not be used because concentrations were far outside the range of the sensor (chapter 4), so that for the gully-flume the barrel data should also be used. Finally, concentrations expressed in gram per litre clear water were calculated in the same way as described for the weir (equation 5.19).

5.5.3 Settling velocity

Application of the Stokes equation (equation 5.11) for the conditions of the Danangou catchment (d50 about 35 mu) showed that settling velocity in clear water would be about 1 mm/s. Considering the concentrations measured at the flume real settling velocity could be about half that, i.e. 0.5 mm/s. This shows that settling velocity reduction is not likely to be important in shallow flows with depths of several millimetres only.

5.6 Conclusions

High sediment concentrations are a characteristic feature of the Loess Plateau. These high concentrations are probably caused by a combination of factors, in particular the occurrence of erodible materials on steep slopes, the structure and chemical constitution of the loess and the harsh climate that causes plant cover to be low.

When sediment concentration increases fluid density increases, viscosity increases and settling velocity decreases. The effect of this becomes increasingly important with increases in concentration and can result in flow behaviour that is quite different from that of normal streamflow. For large concentrations transport capacity might for example increase. The net effect of these changes on the flow is not always evident, for example the effect on flow velocity and flow resistance remains unclear. Despite this, erosion models that are dealing with high sediment concentrations cannot afford to neglect these effects altogether.

The data collected in the Danangou catchment indicate that even though sediment concentrations were considerable this did not change the fluid flow to such extent that special adaptations are needed to soil erosion models such as LISEM. A number of corrections are, however, necessary to be able to compare field measurements with results of soil erosion models. For the weir sediment volume should be subtracted from runoff volume and a density correction is needed to use data from the pressure transducer. For the flumes, the measured water level should be corrected by subtracting the sediment level in the flume from the water level, while the sediment volume should also be subtracted from the discharge. Finally, measured concentration should be corrected to give concentration expressed as gram per litre clear water.

Literature data show that for the sediment concentrations occurring in the catchment the settling velocity will be significantly reduced, so that soil erosion models should be adapted to incorporate a correction for settling velocity.