1 Introduction

A characteristic feature of many alluvial rivers is the gradual decrease in bed grain size in downstream direction, which is called *downstream fining* (figure 1). The first methodological study of this phenomenon was made by Sternberg (1875), who observed an exponential decrease in grain size on the river Rhine. He ascribed this to the progressive breakdown of particles during their downstream transport, which is usually called *abrasion*. Only four years later Daubrée (1879, *op. cit* Krumbein 1941) gave an alternative explanation for the observed downstream fining pattern, namely *selective transport*, the preferential downstream transport of fine particles.

Most authors, however, did not regard selective transport as a significant cause of downstream fining. This situation lasted until the middle of the twentieth century, when Kuenen's abrasion experiments demonstrated that downstream fining rates in natural rivers were much greater than could be attributed solely to abrasion effects (Kuenen 1956). From then on, most authors considered selective transport as the primary abrasion mechanism.

The downstream change in bed grain size in natural rivers, however, cannot be explained solely by selective transport and abrasion. The introduction of sediment of different origin (for instance at tributary confluences) can obscure the effects of selective transport and abrasion completely (*e.g.* Knighton 1980). The same accounts for the size-selective extraction of sediment from the main channel, for instance through floodplain sedimentation. In lowland rivers, the downstream fining patterns are additionally influenced by the sediment distribution at river bifurcations (*see* Gruijters *et al.* 2001). This influence on downstream fining has hardly received attention in literature.

In this report a detailed overview is given of the present knowledge about downstream fining. The influence of abrasion (chapter 2), selective transport (chapter 3), sediment addition, extraction and redistribution (chapter 4) is described and discussed. Then some attention is attained to a very striking downstream fining phenomenon, the rapid gravel-sand transition. Afterwards an overview is given of available numerical downstream fining models (chapter 6). The most important conclusions are summarized in chapter 7.

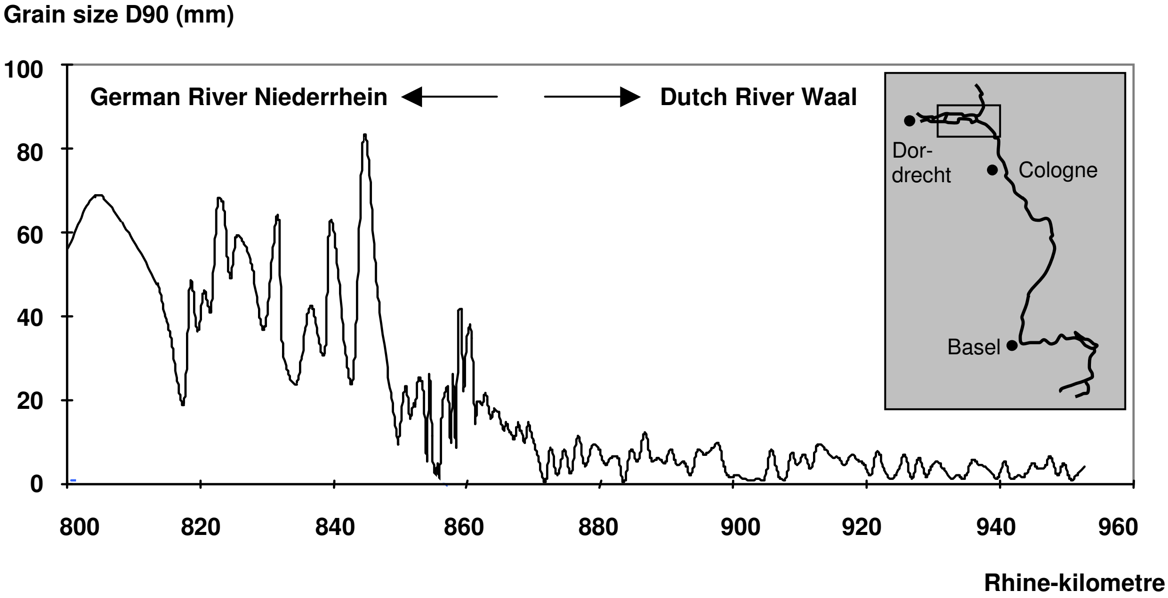

Figure 1. Downstream change in grain size in the river Rhine (*after:* Ten Brinke & Gölz 2001).

2 Abrasion

This chapter focuses on the earliest explanation of downstream fining, *abrasion*, which is a summary term for a range of wearing processes that reduce the size of individual grains during transport. First a description will be given of the various abrasion processes (§2.1), after which the different types of abrasion experiments will be reviewed (§2.2). The findings of these experiments will be discussed in paragraphs 2.3 and 2.4. Paragraph 2.3 describes the factors that determine which abrasion process is dominant, while paragraph 2.4 gives an overview of the factors which influence the abrasion rates. The chapter ends with a synthesis in which an explanation is given for the downstream decrease in abrasion rates, that is frequently observed (§2.5).

2.1 Classification of abrasion

Abrasion has been classified in several ways. To avoid confusion only Kuenen's (1956) classification will be discussed here (figure 2). According to him abrasion can be caused by seven processes: *splitting,* the breaking of grains into two or three parts of roughly equal size, *chipping*, the loss of small flakes form sharp edges, *crushing*, the pulverizing of particles, *cracking*, the formation of small superficial fissures, *grinding*, the loss of fine material during the rubbing of grains against each other, *solution,* the chemical dissociation of grains and *sandblasting,* the grating action of fine grains against large grains. *Sandblasting* can be considered as a special case of *grinding*, and will not be treated separately in this chapter. *Solution,* which can be a very important abrasion mechanism in limestone areas, will neither be treated in this chapter. This is because it has hardly been studied in relation to abrasion.

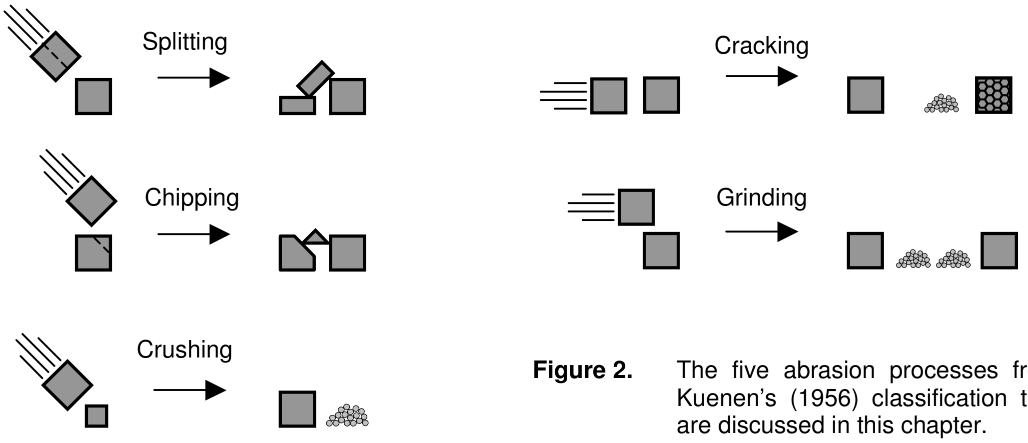

Figure 2. The five abrasion processes from Kuenen's (1956) classification that are discussed in this chapter.

There is a marked distinction between *splitting* and *chipping* on the one hand and *crushing*, *cracking* and *grinding* on the other hand. The first two processes produce particles in the sand-gravel range, while the other processes mainly produce mud and silt (*see* Daubrée 1879 *op. cit.* Krumbein 1941, Krumbein 1941 *and* Bradley 1970). In addition, *crushing*, *cracking* and *grinding* always make the particle roundness increase, while *splitting* and *chipping* often lead to a decrease in particle roundness (Brewer & Lewin 1993).

All processes have in common that they can produce mass loss during particle transport, but processes like grinding can also cause mass loss while a particle is at rest, or slightly vibrating. This was called *abrasion in place* by Schumm & Stevens (1973).

2.2 Abrasion experiments

To gain a better understanding of these abrasion processes many abrasion experiments have been carried out, in de field as well as in the laboratory.

2.2.1 Field studies

Originally, field studies of abrasion processes were mainly based on evaluation of the downstream change in particle roundness (*e.g*. Poser & Hövermann 1951, *op. cit.* Kuenen 1956). A strong increase in particle roundness was seen as a proof for severe abrasion. It has become clear however that an increase in particle roundness not necessarily results from abrasion. It can also be the result of selective transport (*e.g.* Frostick & Reid 1980). On the other hand, strong abrasion can occur without significant increase in particle roundness (Kuenen 1956, Kodama 1994b). It can thus be concluded that abrasion studies based on changes in particle roundness are not very reliable.

Since the middle of the twentieth century field studies of abrasion processes are usually based on an evaluation of the mineralogical composition of the sediment mixture. This can be done in two ways. Some authors compare downstream fining rates of durable and weakly durable minerals and ascribe eventual differences to differences in abrasion rates (*e.g.* Werrity 1992). Others measure the downstream change in mineralogical composition and take a gradual disappearance of weakly durable minerals as a proof for abrasion (*e.g.* Koldewijn 1955, Terwindt *et al.* 1963, Kodama 1994b). Both types of abrasion field studies are based on the assumptions that the influence of selective transport is negligible and that no foreign material is introduced into the river from beneath or from the sides. In rivers in which these assumptions are met, field studies based on mineralogical composition can give useful information about abrasion processes and rates. In most rivers however, the assumptions are not fully met, and for these rivers abrasion studies based on mineralogical grounds are not much more reliable than abrasion studies based on particle roundness.

2.2.2 Tumbling mill experiments

The difficulties in accurately studying abrasion processes in the field, have induced a large quantity of laboratory abrasion studies.

The majority of laboratory abrasion experiments was carried out using a tumbling mill (*e.g.* Krumbein 1941, Gölz *et al.* 1995, Kodama 1994a *and* Jones & Humphrey 1997). In tumbling mill experiments a barrel is filled with water and sediment, put horizontally or slightly tilted on a driving mechanism, and rotated during a given period (figure 3). The sediment load becomes tilted by the rotational movement of the barrel and at its uppermost part grains start moving and become abraded while rolling down.

Since Daubrée's experiment in 1879 barrels of varying size and rotation speeds have been used. Barrel diameters reported range from 10 to 105 cm, barrel lengths from 7 to 125 cm and rotation speeds from 2 to 150 revolutions per minute (Lewin & Brewer 2002). Most barrels have been made of metal, and either left unlined or lined with wood to reduce particle to barrel impact (Lewin & Brewer 2002). Most of the barrels are for the greater part filled with water, but sometimes the water level is lowered, so decreasing the friction exerted on the falling grains and increasing the intensity of the abrasion process (Kodama 1994a). In some cases steel balls are added to increase the intensity of the abrasion process (Gölz *et al.* 1995).

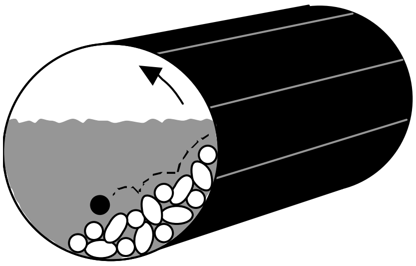

Figure 3. A tumbling mill.

2.2.3 Abrasion tank experiments

An alternative to tumbling mill experiments are the abrasion tank experiments, which were developed by Schoklitsch (1933, *op. cit.* Lewin & Brewer 2002) and Kuenen (1956). Abrasion tanks consist of a circular flume in which water is moved to transport particles across an abrading bed.

Flumes used in these tank experiments vary widely in properties. Channel widths range from 15 to 60 cm, channel depth from 15 to 30 cm and channel circumference from 2.21 to 4.27 m. The flumes were constructed out of concrete, metal or fibre glass while the bed material consisted of concrete or resin, either with or without fixed sand or gravel particles. Water movement was driven by churns, paddles or water jets (Lewin & Brewer 2002).

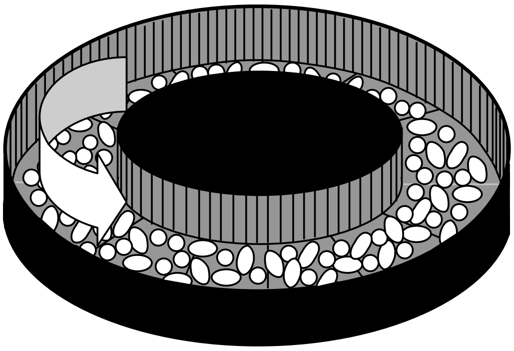

Figure 4. An abrasion tank.

2.2.4 A comparison of tumbling mill and abrasion tank experiments

The main difference between tumbling mill experiments and abrasion tank experiments is the type of grain movement (Kuenen 1956). Grains in tanks are continuously rolling, sliding and saltating over a horizontal surface, while grains in mills alternately drop down a steep slope and then lie still until dropping again. Furthermore, grains in mills interfere with one another's movements, whereas in tanks pebbles nearly always roll separately.

These differences in grain movement have a clear influence on the abrasion process. Abrasion in a tank takes place trough particle-bed contacts, making chipping the principal abrasion process (Brewer & Lewin 1993). Cracking and splitting occur in minor amounts. Abrasion in a mill primarily takes place through particle-particle contacts (Kuenen 1956). Only at low sediment loads some particle-bed abrasion takes places (Lewin & Brewer 2002). The main abrasion process in a mill is therefore grinding, but also some chipping, cracking and splitting occur (Brewer & Lewin 1993).

There has been some debate whether tumbling mill experiments or abrasion tank experiments better represent natural abrasion processes. This is not easy to say. It is possible that tank experiments better represent abrasion processes in some rivers, while mill experiments better reflect abrasion processes in other rivers.

Neither tank experiments nor mill experiments however, are able to properly simulate abrasion by splitting, while this is supposed to be a major abrasion mechanism in the upstream part of rivers (Lewin & Brewer 2002*, see also next paragraph*). The same accounts for abrasion in place. According to Schumm & Stevens (1973) this is of primary importance in natural rivers, but not reproduced in laboratory abrasion experiments.

2.3 Factors determining the dominant abrasion process

From the laboratory abrasion experiments it has become clear that there are marked differences in the intensity of the various abrasion processes. Which abrasion process is dominant depends largely upon two factors: the sediment size and the distance from the river source.

In gravel bed rivers the downstream variation in dominant abrasion processes is as follows. During the first kilometres of transport chipping is generally thought to be the major abrasion mechanism (Kuenen 1956, Abbott & Peterson 1978), possibly in combination with splitting (Krumbein 1941). In these first few kilometres, also some crushing and grinding will occur, because particles are still angular here (Kuenen 1956). Afterwards, grinding (Abbott & Peterson 1978) and cracking (Kuenen 1956) are the principal wearing mechanisms. The influence of splitting and chipping is negligible here (e.g. Abbott & Peterson 1978, Kuenen 1956), but some crushing can occur (Bradley 1970).

For sand bed rivers the pattern of changing abrasion mechanisms in downstream direction is different. Grains rolling on a sandy bottom establish countless contacts all the time, each of which only involves a minute amount of energy. Surface cracks cannot be induced and this leaves only grinding to affect the grains, which is a much less effective mechanism than cracking. During the first few kilometres of transport, however, some chipping can occur (Kuenen 1956).

Relatively little is known about abrasion mechanisms in mixed sand-gravel bed rivers. Bradley (1970) observed that abrasion of sand grains in gravel-sand mixtures was chiefly the result of splitting and crushing probably caused by the impact of gravel. According to Yatsu (1955) also gravel grains in sand-gravel bed rivers are subject to crushing and splitting. Sand-gravel bed rivers often show a rapid transition from a gravel bedded river (D50 about 8 mm) into a sand bedded river (D50 about 2 mm) and Yatsu (1955) explained this by the tendency of fine gravel grains to be crushed and splitted into their individual minerals. Other authors however believe that this is incorrect (*c.f.* Russel 1968).

Though sediment size and distance from the river source are the major factors determining the relative importance of the various abrasion processes, they are not the only factors. Sediment lithology, grain shape and sediment load are also of importance. Kodama (1994a) observed that chert is abraded mainly by splitting (perhaps in combination with chipping), while andesite is abraded by processes like cracking and grinding. Knighton (1982) observed that particles with a platy shape are much more susceptible to splitting than more spherical grains. Lewin & Brewer (2002) noticed that high sediment loads promote abrasive processes which increase particle roundness (crushing, grinding and cracking), while low sediment loads promote abrasion processes which decrease particle roundness (splitting and chipping).

2.4 Factors influencing the abrasion rate

The several laboratory abrasion experiments have also brought insight in the factors that influence the abrasion rate. Important factors are: grain size, lithology, grain roundness, grain shape, grain velocity, mixture composition, river bed grain size, amount of weathering, number of moving particles and the dominant abrasion process. It is difficult to quantify the effects of these factors, because they can hardly be isolated. Yet, some values of the abrasion rates will be given in this paragraph. These values should be taken as indications of the size order of the abrasion rate, not as exact predictions. The exact abrasion values depend upon the interaction between all the factors mentioned, which will be discussed in paragraph 2.5.

2.4.1 Grain size

Both abrasion tank experiments and tumbling mill experiments have revealed that the abrasion rate increases with grain size. In gravel bed experiments by Bradley (1970) and Lewin & Brewer (2002), the absolute size reduction was directly proportional to initial grain size, which implies that the percent size reduction was equal for all grain sizes. This was also the case in Kuenen's (1956) sand bed experiment. In Kuenen's gravel bed experiment however, both the absolute and the relative size reduction increased with grain size (figure 5).

The common increase of abrasion rates with grain size can be explained by the lower kinetic energy of smaller particles and by the fact that smaller particles have longer saltation lengths and do not hit the river bed as much as larger particles (Bradley 1970).

Sand-size grains are transported in suspension at most discharges and therefore have extremely low abrasion rates (Kuenen 1956, 1959). This is consistent with Cailleux's findings (1942), who noticed that sand-sized sandstone particles did not loose their coat of iron oxides during transport in the river Rhine.

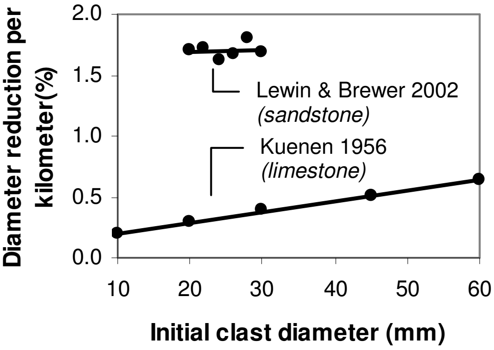

Figure 5. Percent diameter reduction per kilometre of travel in two gravel bed tanks.

2.4.2 Lithology

The influence of grain lithology on its resistance (durability) against abrasion was already noticed by Daubrée (1879 op. cit. Krumbein 1941) and later on confirmed in laboratory and field experiments.

Morris & Williams (1999a) give an overview of abrasion durability of some common lithologies, based on laboratory mill and tank experiments carried out by Krumbein (1941), Kuenen (1956, 1959), Bradley (1970, Bradley *et al.* 1972), Kodama (1994a) and others. This overview is presented in table 1. Table 1 also contains data for other lithologies, based on experiments by Abbott & Peterson (1978). It should be realized that in most experiments tap water was used, which can be chemically very different from river water. The data shown in table 1 are thus not necessarily representative for the abrasion durability in natural rivers.

For each lithology mentioned in table 1, information is given about its physical structure and about the hardness of the constituting minerals, which is given on Moh's scale. It can be seen that in general non-massive lithologies and lithologies made up of minerals with a low hardness have the least abrasion durability, as can be expected.

The exact order of rank of lithologies found by Morris & Williams (1999a) is slightly different from that found by Abbott & Peterson (1978). Especially the different position of chert is striking. According to Morris & Williams chert is easily abraded, while Abbott & Peterson put it among the most durable lithologies. This difference is caused by differences in the size of the chert grains used in the abrasion experiments. The chert grains reported by Morris & Williams were relatively large (about 10 cm) and heavily jointed, while the grains tested by Abbott & Peterson were much smaller (about 4 cm) and probably not jointed, but compact (see Kodama 1994a). From this it can be presumed that abrasion durability for small grains primarily depends on the hardness of the composing minerals, while the abrasion durability for large grains primarily depends on the physical structure of the grain (jointing susceptibility). This is supported by the rank order of lithologies for sand-size grains (lower part of table 1). The durability rank order perfectly mirrors the hardness rank order.

Abbott & Peterson (1978) performed monolithological and polylithological experiments in an tumbling mill. In the polylithological experiments they found the same sequence of abrasion durability as in the monolithological experiments, but abrasion rates of less durable rock types increased whereas rates of more durable types decreased. Abbott and Peterson suggested that in polylithological sediment mixtures first the weakly durable rocks are abraded by impacts of more durable rocks. When the weakly durable rocks have disappeared abrasion of moderately resistant rock types starts.

Morris & Williams (1999a)		**Abbot & Peterson (1978)**	
Gravel		***Gravel***	
Andesite	(6, often layered)	Schist	(5-6, well-foliated)
Chert	(6.5-7.5, variable)	Marble	(2.5-6, massive or foliated)
Shale	(< 3, layered)	Basalt	(3-6, variable)
Coal	(3-5, massive but brittle)	Gabbro	(5-6, massive, but banded)
Limestone	(3-4, layered)	Granodiorite	(6-7, massive)
Gneiss	(6-7, slightly foliated)	Gneiss	(6-7, slightly foliated)
Granite	(6-7, massive)	(meta-)Sandstone	(6-7, massive or layered)
Obsidian	(5-6, massive)	Obsidian	(5-6, massive)
Greywacke	(6-7, coarse layered)	(meta-) Breccia	(, variable)
Amphibolite	(4-6.5, massive or foliated)	Rhyolite	(6-7, variable)
Quartz	(7, massive)	Quartzite	(7, massive or foliated)
Aplite/pegmatite	(6-7, massive or banded)	Chert	(6.5-7.5, variable)
Quartzite	(7, massive or foliated)		
Quartz porphyry	(7, massive)		
Flint	(6.5-7.5, massive, fractured)		
Agate	(6.5-7.5, banded)		

Increasing durability

1 % Approx. size reduction per km 0.01 %

Sand	
Limestone	(3-4)
Apatite	(5)
Hornblende	(5-6)
Garnet	(6-7.5)
Quartz	(7)
Orthoclase	(6)

Increasing durability

Size reduction per km

0.01 % 0.0001 %

Table 1. Abrasion durability of some common lithologies. Information about Moh's hardness and the physical structure of the lithology is given between brackets (largely based on Mottana *et al.* 1978).

3.4.3 Grain roundness

The abrasion rate is noticeably more rapid for angular particles than for well-rounded particles. This was already noticed by Daubrée (1879 op. cit. Krumbein 1941) and later on confirmed by Kuenen (1956), Gölz *et al.* (1995) and others, both in laboratory and field experiments. The primary reason for this is that sharp-angled pieces are very prone to chipping, which is one of the most effective abrasion mechanisms (§ 2.4.10). Another effect is that angular particles have a larger surface area than rounded particles, which promotes the mass loss through grinding and cracking.

A more quantitative idea of the influence of roundness on the abrasion rate can be obtained from figure 6. Shown are the abrasion rate per kilometre and the associated change in roundness for a granite grain travelling in a gravel-bed flume. It can be seen that the abrasion rate is about five times as high during the first kilometres of travel (where the grains are still angular), than further on (where the grains are well-rounded). It is possible however, that this reduction in abrasion rate only partly results from the change in grain roundness. Apart from a grain size effect (see § 2.4.1), also a lithology effect can be present. It is probable that during the first kilometres of transport the weakest grains within a given lithology will all quickly disintegrate, resulting in high abrasion rates, while further downstream, when these weakest grains have disappeared, abrasion rates are much lower.

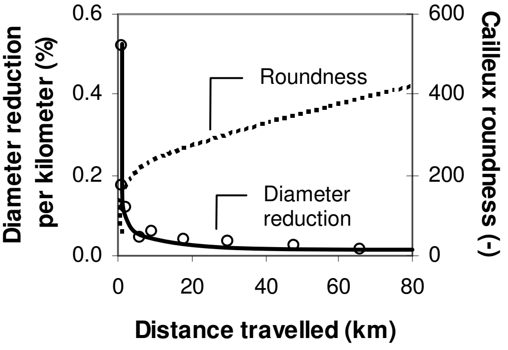

Figure 6. The influence of clast roundness on the abrasion rate for a granite grain of 44 mm (graph based on Kuenen's 1956 experiment E).

2.4.4 Grain shape

The precise influence of grain shape (a measure of the ratio between the three grain axes) on abrasion rates is unclear. Kuenen (1956) considered shape influence negligible compared to the influence of grain roundness and size. Lewin & Brewer (2002) however observed a clear shape influence. This influence was quite contrasted in their tank and barrel experiments, with cubes losing most weight in the barrel but least in the tank.

2.4.5 Grain velocity

The grain velocity influences the abrasion rate because it determines the force (or energy) with which sediment particles bump into each other. Since the energy of a moving grain is proportional to the square of its velocity, it can be expected that the abrasion rate increases quadratically with particle velocity. According to Kuenen (1956) this is indeed the case for gravel bed rivers, but he found only a minor influence of grain velocity on abrasion rate in his sand bed tank experiments.

In the gravel bed tank experiments performed by Lewin & Brewer (2002), the *flow* velocity was measured as surrogate for the grain velocity. They found only a weak dependency of abrasion rates on flow velocity. They argued that while impact energy increases at increasing flow velocities, the number of rolling grains decreases because more grains start saltating.

2.4.6 Mixture composition

The effect of combining coarse and fine grains on abrasion rates has been studied in tumbling mill experiments (Kodama 1994a, Marshall 1927, *op. cit.* Krumbein 1941) and in abrasion tank experiments (Kuenen 1956, 1959).

The abrasion rate of coarse particles diminishes when fines are added, according to Kuenen (1956, 1959). This was also observed by Kodama (1994a) (compare the two black bars in figure 7). Both Kodama and Kuenen only compared abrasion rates for experiments with and without fines. Their experiments give no information about the exact decrease in abrasion rate for a given amount of fines added.

It is unclear what happens to the abrasion rate of fine grains when coarse grains are added. Marshall (1927 *op. cit.* Krumbein 1941) and Kodama (1994a) observed a strong increase in fine grain abrasion rates (grey bars in figure 7), but others found no change in fine grain abrasion rates (Kuenen 1959). This difference could be due to differences in experimental equipment.

Due to these opposing findings, the result of combining coarse and fine grains on the overall abrasion rate is still unclear. Kodama's (1994a) findings imply that the overall abrasion rates in bimodal sediment mixtures are much higher than in uniform sediment mixtures, but from Kuenen's (1956, 1959) results it can be surmised that this is not true.

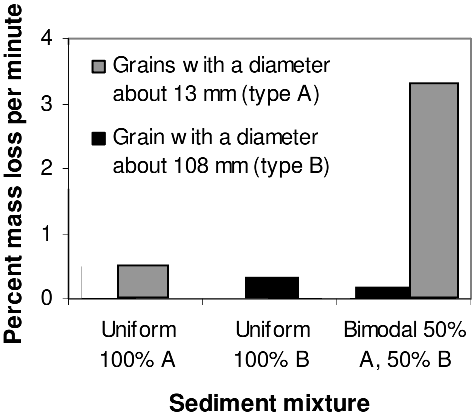

Figure 7. The influence of mixture composition on the abrasion rate for andesite grains (Kodama 1994a) Note: Kodama's ab-rasion rates cannot easily be expressed in percent diameter reduction per kilo-metre.

2.4.7 River bed grain size

The effect of the bed grain size on abrasion rates was studied by Kuenen (1956) and Lewin & Brewer (2002) in their abrasion tanks. Both Kuenen and Lewin & Brewer found that grains moving over a smooth bed experienced much less abrasion than grains moving over a pebble bed.

Kuenen (1956) explained this by differences in bed roughness and type of motion. A smooth (fine-sandy) bed is very smooth to moving gravel grains, while a pebble bed is relatively rough to moving gravel grains. According to Kuenen (1956) this leads to a skidding type of motion on sand beds, while grains on gravel beds are mostly rolling. Since rolling produces more abrasion than skidding, abrasion rates are higher on gravel beds than on sand beds.

However, the situation is more complicated than assumed by Kuenen. Lewin & Brewer (2002) also performed experiments with a bed made of grit (coarse sand). They found that abrasion rates on grit beds are markedly higher than abrasion rates on both coarser and finer beds (figure 8). An explanation was not given. Nevertheless, it is clear that the effect of bed grain size on abrasion is still not fully understood.

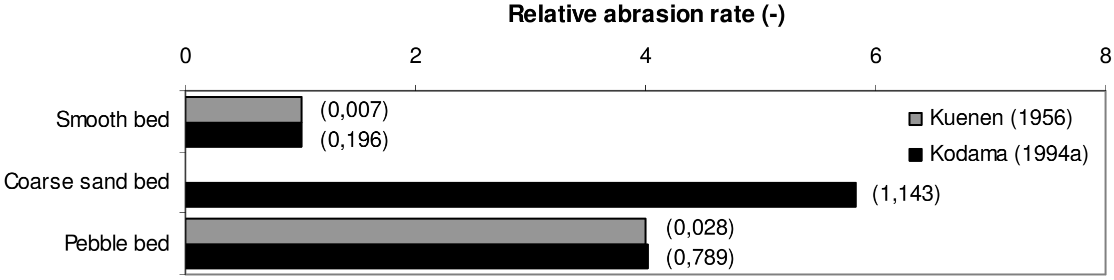

Figure 8. The influence of river bed grain size on the abrasion rate according to Kuenen (1956) and Kodama (1994a). Because both authors used different lithologies (respectively quartzite and sandstone), the abrasion rates were scaled by dividing the abrasion rate for a given bed by the abrasion rate of a smooth bed. Unscaled abrasion rates, expressed as percent diameter reduction per kilometre, are given between brackets.

2.4.8 Weathering

In abrasion tank experiments and tumbling mill experiments it is regularly observed that natural grains show higher abrasion rates than prepared (freshly broken) grains (Kuenen 1959, Bradley 1970, Jones & Humphrey 1997, Lewin & Brewer 2002). This is because natural grains have often suffered from weathering due to chemical, mechanical or biological processes, while prepared grains have not.

Bradley (1970) argued that grains that are not submerged during low flow conditions, suffer from severe weathering. When these sediments are mobilized during floods, abrasion only removes the weathered skin of the particle, irrespective of flood duration.

According to Jones & Humphrey (1997), grains are not remobilised in each flood period. When a grain has become deposited on a point bar, it is supposed to be immobile for centuries. In this period the grain becomes heavily weathered. When the main channel migrates the grain can be taken up again by the flow in the outer bend. It then is transported to the next point bar where it is deposited again. During this short period of transport, the weathered skin of the particle is removed by abrasion.

It is not clear yet whether these hypotheses are correct or not. Weathering however mainly attacks grains resting above the water level, especially if biological activity on the river bottom is minimal. In rivers in which most sediment particles stay submerged for the whole of the year, weathering will be negligible (Jones & Humphrey 1997).

The hypotheses have some interesting implications (Jones & Humphrey 1997). In the first place is it implicitly assumed that small particles have a larger percent of diameter reduction during abrasion than coarse particles, because the weathered skin is as thick for the small particles as for the coarse particles, but the grains are much smaller. This is opposite to common observations (see above). Secondly, floodplain sediments should show much higher abrasion rates in laboratory experiments than sediments from the main channel. This is not the case either, but it can be argued that the mean distance travelled in the river between times of weathering is so short that only a fraction of the weathered skin layer is removed (Jones & Humphrey 1997).

Another implication of the hypotheses is that abrasion in laboratory experiments also only involves the removal of the weathered skin layer (Kodama 1994a). This implies that it is useless to conduct long-duration abrasion experiments, for a longer experiment duration will not increase the mass loss.

2.4.9 Amount of moving particles

In their tumbling mill experiments, Lewin & Brewer (2002) observed an increase in abrasion rates as the sediment load decreased. Gölz *et al.* (1995), however, noticed a much higher mass loss in tumbling mill experiments with a large sediment load than in experiments with a small sediment load. A satisfying explanation for this difference is not yet available.

2.4.10 Dominant abrasion process

The final factor that influences the abrasion rate is the dominant abrasion mechanism. Splitting and chipping involve the largest mass losses and therefore lead to higher abrasion rates than the other abrasion processes. Abrasion rates caused by crushing are less, but still higher than the abrasion rates caused by cracking. Grinding is the least effective abrasion mechanism (Kuenen 1956).

2.5 Synthesis

In the past 125 years many experiments have been carried out to determine the factors that control abrasion processes and rates. Most of these experiments have been carried out in laboratories because it is very difficult to distinguish between the effects of abrasion, selective transport and sediment exchange in the field. Laboratory experiments fall apart in tumbling mill experiments and abrasion tank experiments. Abrasion mechanisms differ significantly between mill experiments and tank experiments as a result of differences in the type of grain movement. It is still unclear whether tumbling mill experiments or abrasion tank experiments better represent natural abrasion processes.

From the several abrasion experiments it has become clear that the relative importance of the various abrasion processes primarily depends on the sediment size and the distance from the river source. Near the source of gravel bed rivers, the dominant abrasion processes are chipping and splitting. Further downstream grinding and cracking become the principal processes. The dominant abrasion process in sand bed rivers is grinding, though some chipping can occur during the first kilometres of transport. Little is known about the dominant abrasion mechanism in mixed sand-gravel bed rivers. There are some indications that splitting and crushing are highly important.

Abrasion rates are influenced by the size, lithology, roundness, shape, velocity and number of moving grains, by the size distribution of the sediment mixture, by the river bed grain size, by the relative importance of abrasion processes and by weathering processes. Grain size and lithology are the most important among these factors. If it is assumed that the percent size reduction is equal for each grain size (§ 4.2.1), than the absolute size reduction per kilometre for fine sandy sediments must be about a factor 1000 less than the absolute size reduction for coarse gravely sediments (cobbles). Variations in lithology can cause differences in abrasion rates of a factor 100 (table 1). The other factors only cause variations in abrasion rates of a factor 10 or less under natural conditions.

Though is has been possible to distinguish between the separate influences of these factors in laboratory experiments, this is very difficult in field conditions. However, in many rivers some of the factors determining the abrasion rate are interrelated, amplifying each other's influence and causing a strong decline in abrasion rates in downstream direction. Thus, while it remains impossible to predict the exact abrasion rate at a given location, the downstream change in abrasion rate can easily be estimated. This is illustrated in figure 9 for a hypothetical river.

In the upstream part of the hypothetical river the strong bed curvature produces high shear stresses and particle velocities. Furthermore, there is a continuous supply of fresh, angular, coarse rock fragments. This all leads to high abrasion rates, primarily by splitting and chipping.

These high abrasion rates result in a strong increase in grain roundness and a decline in the presence of weakly durable lithologies. Furthermore the grain size strongly decreases, because of the high abrasion rates in combination with selective transport (*see next chapter*). Particle velocity also decreases as a result of the lower bed slope. Splitting and chipping do not occur anymore due to the decreased roundness and make place for the less effective processes cracking and grinding. All these changes contribute to a strong decline in abrasion rates. The decrease in grain size, weakly durable lithologies and particle velocity, and the increase in grain roundness still continue, however. Combined with the gradual transition to sand beds this leads to a further reduction in abrasion rates.

Though in general abrasion rates decrease in downstream direction, it is possible that during the transition from a gravel bed to a sand bed locally higher abrasion rates prevail. It has been argued that the presence of both coarse and fine grains in this section promotes crushing, which leads to an increased abrasion of the fine grains (see Bradley 1970 & Kodama 1994a), but this is not generally accepted (see Russel 1968 and Kuenen 1956, 1959).

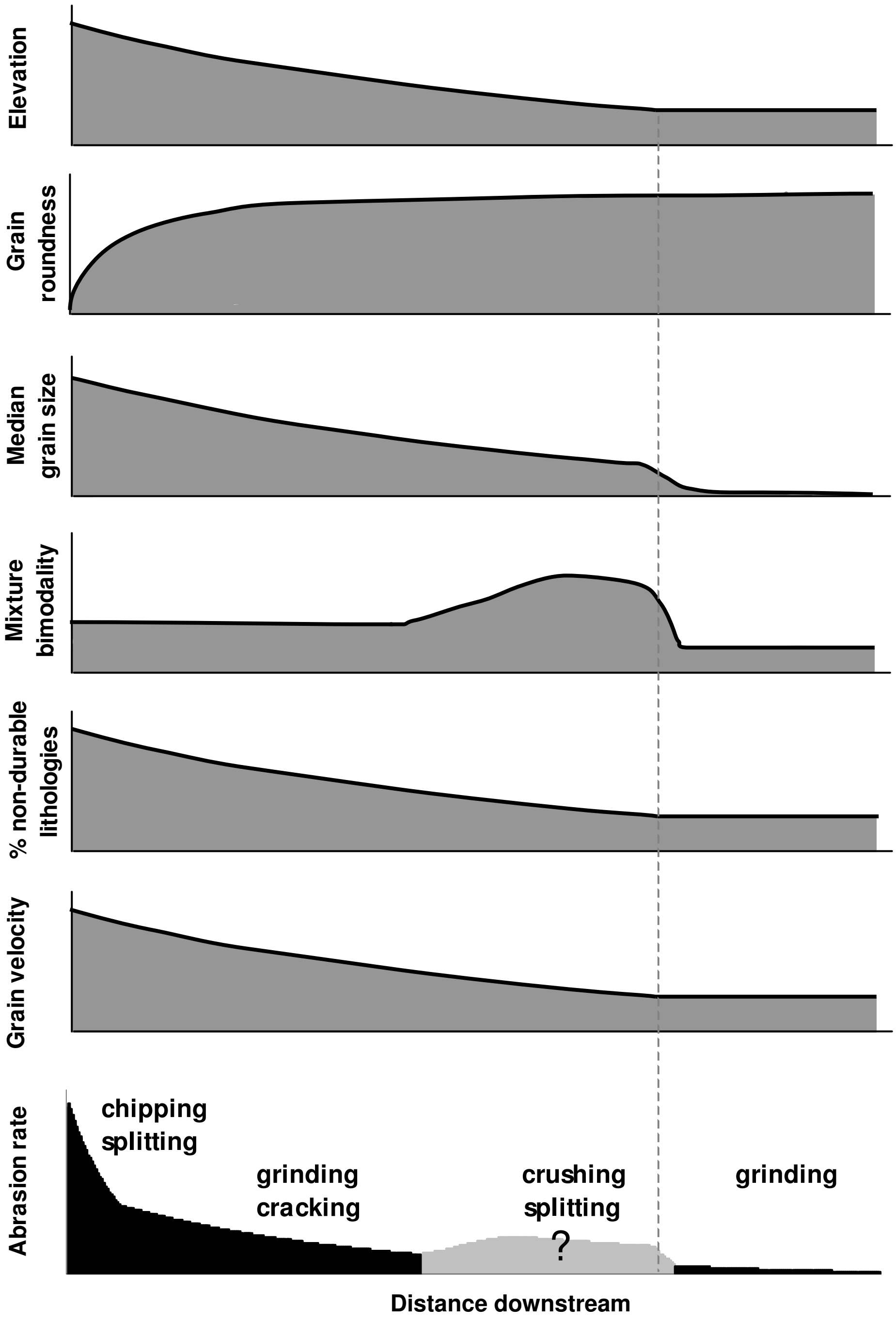

Figure 9. Downstream change in abrasion processes and rates in a hypothetical river, as the result of the downstream change in grain roundness, grain size, sediment mixture, grain velocity and lithology. (In the downstream change in grain size also the effects of selective transport are incorporated.)

3. Selective transport

Though several forms of selective transport can be distinguished (*e.g.* Frostick & Reid 1980, Steidtmann 1982, Reid & Frostick), only one of them causes downstream fining: *grain size selective transport*, the preferential downstream transport of fine particles. In this chapter the term *selective transport* therefore only refers to the process of *grain size selective transport*.

Selective transport can result from selective grain entrainment, selective grain movement and selective grain deposition, but also from local sorting processes like dune sorting and armouring. These four types of processes will be discussed successively in this chapter (§3.1-§3.4). In paragraph 3.5 the overall degree of size-selectivity due to the combined action of selective entrainment, selective movement, selective deposition and sorting will be evaluated, while paragraph 3.6 describes how this selectivity affects downstream fining.

3.1 Selective entrainment

The entrainment process is called selective if fine grains are entrained at a different (usually lower) shear stress than coarser grains. To evaluate the size-selectivity of the entrainment process, the conditions under which grains of different sizes start to move will be discussed.

3.1.1 Entrainment criteria

Criteria for the beginning of motion of sediment grains can be expressed in terms of a critical discharge (*e.g.* Meyer-Peter et al. 1934, *op. cit.* Lenzi *et al.* 1999, Ferguson 1994) or a critical settling velocity (*e.g.* Komar & Clemens 1986), but commonly a criterion based on the critical shear stress is used. Grains are set into motion when the instantaneous (turbulence-driven) value of the bed shear stress (τ, N/m^2) exerted on the grains becomes greater than a critical value (τ_c, N/m^2), which depends on the grain size and the friction with the neighbouring grains:

$$\tau \geq \tau_c \qquad 3.1$$

The instantaneous bed shear stress τ exerted on a grain can be estimated in two steps. First, the total time-averaged bed shear stress need to be corrected for the ‘loss’ of shear stress in the eddies behind bed forms, in order to get the time-averaged shear stress exerted on the bed surface grains (*e.g.* Carson 1987). Second, by combining the time-averaged grain shear stress with a theoretical probability density function, instantaneous bed shear stress values can be estimated (*e.g.* Bridge & Bennett 1992, Kleinhans & Van Rijn 2002).

The critical bed shear stress (τ_c) is much more difficult to determine. The just described way to estimate the instantaneous bed shear stress is stochastical in nature and thus not capable of reproducing the exact value of the instantaneous bed shear stress that caused an observed grain movement. Therefore, usually the value of the *time-averaged* bed shear stress that prevailed at the moment of incipient motion is taken as the critical shear stress. Fernandez Luque & Van Beek (1976) estimated that the true value of τ_c, based on the instantaneous bed shear stress, is about two times as large as the time-averaged value (see also Zanke 2003). Because the critical shear stress τ_c is usually based on the time-averaged bed shear stress, it is questionable whether it still makes sense to use instantaneous values for the bed shear stress τ.

In most cases this problem is not relevant, however, because generally both τ and τ_c are expressed in terms of the time-averaged bed shear stress. Yet, this induces another confusion. When using time-averaged values of τ and τ_c, validity of equation 3.1 is no longer an absolute requisite for motion. There can always be a turbulent sweep that sets a grain into motion, while the time-averaged bed shear stress is still lower than the critical value. This is even more the case when a sediment mixture is considered. Each grain in the mixture has its own critical shear stress value, and a grain that is resting very loose on the other grains will be entrained before the shear stress overcomes the mixture-representative critical shear stress.

3.1.2 The critical shear stress in uniform sediments

Shields (1936, *op. cit* Lenzi *et al.* 1999) determined the critical shear stress for the beginning of motion in uniform sediment mixtures. He found the following relation:

$$\tau_c = (\rho - \rho_s) g D_x \theta_c \qquad 3.2$$

with ρ the fluid density [kg/m^3], ρ_s the particle density [kg/m^3], g the acceleration due to gravity [m/s^2], D_x a representative mixture grain size [m] and θ_c the dimensionless mobility parameter of Shields [-].

Shields (1936 *op. cit.* Andrews 1983) conducted several experiments in a flume with a horizontal bed and nearly uniform sediments (D_{50} from 0.36 to 3.44 millimetres) and found that θ_c varies with the particle Reynolds number ($Re*$). Bonnefille (1963 *op. cit.* Van Rijn 1993) and Chabert & Chauvin (1963) expressed Shields' θ_c - $Re*$ relationship in terms of θ_c and a dimensionless grain size $D*$. From their graphs it can be seen that θ_c is slightly dependent on D_{50} for grains smaller than 1 millimetre, but virtually grain-size independent for larger grains. This implies that the critical shear stress τ_c is directly proportional to the grain size for grains larger than 1 millimetre (figure 10). There have been numerous additions, revisions and modifications of the Shields curve since its original publication. Most of these studies also found a linear relationship between τ_c and D, the exact value of τ_c depending on bed slope (*e.g.* Fernandez Luque & Van Beek 1976), grain shape (Gomez 1994), bed structure (Church *et al.* 1998), water depth (Zanke 2003) and on the method used (Buffington & Montgomery 1997).

3.1.3 The critical shear stress in non-uniform sediments

For non-uniform sediment mixtures usually for each size fraction a different critical shear stress is used. It must be noted that this represents the shear stress at which the first grains in a grain size fraction begin to move, rather than being the average of all the individual particle τ_c values in that size fraction. Equation 3.1 thus changes into:

$$\tau \geq \tau_{c,i} \qquad 3.3$$

with $\tau_{c,i}$ the critical bed shear stress for grain size fraction i [N/m^2]. Equation 3.3 provides the basis for evaluating the degree of size-selectivity of the entrainment process. The most obvious requisite for selective entrainment is that $\tau_{c,i}$ is larger for coarse grain size fractions than for fine size fractions. Equally important, however, is the requisite that τ should lie in between the $\tau_{c,i}$ for the coarsest size fraction and the $\tau_{c,i}$ for the finest size fraction for a considerable part of the year (Pitlick 1989). If this requisite is not satisfied, it is impossible to get selective entrainment, because either all or none of the grain size fractions will be in motion. In literature, however, usually only attention is paid to the first requisite. Therefore, in the following evaluation of the degree of selectivity during sediment entrainment, only the grain-size dependence of the critical bed shear stress is discussed.

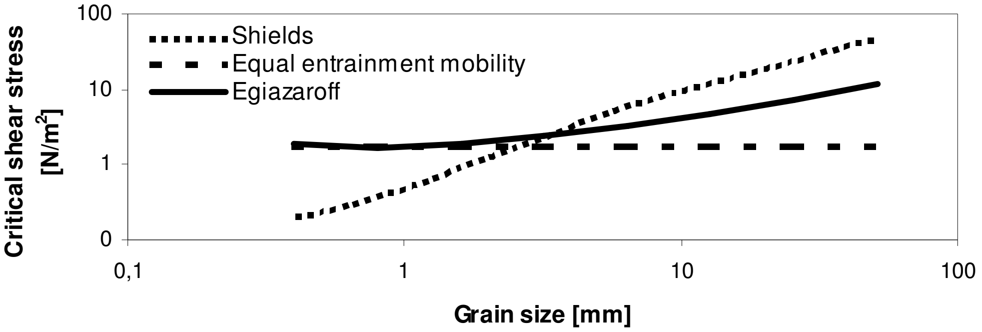

Figure 10. The critical shear stress according to Shields (1936) and Egiazaroff (1965) for a mixture with an average grain size of 2 millimetres. One example of equal entrainment mobility is also given (the horizontal line can be shifted up and down ad libitum).

If a non-uniform sediment mixture is seen as a set of uniform sediment mixtures with different grain sizes, the critical shear stress for each size fraction could be determined with Shields' curve. This approach is not correct, however, because it assumes that all size fractions behave independently of each other, which is not the case. In non-uniform sediment mixtures the entrainment of the different size fractions is strongly affected by hiding and exposure effects. Coarser particles are more exposed to the flow and therefore have smaller critical shear stresses than they would have in uniform sediments. Finer particles are hiding in the wake of coarser particles and therefore have larger critical shear stresses than they would have in uniform sediments (Andrews 1983). Hiding and exposure effects thus reduce the differences in critical bed shear stress between the various grain size fractions.

Only a few theoretical analyses have been taken out to determine the critical shear stress in sediment mixtures. The most famous of these relationships is that derived by Egiazaroff (1965), which can be written as:

$$\tau_{c,i} = \frac{0.1(\rho_s - \rho)gD_i}{\left[\log 19 \dfrac{D_i}{D_{avg}}\right]^2} \qquad 3.4$$

with: D_i the average grain size of size fraction i [m] and D_{avg} the average grain size of the entire mixture [m].

In figure 10 Egiazaroff's relationship is plotted, together with Shields' curve. In the graph also the situation of equal entrainment mobility is sketched, which implies that all grain size fractions start moving at the same shear stress. From figure 10 it can be seen that the grain size influence on the critical bed shear stress according to Egiazaroff is much less than predicted by Shields' curve, due to hiding-exposure effects. For grain sizes smaller than the average grain size, the critical shear stress is even independent of grain size.

Usually, the grain size influence on the critical shear stress is determined experimentally. Therefore measured values of the critical shear stress are made dimensionless [$\theta_{c,\,i} = \tau_{c,i}/(\rho_s - \rho)gD_i$] and plotted against the ratio D_i/D_{50} after which a power function is fitted:

$$\theta_{c,i} = a\left[\frac{D_i}{D_{50}}\right]^t \qquad 3.5$$

This can be written dimensionally as:

$$\tau_{c,i} = a(\rho_s - \rho)gD_i\left[\frac{D_i}{D_{50}}\right]^t \qquad 3.6$$

with $\theta_{c,\,i}$ the dimensionless critical shear stress for size fraction i [-], D_{50} the median grain size of the entire mixture [m], t the hiding-exposure coefficient [-] and a a constant [N/m^2], that is usually taken to represent the dimensionless mobility parameter (θ_c) in a uniform sediment mixture with the same D_{50} or D_{65}.

Though there is no theoretical justification for the fitting of a power function (see also §3.1.5), the empirical approach has been widely used to estimate the degree of selective entrainment in field and laboratory situations. From equation 3.6 it can be seen that equal entrainment mobility will occur if the hiding-exposure coefficient (hereafter denoted as t-value) equals –1. The more the t-value deviates from –1, the stronger the degree of selective entrainment will be.

An overview of t-values from both field measurements and flume experiments is given in appendix 1. It becomes clear that t-values show a broad range, indicating strongly selective entrainment in some rivers but equal entrainment mobility in others. The huge variation in t-values stems partly from differences in sediment mixture and bed characteristics and partly from differences in the experimental methods used to determine the t-values.

3.1.4 The influence of mixture and bed characteristics on hiding-exposure coefficients

The degree of selective entrainment generally grows as the bimodality of the sediment mixture increases (Church *et al.* 1991, Wilcock 1993, Kuhnle 1993). This is illustrated in figure 11. Unimodal and weakly bimodal sediments exhibit severe hiding-exposure effects and the entrainment process is only weakly size-selective with *t*-values close to –1. In strongly bimodal sediments, however, grains move independently of each other. Hiding-exposure effects therefore are virtually absent and sediment entrainment is size-selective, with *t*-values from -0.28 to -0.68.

In addition to the mixture bimodality, also the median grain size influences the *t*-value. For a give value of the mixture bimodality, Shvidchenko *et al.* (2001) found that the hiding and exposure effect is most pronounced (*t* closest to –1) for mixtures with a D_{50} around 5 millimetres. For smaller and larger values of D_{50}, the hiding-exposure effect is reduced and a higher degree of size-selectivity is observed.

Another factor that influences the t-value is the structure of the riverbed. The degree of hiding of small grains is strongly determined by the concentration of coarse grains on the bed surface, according to Wörman (1992). This is because small grains between closely spaced coarse grains experience a different flow field than small grains between sparse coarse grains. The influence of bed structure on the critical shear stress and the degree of hiding-exposure was also stressed by *e.g.* Reid & Frostick (1987), Gomez (1994) and Church *et al.* (1998). They focused respectively on pebble clusters, imbrication and stone cells.

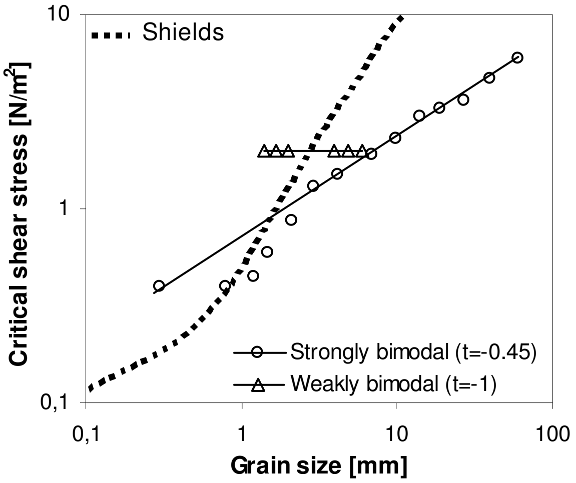

Figure 11. The influence of mixture bimodality on the critical shear stress (the examples shown are the BOMC and MC-50 mixtures analysed by Wilcock 1993 and Wilcock & McArdell 1993).

3.1.5 The influence of experimental methods on hiding-exposure coefficients

Methodologies to determine *t*-values differ on five important aspects: the method to determine the critical bed shear stress, the range of D_i/D_{50} values over which *t*-values are determined, the definition of incipient motion, the choice for D_{50} and the way corrections are made for specific environmental conditions.

The critical bed shear stress $\tau_{c,i}$ can be determined using either the largest grain method, the visual observation method, or the reference transport method. In the largest grain method the actual bed shear stress (τ) is assumed to be equal to the critical bed shear stress ($\tau_{c,i}$) for the coarsest grain in motion as long as there are coarser grains on the bed that are not moving (Andrews 1983). According to Batalla & Martín-Vide (2001) this method is inaccurate because of the low probability of capturing the largest moving particle in a bed load sampler. In the second method the bed shear stress at which a given grain size fraction starts moving is determined visually. This method is direct, but can be subjective depending on one's observation of how much movement constitutes initial motion (see Buffington & Montgomery 1997). The reference transport method estimates the critical shear stress ($\tau_{c,i}$) as the shear stress that produces a low but measurable reference transport (see Parker *et al.* 1982). The $\tau_{c,i}$-values obtained with this method are especially sensitive to the extrapolation method and the particular reference transport value that is chosen (see below). It can be seen from appendix 1 that studies, in which the critical shear stress was determined with the largest grain method or by visual observation, indicate a larger degree of selective transport, than studies in which the reference transport method was used. Average *t*-values are –0.70 for the largest grain method and –0.87 for the reference transport method (see also Batalla & Martín-Vide 2001 and Wathen *et al.* 1995).

The influence of the D_i/D_{50} range over which the t-value is determined is shown in figure 12. It can be seen that the smaller the start value of the D_i/D_{50} range, the higher the t-value. This leads to the conclusion that the finer fractions in a sediment mixture are associated with higher t-values. This implies that the fitting of a power function with a constant power for all grain size fractions (equation 3.5) is not correct. Some authors have tried to solve this problem by fitting two power functions through their critical shear stress data, one for the fine fractions and one for the coarse fractions (*e.g.* Ashida & Michiue 1971 *op. cit.* Patel & Ranga Raju 1999, Kuhnle 1993, figure 13). The t-value for the fine fractions then usually lies around –1, while the t-value for the coarse fractions are between –0.3 and –0.7. A better solution, however, would be to fit an Egiazaroff-like function. Egiazaroff's curve (figure 10) correctly predicts that the entrainment of grain size fractions smaller than the median grain size is grain-size independent, while grain size fractions much coarser than the median grain size exhibit selective entrainment.

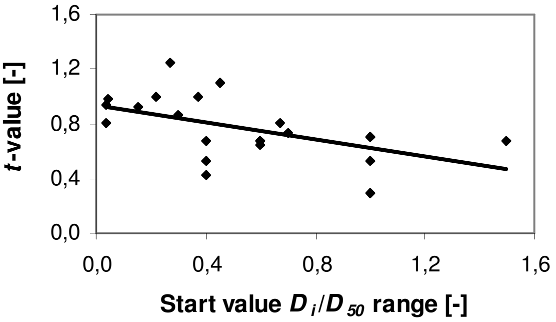

Figure 12. The influence of the D_i/D_{50}-range on the t-values (data taken from appendix 1).

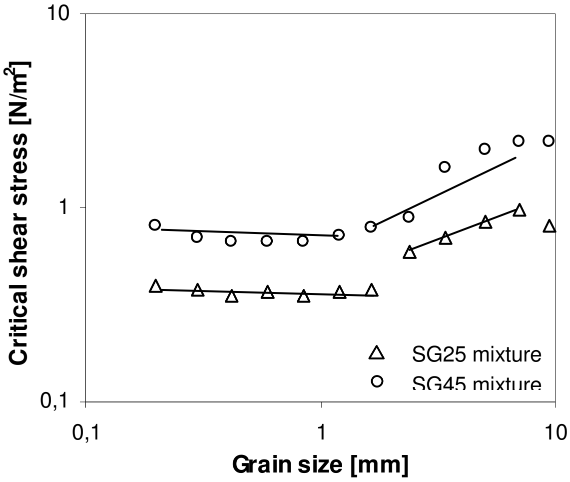

Figure 13. The fitting of two separate power functions through experimental critical shear stress data (after: Kuhnle 1993).

The t-values are also affected by the definition of beginning particle motion (Diplas 1987 *op. cit.* Komar & Shih 1992). Most of the definitions used correspond to a small (and often measurable) transport rate, but the magnitude of this minimum transport varies (box 3.1). For instance, the reference transport used by Parker *et al.* (1982, Parker 1990) corresponds to a higher transport rate than the Shields criterion. Notable is the fact that the reference transport often is a function of the shear stress or the grain size (see box 3.1). This can lead to the strange situation that, when equal volumes (expressed as q_i/f_i) of two size fractions are in motion, the one is considered as being mobile, while the other (the coarser one) is still considered immobile.

For the median grain size (D_{50}) needed in equation 3.5, either the value corresponding to the bed surface can be used, or the value corresponding to the bed subsurface. Because it is the conditions at the bed surface which determine whether a grain will be entrained or not, the bed surface value is the correct one (Buffington & Montgomery 1997, Wilcock & McArdell 1997). In many cases however, the subsurface value is used (*e.g.* Parker & Klingeman 1982). In those cases, the t-value not only reflects the effect of hiding-exposure (which is a grain-scale process, acting at the bed surface), but also the effects of pavement formation. Pavement formation is a vertical sorting process, which concentrates coarse grains at the bed

$$W_i = \frac{q_i}{f_i}\frac{1}{\tau^{1.5}} \quad (\rho_s - \rho)\rho^{0.5} \quad = \quad 0.002$$

e.g. Parker *et al.* 1982, Parker 1990

$$W_i = \frac{q_i}{f_i}\frac{1}{\tau^{1.5}} \quad \frac{(\rho_s - \rho)\rho^{0.5}}{\rho_s} \quad = \quad 0.002$$

e.g. Kuhnle 1993, Wilcock & McArdell 1993

$$q_i^* = \frac{q_i}{f_i}\frac{1}{D_i^{1.5}} \quad \frac{\rho^{0.5}}{(\rho_s - \rho)^{0.5}\rho_s g^{0.5}} \quad = \quad 10^{-4}$$

e.g. Shvidchenko 2001

Box 1 Expressions for the reference transport rate. With: W_i and q^*_i dimensionless transport rates [-], q_i the transport rate of size fraction i [m^3/s] and f_i the frequency of the ith size fraction in the bed [-].

surface, so promoting the transport of those grains. It has an effect similar to hiding-exposure, reducing differences in mobility between coarse and fine grains (see paragraph 3.4.1), so it is obvious that the use of the subsurface D_{50} will result in *t*-values closer to –1 than the use the surface D_{50}. For example Parker found a *t*-value of 0.90 for Oak Creek when using the bed surface D_{50}, and a *t*-value of 0.98 when using the subsurface D_{50} (Parker 1990, Parker *et al.* 1982).

Experimental conditions can have an important influence on *t*-values because the critical bed shear stress is not only dependent on the ratio D_i/D_{50}, but also on a large range of other factors. An exact determination of the D_i/D_{50} influence on the critical bed shear stress requires a correction for the influence of those other factors on the bed shear stress. In many studies however, these factors have not been taken into account. Experimental conditions that can influence the value of the critical bed shear stress and therefore the *t*-value are bed form characteristics, transverse and longitudinal bed slopes, amount of cohesive bed material, particle shape and particle density (see Fernandez Luque & Van Beek 1976,Van Rijn 1993, Gomez 1994, Lewin & Brewer 2002).

3.1.6 Entrainment selectivity during partial transport

Wilcock & McArdell (1993, 1997) observed that at a given shear stress a grain size fraction may be composed of two populations: grains that move with some measurable regularity and grains that remain immobile, even though they are exposed on the bed surface. This situation is called *partial transport*. The concept of partial transport emanated from recirculating flume experiments. In sediment feed flume experiments, no partial transport can be maintained under equilibrium situations, since the material that passes though the flume must be identical to that entering it (Parker & Wilcock 1993). However, under non-equilibrium situations partial transport will also occur in sediment feed flumes. The same accounts for natural rivers.

During a discharge wave, initially all grains in a grain size fractions will be immobile. Then, at initial motion conditions, the first few grains in a grain size fraction become mobile. The remaining surface grains are still immobile (partial transport). These grains gradually become mobile when the bed shear stress increases (*e.g.* Fernandez Luque & Van Beek 1976). Just like the entrainment of the first grains in a fraction, the mobilisation of these remaining grains is affected by hiding-exposure effects, but the severity of the hiding-exposure effects is not necessarily the same. Furthermore, the mobilisation of the remaining grains is not only caused by the flow drag, but also by grain collisions.

In deriving their transport relation, Parker *et al.* (1982) assumed that the proportion of mobile grains increases equally fast for fine grain size fractions as for coarse grain size fractions. Their field data, nevertheless, did not justify this assumption (Komar & Shih 1992). From Wilcock & McArdell (1993, 1997) it can be concluded that the proportion of active grains increases faster for fine than for coarse size fractions. In combination with the lower critical bed shear stress for finer grains, this means that fine fractions reach a state of fully mobilized transport (*i.e.* all grains are mobile) at much lower bed shear stresses than coarse fractions (figure 14).

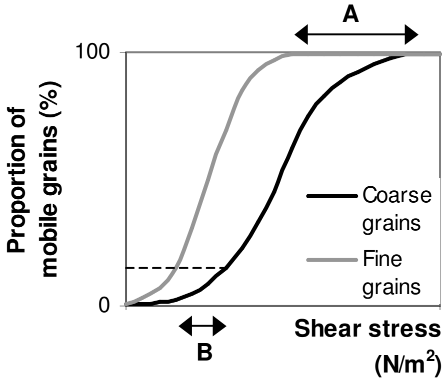

Figure 14. Increase of the proportion of mobile grains for two grain-size fractions. Indicated are: A, the shear stress range in which fine grains are fully mobilized while coarse grains are partially transported, and B, the range in critical shear stress.

So there is a range of shear stresses in which fine fractions are fully mobilised, while coarse size fractions are only partially transported (*indicated with* 'A' *in figure 14*). This is a form of selective entrainment that has hardly been studied, but probably is more important than the exhaustively studied selectivity during initial motion (Wilcock & McArdell 1997, Wilcock & Ellis 1989).

3.2 Selective movement

When a grain has become active in a discharge event, it will travel downstream. The process of grain movement is called selective if coarse grains travel a shorter distance in the same time period than fine grains, provided that both the fine and the coarse grains have already been entrained (Wilcock & Ellis 1989).

First size-selectivity during bed load transport will be discussed; then attention is focused on size-selectivity in combinations of bed load and suspended load transport.

3.2.1 Bed load transport

Surprisingly, in literature no studies were found that address the selectivity of the bed load grain movement process directly. Information about the degree of selective grain movement during bed load transport thus has to be derived from substitute data like the velocity of movement, the mean step length and the distance of travel during a flood period. Existing bed load transport formulae can also provide insight in the selectivity of the grain movement process.

Velocity of movement
The actual velocity of moving grains has been determined in laboratory flumes with a variety of methods. In some studies the velocity of grains in two unimodal sediment mixtures with different sizes were compared (*e.g.* Gilbert 1914, Meland & Norrman 1966, Fernandez Luque & Van Beek 1976). Other studies measured the velocity of solitary grains moving over a fixed bed (*e.g.* Steidmann 1982), or followed painted grains in non-uniform sediment mixtures (*e.g.* Wilcock & McArdell 1993).

Despite the broad range of experimental conditions, the results are quite similar. Generally, coarse grains appear to have a larger velocity of movement than small grains, both in sand mixtures and in gravel mixtures.

This does not necessarily imply, however, that coarse grains are preferentially transported downstream. Even when a grain-size fraction is fully mobilised, the grains inside that fraction are not constantly in motion, but alternate periods of motion (*steps*) with periods of rest (Wilcock & McArdell 1993; Habersack 2001). If the time fraction that a grain is actually moving is smaller for coarse grains, which is probably the case, the movement process can still be size-independent, despite the larger velocity for coarse grains.

Step length
The length of individual grain steps in gravel bed rivers has been determined by marking particles and following them during transport. Probably only grain steps that are not induced by grain-grain collisions were studied. Most step length experiments were conducted in laboratory flumes with uniform sediment (*e.g.* Einstein 1950, Lee & Hsu 1994). Step lengths in non-uniform sediment mixtures, however, can be totally different, because hiding-exposure effects not only act during grain entrainment, but also during grain movement (Samaga *et al.*1986a,b). Only a few step length studies considered non-uniform sediment (*e.g.* Wilcock 1997, Habersack 2001).

A compilation of some step length data is shown in figure 15. In most studies it was observed that coarse particles have larger step lengths, often with a direct proportionality between grain size and step length. Only Habersack (2001), who reanalysed some existing data about step lengths in gravel bed rivers, found a different relation. His results show that grain size fractions smaller than the median grain size all have similar step lengths, while step lengths for coarser particles are much smaller. An explanation for this contradictory result was not given.

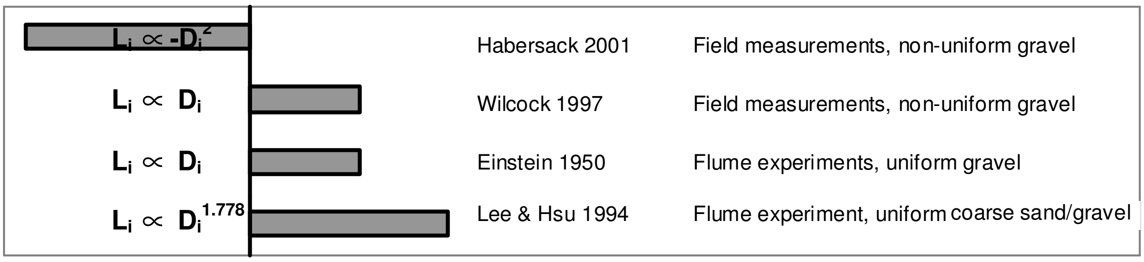

Figure 15 The relation between step length (L_i) and particle size (D_i).

The results shown in figure 15 are only valid for gravel-size particles in gravel bed rivers with a plane bed. So it is possible that sand-size particles in gravel bed rivers have a larger step length than the gravel particles. This is unlikely, however, because sand grains have a high probability of getting entrapped in pores between the surface gravels (*e.g.* Diplas & Parker 1992). It is also possible that sand bed rivers show a different relation between step length and grain size than suggested by figure 15. Grain steps in sand bed rivers do not occur because grains settle at the leeside of immobile roughness grains like in gravel bed rivers, but occur because grains settle at the leeside of bed forms. Coarse grains always settle at the leeside of bed forms, but fine grains can bypass the troughs, traveling in semi-suspension (see Wilbers 2004 for an discussion of bed load transport over dunes). This will result in a decrease in step length with grain size. In strongly bimodal mixtures, however, the opposite seems to be the case. Wilcock & McArdell (1993) observed that coarse grains traverse the body of the dunes and the intervening dune troughs without stopping, while smaller grains tend to deposit temporarily in the dune troughs.

From the above observations in gravel bed rivers and sand bed rivers, it cannot be concluded unambiguously whether coarse grains or fine grains have larger step length, though a small tendency towards larger step lengths for coarser grains is suggested. This does not necessarily imply selective movement of coarse grains, however. The selectivity of bed load movement does not only depend upon the step length, but also upon the number (or preferably *mass*) of grains that start a grain step per second. This is usually called the pick-up rate, which is a confusing term, because it is also used for the taking of grains into suspension (*e.g.* García & Parker 1991, Admiraal & García 1999).

Experiments to determine the pick-up rate were conducted by *e.g.* Van Rijn (1984c) and Fernandez Luque & Van Beek (1976) using uniform sand and fine gravel sediments. They both found the pick-up rate decreasing with grain size. Possibly this balances the increased step length for coarser grains, resulting in non-selective grain movement. However, an accurate evaluation of the size-selectivity of the grain movement process requires simultaneous measurements of the step length and the pickup rate for non-uniform sediments, both for situations with and without bed forms and both for gravel bed rivers and for sand bed rivers.

Distance of travel

The distance travelled by grains during a flood event has been determined in many field studies using tracer grains. In most of the studies no significant relation was found between the travel distance and the grain size (*e.g.* Marion & Weirich 2003), which can be due to methodological shortcomings (Hassan *et al.* 1992). Hassan & Church (1992) reanalysed travel distance data from a variety of gravel bed rivers. All data probably refer to plane bed situations and a low intensity transport regime, which means that individual grain movements are isolated events. They found that the travel distance is primarily determined by the flow intensity. The grain size of the moving particles, however, also had a clear influence (see also Church & Hassan 1992). This is illustrated in figure 16.

Figure 16 shows that grains smaller than the median grain size all move a similar distance. A possible explanation is that small grains have a high probability of getting trapped, making the travel distance relatively insensitive to grain size.

For grain size fractions larger than the median grain size, a rapid decrease in the travel distance with grain size is observed. This seems contradictory with the results from the step length and grain velocity experiments, but figure 16 should be handled with care. The data used stem from flood events and it is probable that coarse grains have started moving later during the event than fine grains. The plotted travel distance for coarse grains thus has been covered in a shorter time period. So the entire movement process will be much less size-selective as suggested by figure 16. A meaningful way to evaluate the resulting degree of size selectivity would be to divide the to divide the travel distance of a grain by the time period that that grain was mobile.

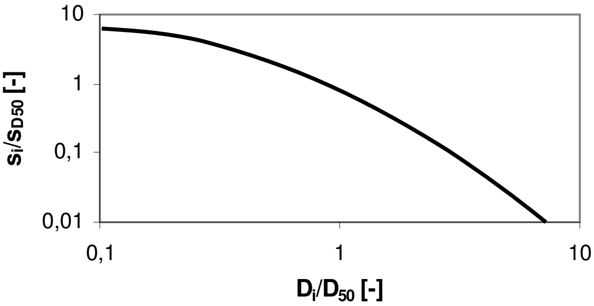

Figure 16. The influence of the grain size on the distance travelled during a flood event. The latter is expressed as the distance moved by grains of size *i* (s_i), scaled by the distance moved by grains which size is equal to the median grain size (s_{D50}). After Church & Hassan (1992).

Bed load transport formulae

From the preceding sections it has not become clear whether the bed load movement process is selective or not. There are some indications that coarse grains have a larger velocity of travel and a larger step length, but this is counteracted by the fact that they also have a smaller pick-up rate. Other observations show that coarse grains travel shorter distances during a flood event, but this is probably because they are only mobile during a part of the event. The movement of bed load grains thus seems to be only weakly size-selective, or even non-selective.

An idea of the size-selectivity during grain movement can also be drawn from existing fractional bed load formulae. Many formulae compute the fractional transport rate as $q_i = f(\tau - \tau_{c,i})$, for example the Meyer-Peter & Müller (1948) formula:

$$q_i = f_i\left[\frac{8}{g(\rho_s - \rho)\rho^{0.5}}\right](\tau - \tau_{c,i})^{1.5} \qquad 3.7$$

If the shear stress τ is large, the influence of $\tau_{c,i}$ becomes negligible, and for each size fraction the same transport rate will be computed, provided their availability in the bed is the same. These are conditions in which all size fractions are fully mobile, so it is implicitly assumed that the process of grain movement is not size-selective. Because sediment transport formulae like equation 3.7 have been used relatively successfully, the assumption of non-selective grain movement cannot be far beside the truth.

3.2.2 Suspended load transport

In the preceding paragraphs it was assumed that sediment transport only involves the transport of particles over the bed by rolling, sliding and saltating, commonly called bed load transport. When bed shear stress is high enough particles will become suspended, the immersed particle weight being supported by turbulent diffusion.

Usually a part of the total sediment load is transported as bed load, while the remaining part is transported in suspension. Because suspended sediment particles move continuously and with a much larger velocity than bed load particles, suspended grain movement is much faster than bed load grain movement.

The existence of suspended load transport can be a primary cause for size-selective grain movement if finer particles predominantly move in suspension, while coarser particles predominantly move as bed load. Size-selectivity can also result from the suspended transport process itself: when coarser suspended particles travel with a lower velocity than finer particles. Both causes will be discussed successively. Doing so, only attention will be attained to grain sizes that show interaction with the river bed somewhere during the course of the river, for suspended sediment that is never deposited on the river bed will not influence downstream fining. This does not mean that all wash load will be left out of consideration, because grains that are transported in suspension can become part of the bed load transport further downstream (*e.g.* Deigaard 1980).

Selective movement by differences between bed load and suspended load size distributions

A logical first step to determine the size-difference between the bed load and suspended load is to evaluate the critical conditions for the beginning of suspension. This was done by several authors (see Cheng & Chiew 1999 and Van Rijn 1984b for an overview). Most of them expressed the critical condition for the beginning of suspension in terms of a critical ratio u^*/w_i, where u^* represents the shear velocity [m/s] and w_i the fall velocity of grains of size i. In figure 17 these criteria for the beginning of suspension are plotted in a shear stress-grain size plot.

It can be seen that all criteria predict a strong increase in the critical shear stress for the beginning of suspension with grain size. Nevertheless, there are marked differences between the curves. This is

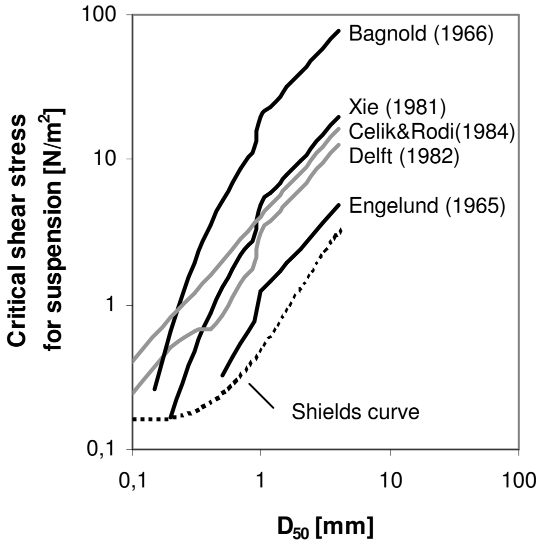

Figure 17. Critical shear stress criteria for the beginning of suspension in uniform sediments. (Curves constructed by combining the suspension criteria (*op. cit.* Cheng & Chiew 1999, Van Rijn 1984b) with the fall velocity relations given by Van Rijn 1993.)

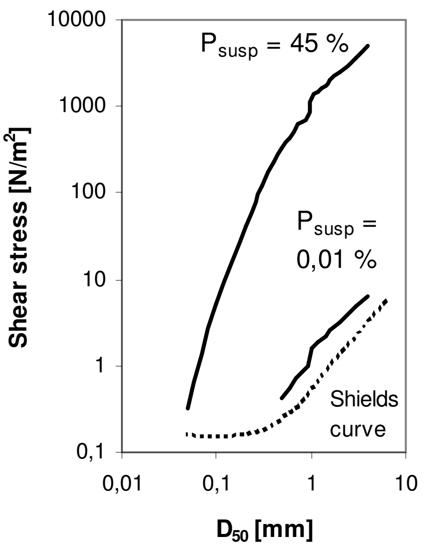

Figure 18. Probability of suspension in uniform sediments as function of the shear stress and the median grain size (after: Cheng & Chiew 1999)

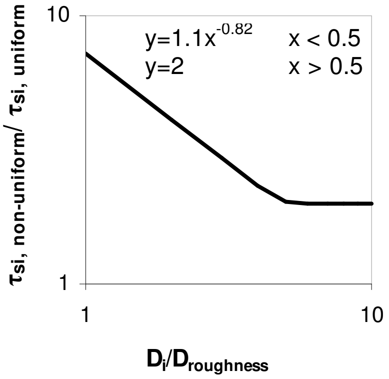

Figure 19. Hiding exposure function for suspended sediment (Niño *et al.* 2003). Symbols used are: $\tau_{si,\ non\text{-}uniform}$ and $\tau_{si,\ uniform}$, respectively the critical shear stress for the beginning of suspension of size fraction *i* in a non-uniform mixture and in a uniform mixture, and D_i and $D_{roughness}$, repesctively the mean grain size of fraction *i* and the size of roughness grains (~D_{90}).

because all criteria point to different probabilities of suspension, as was shown by Cheng & Chiew (1999). The Bagnold criterion corresponds to the largest probability of suspension and Engelund's criterion to the lowest probability.

Based on Cheng & Chiew's equations figure 18 was constructed, in which two probability relations are given. Though figure 18 is only based on theoretical relations, it much more instructive than figure 17. It shows that while a state of 0.01 probability of suspension (~beginning of suspension) can be reached by many grain sizes, a state of 45 % probability of suspension can only be reached by very small grains. Or alternatively, bed grain size fractions up to 5 mm can become suspended for 0.01 %, but only grain sizes smaller than about 0.5 millimetres can become suspended for 45 % or more.

These results are only valid for uniform sediment mixtures. In non-uniform mixtures, the curves shown in figure 18 will be markedly flatter due to hiding-exposure effects, as was shown by Niño *et al.* (2003, figure 19, *see also* Samaga *et al.* 1986). That does not alter the fact that figure 18 suggests that the difference in size composition between bed load and suspended load will increase at increasing bed shear stress. At increasing shear stress the number of coarse grains that will become suspended only increases marginally, while the number of suspended fine grains strongly increases. The bed load becomes thus becomes rapidly depleted of fine grains, which will even lead to pavement formation in some cases. The suspended load will then be much finer than the composition of the bed load, resulting in strongly selective transport. The selectivity of the movement process thus increases with shear stress.

A final remark that has to be made here is that the source for the suspended load is the material that is present in the top of the bed load layer. The composition of this material is not necessarily equal to the average size composition of the bed load layer, introducing an additional difference between the bed load and suspended load composition, which is not incorporated in figure 19, because this is based upon experiments in which a bed load layer was absent. When the river bed is covered with bed forms, this will even be more complicated, because suspended sediment than primarily stems from the top of the bed forms which usually are much finer than the average bed load.

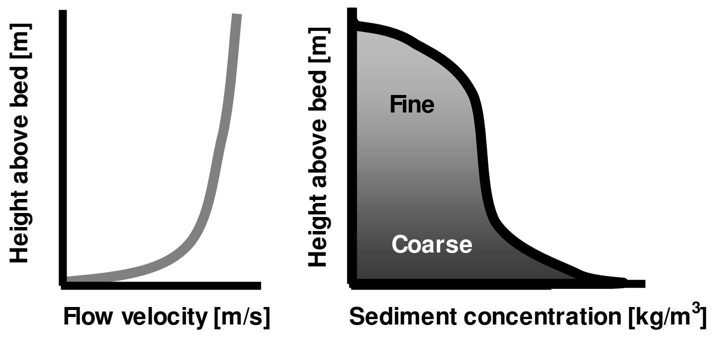

Figure 20. Schematic representation of the suspended load transport process. Coarse grains can be seen to have a much lower average transport velocity, resulting in selective transport of fine fractions.

Selective movement resulting from the suspended load transport itself
Size-selectivity in the suspended load transport itself is the result of the large settling velocity of coarse suspended grains. Therefore they are only present in the lowermost part of the suspended load layer, while fine grains are present throughout the suspended load layer. Because the flow velocity in the lower part of the suspended load layer is relatively low, the average velocity of coarse suspended grains is smaller than the average velocity of fine suspended grains, resulting in selective movement (figure 20). While this mechanism is well known (*e.g.* Deigaard 1980), in literature no studies were found in which the exact degree of size-selectivity involved was determined.

3.3 Selective deposition

Depositional dynamics have received considerably less attention than those of sediment entrainment or sediment movement, but can also be strongly size-selective. Selective deposition is defined here as the situation in which coarse and fine grains are deposited at different shear stresses.

Most authors assume that the deposition process is strongly related to the process of grain entrainment (Powell 1998): coarse grains being deposited at higher shear stresses than small grains, because they also are entrained at higher shear stresses than small grains. The critical shear stress at which a grain will settle however is usually somewhat smaller than the critical shear stress at which it was entrained (*e.g.* Beschta 1987). According to Reid & Frostick (1987) this is partly caused by the difference between static friction and dynamic friction and partly the result of bed microstructures that hamper grain entrainment, but do not affect grain deposition.

An opposite view is given by Frostick & Reid (1980) and Paola *et al.* (1992b). They stated that larger grains protrude farther into the flow and roll over a surface that is relatively smooth, whereas finer grains tend to become entrapped among the larger grains, resulting in selective deposition of fine grains (*c.f.* Diplas & Parker 1992). The degree to which fine grains are preferentially deposited depends primarily upon the bed roughness. The coarser the bed surface, the more the deposition of coarse grains is encouraged and the smaller the likelihood of finer particles settling because of the increased turbulence around the coarse grains (see Powell 1998, Diplas & Parker 1992).

3.4 Size-selective transport due to local sorting processes

3.4.1 Armouring

A common feature in many (sand-) gravel bed rivers is the presence of an armour layer: a thin layer of coarse grains on top of finer material. Two types of armour layers can be distinguished: stable armour layers and dynamic armour layers.

A stable armour layer develops when the bed shear stress is smaller than the critical bed shear stress for the coarsest grains on the bed surface, but larger than the critical bed shear stress for the smallest grains on the bed surface. Fine grains are washed away, while coarse grains remain behind, causing the bed surface to coarsen (*e.g.* Sutherland 1987, Lisle & Hilton 1999). The remaining fine grains on the bed surface become increasingly hided by the coarser grains, and eventually a situation is reached in which none of the surface grains can be moved by the flow. A stable armour layer has developed and the sediment transport rate is very small. When the bed shear stress increases, the armour layer coarsens because increasingly coarse material is winnowed away from the bed (*e.g.* Gomez 1994). The armour layer is suddenly broken up when the bed shear stress becomes equal to the critical value for the armour grains, which is dependent on their size and shape (Gomez 1994). Armour grains *and* underlying finer grains then start to move at the same shear stress, which means that the entrainment of the first grains in a size-fraction is a non-selective process.

The presence of a stable armour layer, however, also can enhance differences in mobility. Generally not all of the fine material is present underneath the armour layer. A part of it moves quickly over the armour layer, in the form of sand waves or dunes, indicating strongly selective transport (Lisle & Hilton 1999, Ryan *et al.* 2002, Kleinhans *et al.* 2002 and Frings 2002, 2003). Furthermore, it must be realised that the formation of an armour layer always involves selective transport (*c.f.* Church *et al.* 1991), something which is generally overlooked.

Dynamic armour layers (*pavements*) form after the stable armour layer has been broken up and there is a continuous supply of sediment from upstream. In this situation, all grain size fractions are mobile, but the mobility of coarse size fractions is still smaller than the mobility of fine fractions. A pavement generally becomes finer when the bed shear stress increases and disappears at high bed shear stresses (*e.g.* Gomez 1995). This is because the mobility differences between coarse and fine grains disappear at increasing bed shear stress, and because the sediment in the pavement becomes mixed with finer underlying material at increasing bed shear stress.

The influence of a pavement on the degree of size selectivity during sediment transport is complex. In developing bed load transport relations, Parker and co-workers hypothesised that the existence of a bed pavement regulates the entrainment of particles by the stream, resulting in a situation in which all grain sizes are entrained at the same bed shear stress and transported at rates in proportion to their presence in the subsurface material (Parker *et al.* 1982, Parker & Klingeman 1982). This is called the equal mobility hypothesis. The proposed mechanism is that the differences in mobility between coarse grains and fine grains which are not effaced by the hiding and exposure effects (§ 3.1.3) are cancelled out by the overrepresentation of coarse grains at the bed surface, making them more available to the flow.

The effect of a pavement on the degree of size-selective transport can be evaluated by comparing the empirical hiding-exposure coefficient (*t*-value) calculated using the median grain size of the bed surface, with the *t*-value calculated using the median grain size of the bed subsurface (see (§ 3.1.5).

3.4.2 Patchiness

Parker & Klingeman's (1982) hypothesis of equal mobility conflicts with the downstream fining process. This paradox can be explained by the fact that the hypothesis neglects size-selectivity during the formation of armour layers, as was seen in the previous paragraph, or by assuming small deviations from equal mobility (Paola & Wilcock 1989). Another explanation for this apparent inconsistency is the patchiness-hypothesis.

In many rivers the bed load composition is not constant spatially, but sorted in coarser and finer patches. Patches can either be organized perpendicular to the flow (Iseya & Ikeda 1987, Dietrich *et al.* 1989) or parallel to the flow (Wilcock & McArdell 1993, Sambrook Smith & Ferguson 1996) and can result from various mechanisms (*e.g.* Paola & Seal 1995, Lisle 1995, Kleinhans 2002, Toro-Escobar *et al.* 2000). A special case of patchiness is the sorting process in meander bends.

Even if equal mobility is satisfied exactly within a patch, the total sediment transport process can be size-selective when coarse patches move less frequent and less fast than finer patches. Whether this is the case depends on the spatial correlation between actual and critical bed shear stress. If the actual and critical bed shear stress are uncorrelated, coarse patches will always be less mobile because they have a higher critical bed shear stress. If a coarse patch develops because of a locally high stream velocity, the increase in critical bed shear stress could theoretically be exactly compensated by the increase in actual bed shear stress, making coarse patches equally mobile as fine patches (Paola & Seal 1995, *see also* Ferguson 2003).. This is not always the case however. Frings (2002) and Duizendstra (2001) found much lower transport rates in the coarse parts of the river Meuse than in the fine parts, despite the larger stream velocity in the coarse parts.

When a fine patch has a larger velocity than a coarse patch, it will ride over or into the downstream coarse patch, if patches are organized perpendicular to the flow. According to Lisle & Hilton (1999) fine patches override immobile coarse patches, but Iseya & Ikeda (1987) describe a mechanism by which coarse patches gradually change in fine patches and reverse, the pattern of patchiness slowly moving downstream.

Paola & Seal (1995, Seal & Paola 1995) applied the patchiness hypothesis to the North Fork Toutle River. They found that most of the observed downstream fining can be explained using patchiness as dominant mechanism. Sambrook Smith and Ferguson (1996) argued that the presence of patches is one of the primary reasons for the abruptness of the gravel-sand transition. In gravel bed rivers with a gradual reduction in stream gradient there will be a location where sandy patches form on the gravel bed. Patches increase the size-selectivity of the transport process because fine patches travel faster than coarse patches. Patches however also cause a decrease of the bed roughness, which leads to a further reduction in bed shear stress, making the gravel more immobile.

3.4.3 Bed form sorting

The presence of bed forms can have an effect opposite to the pavement-mechanism described by Parker *et al.* (1982). The migration of bed forms leads to a vertical sorting in the channel bed through two processes: sorting in the bed form troughs and sorting at the steep lee-side slopes.

In bed form troughs an accumulation of coarse grains occurs, due to preferential deposition of the coarse fractions (Kleinhans 2001, 2002) and winnowing away of the finer fractions in situations of partial transport (Blom 2003). Whenever an armour layer is present, the winnowing away of fines in the troughs will cause the armour layer to sink and become buried (*e.g.* Willis 1988 *op. cit.* Blom 2003).

Sorting at the lee-side slopes can occur when coarse grains roll further down the lee face than fine ones, because they have larger velocities, experience less friction (relatively) and are less probable to find a suitable pore in which they can become entrapped (Zanke 1976 *op. cit.* Blom 2003, Ribberink 1987. Kleinhans (2001, 2002) proposes another mechanism, explaining sorting at the lee-side slope in three phases. First, the grains fall form suspension at the top of the slope. Second, small grains on the leeside slope are worked down by kinematic sorting and percolation, while coarse grains are worked up. Third, the sediment on the lee slope flows downstream as a grain flow in which preferentially large grains are dragged downstream.

Blom (2003) estimated the time needed for a bed layer to reach its equilibrium composition under conditions of a constant average bed level, a constant average bed load transport and a constant probability distribution of bed form trough levels. She found that the uppermost layers reach their equilibrium composition before the bed forms have moved one dune length. The lowermost layers however, required much more time to reach their equilibrium composition. In nature these lowermost layers probably never will reach their equilibrium composition, due to the continuously changing hydraulic conditions.

The resulting fining upward profile extends from the top to the bottom of the transport layer, which is about twice the bed form height. Because bed form height generally increases with discharge, during low flow periods only the upper part of the fining upward profile established during floods will be re-sorted and transported downstream. The bed load composition during low flows will thus be finer than the bed load composition during high flows. Because low flow conditions occur much more frequent, coarse material will generally be underrepresented in the bed load, causing a distinct difference between transport rates of coarse and fine material (Deigaard 1980, Frings & Kleinhans 2002).

This bed form sorting mechanism can be considered at a special case of the topographic sorting mechanism described by Paola (1989). Topographic sorting that results from the tendency of coarser grains to be concentrated preferentially on topographic low surfaces. These surfaces are buried a greater-than-average proportion of time, so coarse material will generally be underrepresented in the bed load.

3.5 Comparison of bed and bed load grain size distributions

In paragraphs 3.1 to 3.3 the entrainment process, the movement process and the depositional process were studied separately. These processes however interact in a complex way and are affected by bed armouring, patchiness and bed forms (§3.4). A way to evaluate the resulting degree of size-selectivity is to follow tracer particles in their downstream transport over several years (*e.g.* Ferguson *et al.* 1996). This method however requires a lot of field measurements. A good alternative for rivers without suspended load is the comparison of bed (subsurface) and bed load grain size distributions. Any difference between bed load composition and bed composition indicates size selective transport. (For rivers with a combination of bed load and suspended load transport a comparison could be made of total transport load and bed load composition, but this is prone to errors because a part of the suspended load has no relation with the in situ bed composition, *see* Beschta 1987).

Many authors measured bed load grain size distributions and compared them to bed size distributions (*e.g.* Parker *et al.* 1982, Church *et al.* 1991, Kuhnle 1992, Wathen *et al.* 1995, Lenzi *et al.* 1999). They all noticed a coarsening of the bed load at increasing bed shear stresses, the bed load composition generally approaching the bed composition at the highest bed shear stresses. This indicates a tendency toward equal mobility at high bed shear stresses, but a strong degree of size-selectivity at low bed shear stresses; the latter probably resulting from size-selective entrainment (especially in the partial transport range), possibly in combination with sorting processes.

The exact shear stress value at which equal transport mobility is reached varies however. Parker *et al.* (1982) report equal mobility as the shear stress is larger than 1.4 times the critical shear stress for incipient motion, while Lenzi *et al.* (1999) reported that equal mobility is attained when the discharge is 2.5 times larger than the critical discharge for incipient motion. Thorne stated in his discussion to Komar & Shih (1992) that in many (gravel-bed) rivers the actual shear stress will never be twice the shear stress for incipient motion, so equal mobility will never be attained.

From the viewpoint of downstream fining these temporal variations in the degree of size-selectivity are relatively unimportant. What counts is the long-term averaged degree of size-selectivity, which combines all effects of selective entrainment, selective movement, selective deposition and selectivity due to sorting processes. The long-term averaged degree of size-selectivity can be determined by comparing the long-term averaged bed load size distribution with the bed (subsurface) size distribution.

Wathen *et al.* (1995) found the yearly averaged bed load size distribution to be much finer than the bed size distribution, but Kuhnle (1992) noted that the bed load size distribution was equal to that of the bed, when averaged over a four-year period. So according to Kuhnle (1992) all grains are on average equally mobile. One of the most detailed researches was carried out by Church *et al.* (1991) who performed bed load transport measurements in Harris Creek. They computed $f_{i.bl}/f_i$ - ratios for different transport conditions, in which $f_{i,bl}$ the frequency of the i^{th} grain size fraction in the bed-load [-] and f_i the frequency of the i^{th} grain size fraction in the bed subsurface. Their results are shown in table 3.

		Low flow	Flood	Season-average
Grains	**< 0.053 mm**	>>1	± 1	> 1
Grains	**0.053 - 0.425 mm**	>1	< 1	= 1
Grains	**> 0.425 mm**	<1	= 1	< 1

Table 2. $f_{i,bl}/f_i$- values in Harris Creek Church *et al.* (1991)

From the table it can be concluded that intermediate grains (0.05-0.43 mm) are overrepresented during low flows ($f_{i.bl}/f_i > 1$), but underrepresented during flood periods ($f_{i.bl}/f_i < 1$), probably because a part of these grains are taken in suspension or trapped in the bed (Gomez 1995). Averaged over a year these grains exhibit equal mobility ($f_{i.bl}/f_i = 1$), which is in accordance with the finding of Kuhnle (1992b). Large grains however appear to be always underrepresented in the bed load, except for the largest floods. Consequently these grains are almost always size-selectively transported. The smallest grains are largely overrepresented in the bed load during the largest part of the year. These grains must be considered as wash load and do not have a relation with the bed composition.

Lisle (1995) studied spatial variations in $f_{i.bl}/f_i$-ratios in thirteen rivers and found $f_{i.bl}/f_i$-ratios to be close to unity in downstream river reaches, but much smaller than unity in upstream river reaches. Upstream river reaches thus are characterized by selective transport. Because these river reaches have armour layers, which are seldom broken down, the fine material probably moves in patches over the armour layer (Lisle 1995).

3.6 Synthesis

It was seen in this chapter that many empirical studies have been carried out to determine the size-selectivity of the sediment entrainment process. They all focused on size-selectivity during incipient motion; the size selectivity during conditions of partial transport received much less attention. The experiments have suggested strongly selective entrainment during incipient motion in some rivers, but equal entrainment mobility in others. This is partly due to differences in the experimental methods. The observed differences in size-selectivity, however, also partly result from differences in sediment and bed properties between the study areas. The selectivity of the entrainment process was found to increase with sediment bimodality and is most pronounced for grain size fractions coarser than the median grain size.

The size-selectivity of the grain movement process depends on the manner of movement. Bed load grain movement seems to be rather non-selective, though this conclusion was based on incomplete observations of the movement process, primarily performed in plane gravel bed rivers and flumes. Suspended load grain movement, however, is a size-selective process. Coarse grains are only present in the slow-moving lowermost part of the suspended load layer, whereas fine grains are present throughout the suspended load layer, thus having a larger average movement rate. Size-selective grain movement also results from the division between bed load transport and suspended load transport, for the fast-moving suspended load transport has a much fine size-composition than the slow-moving bed load transport. The latter form of size-selective movement will increase at increasing discharge, because the percent of the bed grains that is suspended increases much faster with shear stress for fine grains than for coarse grains.

Information about the selectivity of the depositional process is very scarce. It is usually assumed that coarse grains will settle at higher bed shear stresses than fine grains, but the opposite can also occur.

The degree of size-selectivity due to local sediment sorting was found to depend on the type of sorting process. Armouring often leads to equal mobility, because coarse grains are overrepresented at the bed surface. Dune sorting leads to size-selective transport, because coarse grains are concentrated in deep bed layers, which are only rarely mobile. Patchiness also leads to size-selective transport, because of the difference in mobility between coarse and fine patches.

From the above overview it can be concluded that in most rivers the overall sediment transport process will be size-selective, either resulting from selective entrainment, selective movement, selective deposition, or from selectivity due to sorting processes. An important issue that has to be addressed now is the question how exactly these processes cause downstream fining.

The influence of selective movement on downstream fining depends on the mode of grain movement. If all sediment travels as bed load, while only the bed load grain movement process is selective, all bed load grains are mobile. However, the finer part of the sediment load that is supplied at the upstream end of a channel will quickly run down the river, while the coarser part of the supplied load will only travel downstream slowly. Fine and coarse grain sizes thus become separated and a downstream fining pattern develops. This is a temporal situation, however. Because coarse grains also travel downstream, eventually a situation will be reached in which the bed composition in the entire river is equal to the input sediment load composition (*c.f.* Van Stralen 1999). Downstream fining by selective bed load grain movement thus clearly is a non-equilibrium situation. It must be seen as a process that attributes to a rapid downstream fining in the early stages of a river. In mature rivers the influence of selective bed load grain movement is probably small.

The situation becomes different when a part of the sediment travels in suspension, while the rest still travels as bed load. Selective movement now also results from the fact that coarse grains travel slowly in bed load, while fine grains move rapidly in suspension. Most rivers are characterised by a downstream decrease in bed shear stress ($\tau=\rho gRI$), caused by the downstream decrease in stream gradient (I), but slightly counteracted by an increase in water depth (R) (see Rice & Church 2001, Knighton 1999b, Morris & Williams 1997, 1999b, Ohmori 1991). This indicates that there is a gradual decrease in the size of sediment particles that can travel in suspension. Or alternatively, each suspended load grain size fraction will encounter a point at which it settles and travels further as bed load. The location at which this occurs shifts downstream for finer particles. This results in a downstream fining pattern that is relatively stable. It will only disappear when the stream gradient becomes constant over the entire river. Whether this will occur in natural rivers is still unknown. Anyhow, changes in stream gradient are very slow (*e.g.* Deigaard 1980), so downstream fining caused by the division between bed load transport and suspended load transport must be seen at least as a semi-equilibrium situation. *However*, this mechanism can only explain downstream fining in sand bed rivers. Because gravel grains never travel in suspension, the downstream fining in gravel bed rivers cannot result from the settling of increasingly finer grains in downstream direction.

The situation changes if only the entrainment process is selective. Selective entrainment means that fine grains are entrained in an early stage of the flood wave, when the bed shear stress is still relatively low, while coarse grains are entrained much later. Therefore, fine grains are in motion for a larger part of the year than coarse grains, and a downstream fining pattern develops because coarse and fine grain sizes become separated, just like in case of selective bed load movement. *However*, while in case of pure selective movement all grain sizes are mobile over the entire reach of the river, this is not the case with selective entrainment. The downstream decrease in bed shear stress that is present in most rivers causes a gradual decrease in the size of sediment particles that can be entrained by the flow. This results in a downstream fining pattern, which covers both the gravely part and the sandy parts of a river, and can be seen as (semi-)equilibrium situation. Selective entrainment thus attributes to a rapid downstream fining in the early stages of a river, but also maintains a relatively stable downstream fining pattern in later stages.

If only the grain deposition process is selective, downstream fining will develop in a way very similar to downstream fining caused by selective entrainment. It also attributes to a rapid downstream fining in the early stages of a river, but can also explain downstream fining patterns in mature rivers. However, because the highest downstream fining rates that have been reported all stem from rapidly aggrading river reaches and alluvial fans, selective deposition is usually considered to be the most important downstream fining mechanism (*e.g.* Bradley *et al.* 1972, Brierley & Hickin 1985, Morris & Williams 1999b). This is not necessarily true. The high downstream fining rates in aggrading rivers can also be brought about by selective entrainment. Sediment transport in rivers primarily takes place during floods. After a flood thus both coarse and fine grains become deposited. Even if this depositional process is non-selective, downstream fining can occur, because selective entrainment can make that the coarse grains are not entrained anymore during the following flood, while fine grains are.

Finally the influence of selectivity due to sorting processes on downstream fining will be considered. Dune sorting involves the rapid downstream transport of fine particles and the slow downstream transport of coarse particles. It thus is similar to selective bed load movement and is not able to produce an equilibrium downstream fining pattern. Patchiness in itself does not cause downstream fining at all. The only effect of patchiness is to increase selective entrainment, movement and deposition processes. The effects of armouring on downstream fining are diverse. Pavement formation often causes equal mobility and prevents downstream fining. Stable armour layers also hinder downstream fining, except when fine grains are present on top of the armour layer. In that case it increases the degree of selective entrainment and attributes to downstream fining.

4. Sediment addition, extraction and redistribution

The introduction of sediment of different origin into a river can obscure the effects of selective transport and abrasion completely. The same accounts for the size-selective extraction of sediment from the main channel. In lowland rivers, the downstream fining patterns are additionally influenced by the sediment redistribution at river bifurcations.

An overview of addition, extraction and redistribution processes is given in the form of a sediment balance is given in figure 21. These processes and their effects on downstream fining are successively discussed in the following nine sections (§4.1-4.8), after which a short overview is given of the changing downstream importance of these factors (§ 4.9).

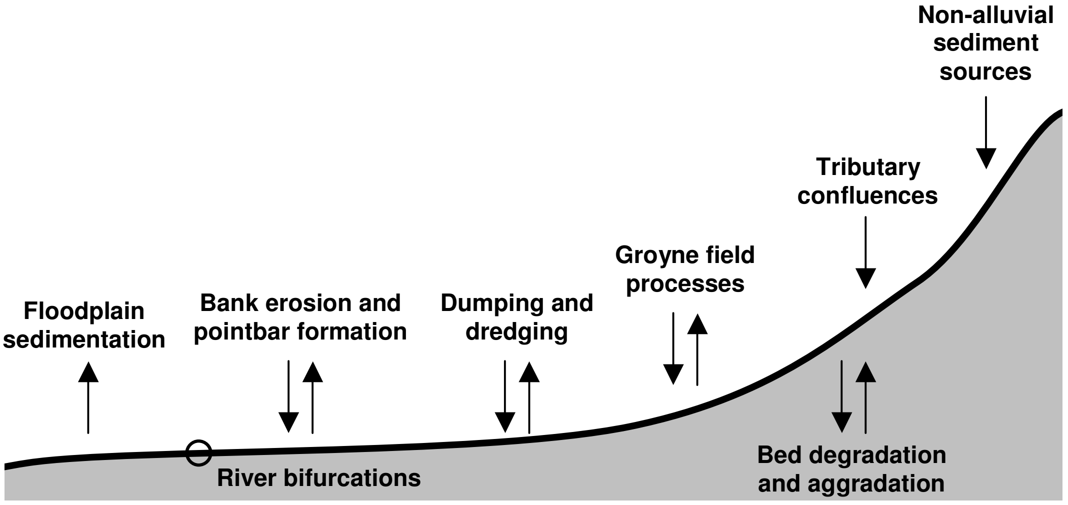

Figure 21. Longitudinal river cross-section, showing the most important sediment addition, extraction and redistribution processes.

4.1 Tributary confluences

Whether or not the main stream downstream fining trend is disrupted at tributary confluences, depends on the relative volume and relative size characteristics of the sediment input (Knighton 1980). In particular, the larger the volume of an input and the greater the grain-size disparity between it and the mainstream material, the greater is the expectation that the mainstream texture is changed significantly. Another important factor is the water discharge of the confluence relative to the water discharge of the main stream. A tributary, which introduces a significant quantity of water but little sediment, could, by increasing bed shear stresses, produce a significant change in mainstream texture (Rice 1998).

Tributaries will only rarely cause a *decrease* in main stream grain size. This can only occur when large quantities of sandy sediment are supplied to a gravel-bed river, which was for instance the case in the Ringarooma river (Knighton 1989, 1999a). When the tributary sand supply is not very large, it is unlikely that it will change the main river grain size. Fine sediments entering a river with coarser material will immediately be washed away, because the relative coarseness of the main stream is indicative of a transport regime capable of removing fine material.

On the other hand, coarse sediment entering a river with fine bed material is often directly deposited because the hydraulic conditions in the main river are not capable of transporting coarse grains. This causes a sudden increase in main stream grain size. Though this is clearly not an equilibrium situation, it is observed frequently (*e.g.* Ichim & Radoane 1990, Brewer & Lewin 1993, Rice & Church 1998). A strong downstream fining usually follows the sudden increase in grain size; especially in short river reaches between two tributaries (sedimentary links). Backwater effects caused by the lowermost tributary lead to a significant reduction in the bed gradient just upstream of this tributary confluence, resulting in a strong decline in bed shear stresses in the sedimentary link (Dawson 1988). Therefore the transport of the sediment delivered by the upstream tributary will be strongly selective.

In large rivers it can occur that tributary confluences only affect the downstream fining trend over a small part of the river width, because the added sediment load does not mix entirely, but travels alongside the bank at which it entered the main river (see Bradley *et al.* 1972, Van Wijngaarden 1999).

Rice & Church (1998) developed a statistical method to identify tributaries with a significant influence on mainstream texture. Only 23 of 156 investigated tributaries (British Columbia) appeared to have a distinct influence on the grain size in the main stream. In his Piave study Surian (2002) also found that most tributaries do not disrupt the downstream fining trend of the main river.

When information about the water and sediment discharge of the tributary is lacking, a prediction of the tributary influence on main stream grain size can be made on basis of catchment characteristics. Rice (1998) reached reasonable results by relating the tributary influence to the catchment area and the product of catchment area and catchment slope, respectively surrogates for sediment production capacity and sediment transport capacity.

4.2 Non-alluvial sediment sources

While tributaries have a very point-wise influence on the main stream downstream fining trend, non-alluvial sediment sources have a more diffuse influence. Non-alluvial sediment sources are especially important in river head waters, where floodplains are small or absent. In these river parts slope processes like creep and mass movements directly deliver sediment to the river (Rice & Church 1996a). Because the supplied material is mostly coarser than the main stream bed load, non alluvial sediment sources usually have a coarsening effect. A different situation occurs when also wood debris are delivered into the river. This often leads to the formation of log jams, with an increase in grain size downward of the jam and an decrease upward of the jam. According to Rice & Church (1996a) non-alluvial sediment supply and storage preclude a systematic diminution of sediment grain size in river head waters.

4.3 Bank erosion and point bar formation

When a (meandering) river is unconstrained by resistant rock or engineering works, the erosion of river banks can form a significant sediment source. The influence on the downstream fining trend again depends on the volume and grain size characteristics of this sediment supply, which can be very variable.

In the meandering process bank erosion often goes aside with point bar formation. In case of point bars without chutes, especially the finer grains are involved in the formation of point bars; this could lead to a coarsening of the bed load, so delaying the downstream fining trend. In case of net sedimentation, however, especially coarse material (the channel lag) is extracted from the sediment load during lateral migration of meander belts (*e.g.* Dietrich 1989, *see* figure 22). This will result in a increase in downstream fining rates.

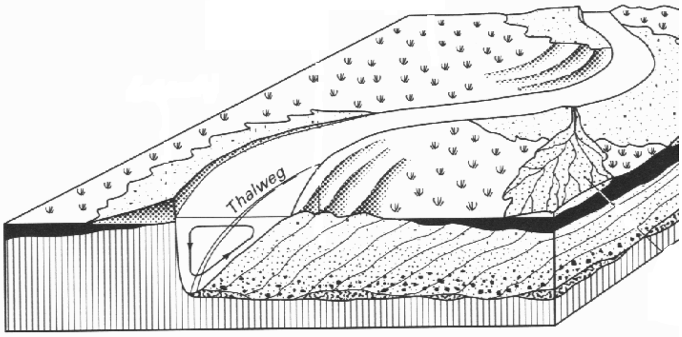

Figure 22. Bank erosion and pointbar formation (after Reineck & Singh 1973). The channel lag deposits (large dots) are permanently removed from the river, in case of net sedimentation, while the finer point bar sediments (specks) can be re-entrained when the meander bends shifts.

4.4 Floodplain sedimentation

During floods a part of the suspended sediment load in the main channel is transferred to the floodplains and deposited. This forms a sediment sink with a very diffuse character.

Mechanisms of overbank deposition

Two processes cause the transfer of suspended sediment to the floodplains: turbulent diffusion and convection. Turbulent diffusion results from the difference in current velocities between the channel and the floodplain and is restricted to a narrow strip close to the channel. Sediment transfer by convection occurs where there is a component of flow perpendicular to the channel, which is especially the case in meander bends (Ten Brinke *et al.* 1998, *cf.* Nicholas & Walling 1998). Convection can distribute sediment over the entire floodplain.

Because of the low stream velocities, sediment suspended in floodplain water will start to settle down. The coarser grain size fractions will settle completely, usually relatively close to the channel, while the finest fractions will only settle in places with ponded water, due to the difference in fall velocity (see § 3.2.2; Nicholas & Walling 1998).

Sources of the suspended sediment load

The suspended sediment stems partly from soil erosion in upstream areas and partly from suspension of local river bed material (figure 23). The suspended sediment stemming from hill slope erosion in upstream areas bears no relation with the in-situ channel bed and deposition of this sediment on floodplains will therefore not influence downstream fining rates. Downstream fining rates will only be influenced if sediment is deposited that results from suspension of bed sediment shortly upstream (arrow C in figure 23).

Determination of the relative contribution of the two suspended sediment sources thus is crucial for estimating the influence of floodplain sedimentation on downstream fining. The suspended sediment source can be determined in various ways (*e.g.* Wood 1977, Klein 1984, De Boer 1997, Collins *et al.* 1998, Asselman 1999), for instance by analysing mineralogical, colorimetric, mineral-magnetic, chemical, organic, radiometric, isotopic or physical sediment properties.

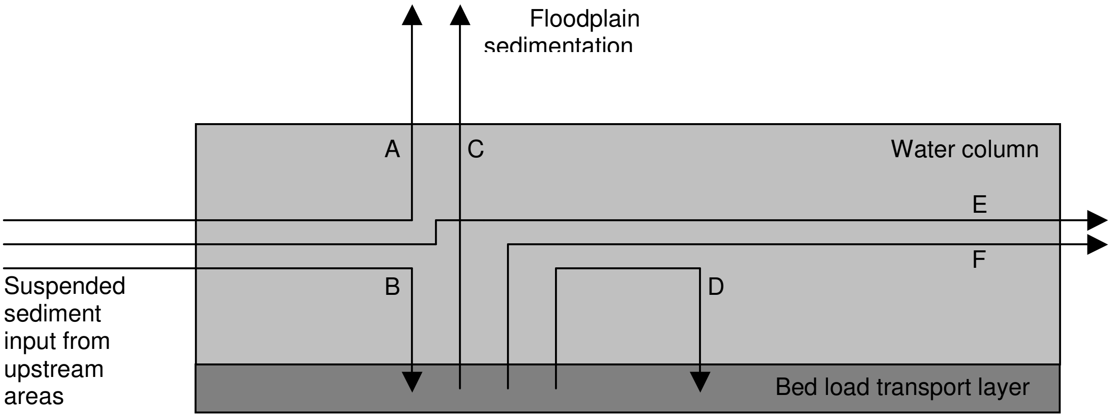

Figure 23. Transport paths of suspended sediment during a single flood. Suspended sediment in a river stems from two sources: hillslope erosion in upstream areas and suspension of local bed sediment. A part of the sediment brought down from upstream areas will directly be transported to the sea (E), another part will become entrapped in floodplains (A), while the rest will be deposited on the river bed where bed gradient and shear stress decrease (B). The sediment that has become suspended locally will either be deposited on floodplains (C), be transported to the sea (F), or be transported a short distance and then be deposited again on the river bed (D, §3.2.2).

Grain size of overbank deposits

The second factor determining the influence of floodplain deposition on downstream fining rates is the difference in grain size between main channel en floodplain deposits. The grain size of the deposited material depends primarily upon the grain size distribution of the suspended load during high discharges, for only the suspended part of the sediment load in the main channel can enter the floodplains. Because the composition of the suspended load is usually much finer than the average composition of the transported load, the material deposited on floodplains will be finer than the average bed composition. If this is the case (and the volumes of sediment involved are relatively large) floodplain sedimentation will cause a decrease of downstream fining rates because the fines are removed from the main channel, while the coarse particles remain behind.

The situation is different in rivers with only suspended load transport. All grain sizes of the sediment transported in the main channel can enter the floodplain, where especially the coarse particles are deposited, because of their larger settling velocity. The fine particles stay suspended and re-enter the main channel at receding flow. Thus especially coarse particles are removed from the main channel, leading to an increase in downstream fining rates (*c.f.* Dietrich 1989).

It is possible that the grain size of the deposited sediment changes with flood magnitude. At increasing flow intensity, the composition of the suspended load in the river usually becomes finer (Asselman & Middelkoop 1998, Xu 2002), because the amount of fine suspended sediment load stemming from soil erosion in upstream areas grows. The suspended sediment locally entrained from the riverbed, however, becomes coarser because high-magnitude floods are capable of transporting coarser particles in suspension (figure 17,18).

Volume of overbank deposits

The last factor determining the influence of floodplain sedimentation on downstream fining is the volume of sediment involved relative to the amount of sediment transport in the main channel. Downstream fining rates can only be influenced by floodplain deposition if relatively large volumes of sediment are involved.

The quantity of sediment involved in overbank sedimentation has been determined for individual floods and as long-term average (see Middelkoop 1997 for an overview of measurement methods). From these measurements it has become clear that the rate of overbank deposition is highest in the downstream part of a river, because of the low gradient, the rising erosion basis and the higher frequency of overbank flows. The rate of overbank deposition also depends on the river's sinuosity. Sinuous rivers have stronger helicoidal flows than straight river and hence a stronger convective sediment transport to the floodplains (Ten Brinke *et al.* 1998). Furthermore, embankments and river training works influence the rate of overbank deposition. Floodplain levels in embanked rivers are generally higher than in natural rivers, lowering the frequency of overbank flow and the rate of overbank deposition. Training works such as groynes dissipate the energy of the currents and protect the fine sediments in between them from erosion and transport to the floodplains. Other factors influencing the rate of overbank sedimentation are the flood magnitude and duration. High-magnitude and high-duration floods usually show much larger volumes of overbank deposits, because they transport more suspended sediment and because they inundate a larger portion of the floodplain for a longer time period (Asselman & Middelkoop 1998, Xu 2002).

Only a few authors compared absolute amounts of overbank deposition with the sediment transport in the main channel. According to Middelkoop (1997) only 4 to 19 percent of the main channel suspended load becomes trapped in the floodplains. Ten Brinke *et al.* (1998) however found that the majority of the sediment transported in the main channel can be deposited on floodplains during high-magnitude floods. These floods however have a low occurrence frequency and therefore the average annual sediment loss to the floodplains is relatively small (*c.f.* Kleinhans 1996), which implies that floodplain sedimentation only has a minor influence on downstream fining rates, especially when considering that a large part of the floodplain sedimentation consists material stemming from soil erosion in upstream areas.

In the future, the influence of floodplain sedimentation on downstream fining rates is likely to increase a bit, at least in Europe. Due to land use changes, less suspended sediment will be fed into the rivers, which will cause a slight decrease in the main channel's suspended sediment transport. In the same time, the expected climate change will probably cause an increase in floodplain sedimentation rates, because of the increase in frequency of occurrence of high discharge stages (Asselman 1998, Middelkoop 1997).

4.5 Groyne field processes

If groynes are present in a river to provide a fairway of sufficient depth and width, the areas in between them usually consist of gently sloping, unvegetated beaches. The sediment in these groyne fields is usually much finer than the sediment in the main channel. Though Sukhodolov *et al.* (2002) did not find any temporal change in groyne field bed levels along the Elbe, this is not the case for other rivers. In the river Waal, for instance, significant temporal changes in groyne field bed levels are observed, implying a significant sediment exchange between the main channel and the groyne fields. Because this sediment transfer involves groyne field sediments that are much finer than main channel sediments, the main channel downstream fining trend could clearly be affected by this sediment transfer. Whether the downstream fining trend decreases or increases depends on the direction of the net sediment transport.

This was studied by Lenders *et al.* (1998) and Ten Brinke (2003, Ten Brinke *et al.* 2001). They found that, during low flow conditions, sediment is transported from the groyne field to the main channel due to currents and waves induced by navigation traffic. Sediment transport from the main channel to the groyne field primarily takes place during times of high discharge, due to natural currents. According to Ten Brinke (2003), the deposition of fines during high discharge periods equals the erosion of fines during low flow periods, so the net transport is zero. This implies that groyne field processes do not affect the downstream fining trend in the main channel.

4.6 Bed erosion and aggradation

Bed erosion can have a significant influence on the grain size of the river bed and therefore on downstream fining rates. During bed erosion especially the finer grain size fractions are removed, leaving the coarser fractions behind. In case of a flat river bed, this leads to the development of an armour layer (see §3.4.1). In the case of a bed covered with bedforms, this leads to the development of a coarse surface layer, which thickness corresponds to the height of the largest dunes (*e.g.* Frings & Kleinhans 2002). In both cases the coarsening of the bed retards further degradation.

Bed aggradation also can influence downstream fining rates. Because especially coarse grains are removed from the bed load, the bed becomes finer during aggradation.

4.7 Dumping and dredging

Dumping of material into the river will only significantly affect the downstream fining trend, if its volume is relatively large and of a distinctly different composition than the local bed (Knighton 1980). Not only the grain size composition of the supplied material is important, but also the lithological composition and the degree of roundness. A supply of weakly durable or angular grains into a river with durable, well-rounded grains on the bed, will fortify abrasion, causing an increase in downstream fining rate downstream of the point of supply (see Gölz *et al.* 1995).

In the Netherlands the dumped material often stems from dredging activities in other parts of the river, for since 1990 net extraction of sediment from the river bed is forbidden in many river reaches (Schans 1998). In this case the supply will not affect the downstream fining rate.

In Germany sediment is primarily dumped to prevent bed degradation downstream of river dams (*e.g.* Gölz 1990, Gölz *et al.* 1995). Though sometimes sediment from upstream of the dams is used, mostly freshly broken sediment from quarries is used. This has a large influence on downstream fining rates, in the first place because the size of the deposited material differs from the local bed material and secondly because the supplied material is abraded rather quickly.

In mining areas often a lot of mining waste is dumped into the rivers. The volumes of sediment involved are usually large, while the supplied material is usually much coarser than the local riverbed material (*e.g.* Cui & Parker 1999, Knighton 1989, 1999a). This has enormous effects on the downstream fining rate.

Dredging activities in rivers have two goals: deepening of shipping routes and economic winning of sand and gravel. Both types of dredging will affect the downstream fining trend if the volume of dredged material is relatively large, while its composition is different from the average bed composition. Dredging can also influence downstream fining by fortifying bed degradation (Frings & Kleinhans 2002), which is probably more important.

4.8 River bifurcations

A very sudden change in downstream fining pattern can occur at river bifurcations. In the meander bend upstream of a river bifurcation the sediment load becomes sorted horizontally through the process of bend sorting. Fine grains are concentrated in the inner bend, coarse grains in the outer bend. Therefore a river branch splitting from the main channel in the outer bend will generally receive a coarser sediment load than the river branch splitting from the main channel in the inner bend. This is probably the case at the bifurcation point IJsselkop in the Netherlands (Gruijters *et al.*2003). The disruption of the downstream fining pattern at river bifurcations thus strongly depends on the process of bend sorting. Bend sorting is the result of the interaction between bend flow and bend topography.

Bend flow

A general characteristic of both straight and curved natural channels is the presence of secondary currents, flowing in the plane orthogonal to the main flow direction (e.g. Thomson 1976, Wilson 1973). The secondary circulation in meander bends is usually dominated by one large circulation cell, that results from the centrifugal force generated by the curvature of the flow (*e.g.* Hey & Rainbird 1996): because the centrifugal force is proportional to the square of the point velocity, the fast-flowing surface water will experience the greatest radial force, driving it towards the outer bend. As a result the water surface in the outer bend will become super-elevated. This generates a cross-stream pressure gradient force, which results in a return flow of near-bed water from the outer to the inner bank (figure 25).

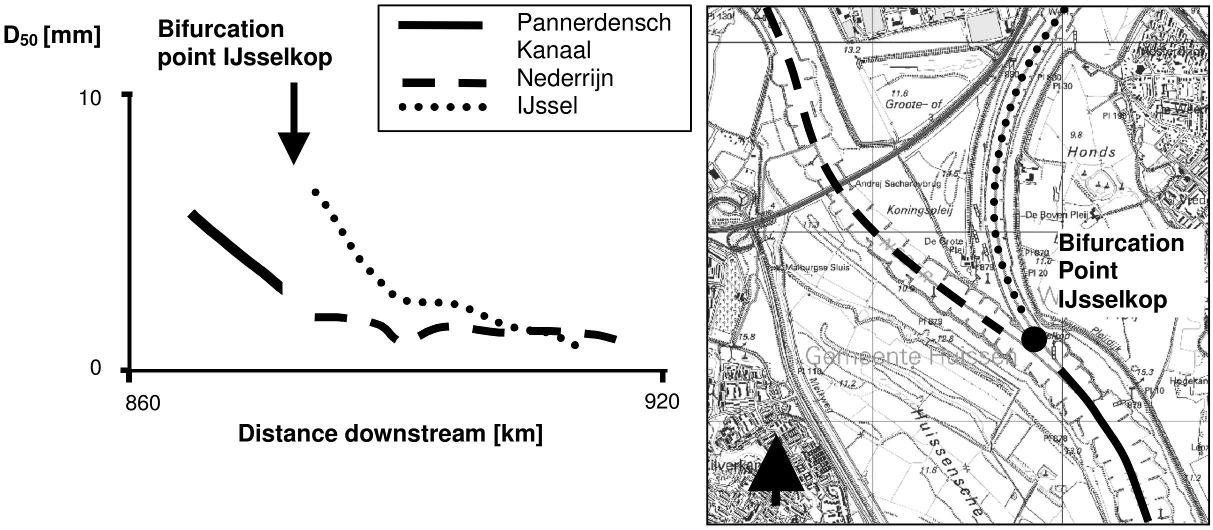

Figure 24. Bifurcation influence on downstream fining (Bolwidt & Jesse 2002).

In addition to this large-scale secondary circulation several small scale phenomena can be present. In the outer bend, for instance, often a small opposite-rotating circulation cell is noticed (Wilson 1973, Thorne & Hey 1979, Geldof & De Vriend 1983). This cell can result from the interaction between the main cell and the outer bank, but it can also be a relic of the circulation cell in the previous bend (figure 25. Near the inner bank usually no circulation cell is visible at all. Shoaling of the flow over the pointbar causes a pressure rise over the bar and a pressure drop over the pool, such that the centrifugal force exceeds the opposing pressure gradient force in the area above the highest parts of the pointbar, resulting in net outward flow (Dietrich & Whiting 1989; figure 25.

Notwithstanding these small-scale phenomena, the water movement in a river bend is largely determined by the main secondary circulation cell together with the primary (downstream) water movement. The combined effect of both is a helicoidal (spiral) water motion, in which the near bed flow is directed obliquely towards the inner bend. The exact direction of the near bed flow depends primarily upon the discharge (flow velocity), the radius of curvature and the width-depth ratio (Yen 1970, Hey & Rainbird 1996 and De Vriend 2001).

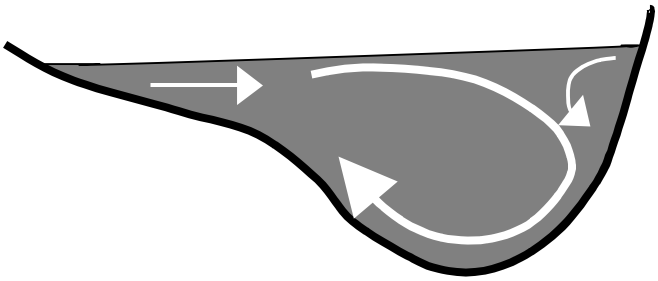

Figure 25. Secondary circulation in meander bends. Shown are the helicoidal motion (large arrow), the effects of shoaling (left arrow) and the outer bend cell (right arrow).

In addition to the direction of the flow in meander bends, also the strength of the flow is important. Field measurements have revealed that the zone of maximum velocity and maximum shear stress gradually shifts towards the outer bank in a meander bend (Leopold & Wolman 1960, Dietrich & Whiting 1989, Julien & Anthony 2002).

For a long period, however, models for water flow in river bends have not been able to reproduce this velocity distribution. The models all predicted the largest velocity and shear stress in the inner bend (*e.g.* Yen 1970), which is incompatible with the common observations of deposition near the inner bank, though some attempts have been made to explain this contradiction (*e.g.* Hooke 1975). Recent models for water flow, incorporating the convective transport of primary flow momentum by secondary currents, however correctly predict the highest flow velocity and shear stress in the outer bend (Johannesson & Parker 1989).

Bend topography

The specific flow pattern in river bends results in a characteristic morphology with the deepest parts near the outer bank and the shallowest parts near the inner bank. The transverse bed slope is highest in the first part of a river bend. This is due to the redistribution of water and sediment, which leads to overshoot effects (De Vriend & Struiksma 1983, Struiksma *et al.* 1985, Struiksma & Crosato 1989). These overshoot effects dampen out and are hardly noticeable in the second part of the bend. In this area the bend flow is fully developed and the transverse bed slope reflects the balance between gravity, which tends to carry particles towards the outer bank, and drag force, which is directed oppositely. Factors influencing the transverse bed slope in this area are the grain size (distribution), the river width and depth, the Froude number, the flow velocity and the radius of curvature (Yen 1970, Wilson 1973, Odgaard 1981 and Bridge 1992).

Bend sorting

The combined effect of bend flow and bend topography is a gradual shift of the locus of the coarsest material towards the outer bank (e.g. Bartholdy & Kisling-Möller 1996, Julien & Anthony 2002). There is some debate, however, about the exact underlying mechanism.

The classical explanation for the horizontal sorting in meander bends is based on the balance between the inward directed drag force and the outward directed gravity (e.g. Deigaard 1980, Parker & Andrews 1985, Bridge 1992 and Julien & Anthony 2002). Because the drag force is proportional to the surface area of the sediment grains ($\sim D^2$), while the gravity force is proportional to the mass (or volume) of the grains ($\sim D^3$), the influence of the gravity force is relatively larger for the coarser grains. Therefore, the net transport direction for the coarse grains is towards the outer bank, while the net transport direction for the fine grains is towards the inner bank. If the bend is sufficiently long, this will result in a progressive segregation of coarse and fine fractions, with ultimately only coarse grains in the outer bend and fine grains in the inner bend.

According to Dietrich & Whiting (1989), this explanation for the horizontal sorting in meander bends is correct for sand-bed rivers but incorrect for gravel bed rivers. The starting point of their reasoning is the spatial variation in bed shear stress. As was described two paragraphs ago, the zone of maximum bed shear stress gradually shifts towards the outer bank in a meander bend. This means that the shear stress near the outer bank gradually increases in downstream direction. Under equilibrium conditions (no erosion or deposition) this increase in bed shear stress must be balanced by either an increase of sediment transport or an increase in bed grain size.

A gradual increase of sediment transport in de outer bend without bed erosion requires a substantial net outward flux of sediment throughout the bend. In gravel bed rivers with low excess-shear stress this does not take place, according to field measurements in two small streams (Dietrich & Whiting 1989). So coarsening of the bed is the only way in which the shear stress increase can be balanced. Coarsening of the bed cannot be established by rolling of coarse grains towards the outer bend, because coarse grains are generally immobile in gravel bed rivers. Coarsening thus must occur due to the winnowing of fine particles in the outer bend, leaving coarse lag deposits behind.

Sand bed rivers usually have high excess-shear stresses. Substantial net outward fluxes of sediment can easily occur and the increase in shear stress in the outer bend is met by an increase in sediment transport. Bend sorting thus can easily be established by the cross-stream transport of coarse particle towards the outer bend and the cross-stream transport of fine particles towards the inner bend.

Local effects

In regular rivers consisting of a series of alternating bends with the same length and curvature, the locus of the coarsest material will lie near the inner bank at the bend entrance, in the middle of the river at the bend apex and near the outer bank at the bank exit, according to Parker & Andrews (1985).

In natural rivers, however, several local effects can modify this pattern and cause the locus of the coarsest material to cross the channel centerline upstream or downstream of the bend apex. For instance, consider the case in which a long, sharp bend is followed by a short, gentle one. If bend sorting results from the outward transport of coarse grains, it is probable that the long, sharp bend will drive so much coarse material to the outer bank, that the locus of the coarsest material will never actually cross to the outside of the succeeding short, gentle bend (Parker & Andrews 1985).

To test the influence of bend length, Frings & Kleinhans (2002) studied bend sorting in 30 river bends in the Dutch river Waal. Their analyses confirm that there is a large variation in the location where the locus of the coarsest material crosses the channel centreline, but the influence of bend length appears to be very small. Apparently other, not yet identified, local factors must play an important role.

Temporal variations

It has regularly been observed that the transverse bed slope changes due to discharge variations (*e.g.* Bartholdy & Kisling-Möller 1996). The way in which discharge variations affect the pattern of bend sorting, however, is still imperfectly understood.

For rivers in which bend sorting is the result of cross-stream transport, computer models predict that an increase in discharge would cause a coarsening of the lower part of the point bar and a change in the location where the locus of the coarsest material crosses the centreline. Field measurements in the river Esk indeed show an increased bed load sorting at high discharges, but the location where the locus of the coarsest material crosses the centreline remains the same (Bridge & Jarvis 1982, *op. cit.* Parker & Andrews 1985).

For rivers in which bend sorting does not result from cross-stream transport, an increase of discharge should also lead to a coarsening of the pool in the outer bend, due to the increase in shear stress. At decreasing discharge the pool will become finer again, due to the deposition of fine particles (Dietrich & Whiting 1989).

It remains a question, however, whether those changes in bend sorting are large enough to affect the size distribution of the sediment loads entering the two branches at a river bifurcation, or not.

4.9 Synthesis

It was seen in this chapter that the effects of abrasion and selective transport can be obscured by sediment addition, extraction and redistribution processes.

Important sediment addition processes are tributary confluences, non-alluvial sediment sources, bed degradation and dredging. Large tributaries usually cause an increase in main stream grain size. Non-alluvial sediment sources generally have the same effect, but have a much more diffuse character. Bed degradation leads to a general coarsening of the bed, because primarily fine grains will be eroded. Dredging mainly influences the main stream grain size by inducing and fortifying bed degradation.

Extraction processes are point bar formation, floodplain sedimentation, aggradation and dumping. Point bar formation often leads to an extraction of coarse material from the main channel, while floodplain sedimentation normally causes an extraction of fine sediments from the main river. During aggradation especially coarse grains will be deposited, leading to a coarsening of the bed. Whether dumping influences the main river downstream fining trend depends on the size and volume of the supplied material.

The only redistribution process that was discussed is the sediment redistribution at river bifurcations. Due to bend sorting, a river branch bifurcating from the main channel in the outer bend will generally receive a coarser sediment load than the river branch splitting from the main channel in the inner bend, leading to a very sudden change in downstream fining trend.

The relative importance of these processes changes in downstream direction. In headwater streams especially non-alluvial sediment sources are important, because floodplains are small or absent here. Processes that are especially important in downstream areas are bed degradation, point bar formation, floodplain sedimentation and sediment redistribution at river bifurcations. These processes require respectively presence of easily erodible (alluvial) deposits, meandering channels, broad floodplains and delta-formation, all features which are characteristic for downstream river reaches. Processes which can affect downstream fining over the entire river reach are dredging, aggradation, dumping and tributary confluences.

Determination of the exact combined influence of these factors requires the construction of a detailed, fraction-wise sediment balance, in which apart from the addition, extraction and redistribution processes also in-channel processes as selective transport and abrasion are quantified.

5 Gravel-sand transitions

One of the most striking expressions of downstream fining is the rapid change from a gravel bedded river (D_{50} about 8 mm) into a sand bedded river (D_{50} about 2 mm), which is observed in many rivers (*e.g.* Knighton 1980, Dawson 1988, Ichim & Radoane 1990, Sambrook Smith & Ferguson 1995). Various mechanisms have been proposed which can explain this phenomenon, which is usually called *the gravel-sand transition*.

The first explanation was put forward by Yatsu (1955). He suggested that grains in the fine gravel range are unstable and tend to be crushed and splitted into their individual minerals. Though this hypothesis has not been falsified yet, it has not been supported by experimental data, so many authors believe that it is incorrect (*c.f.* Russel 1968).

Another explanation was put forward by Sambrook Smith & Ferguson (1996). They argue that there will be a certain location at which the river becomes incapable of transporting sand grains in suspension. At that location, the sand grains will settle and travel further as bed load, while finer grains are entrained into suspension. This implies a huge sand-supply to the river bed, by which the gravel bed can become totally covered.

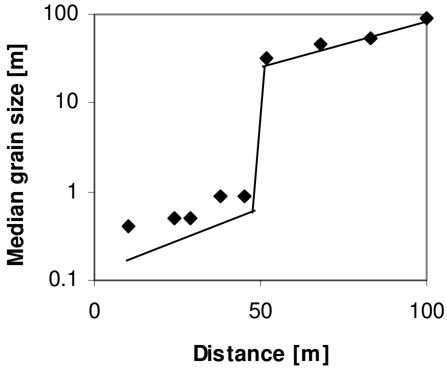

Figure 26. Gravel-sand transition in the Kinu River, Japan (after Parker & Cui 1998).

A gravel-sand transition can also occur when large quantities of sandy sediment are supplied to a gravel-bed river (Sambrook Smith & Ferguson 1995, Sambrook Smith 1996). This was for instance the case in the Ringarooma river (Knighton 1989, 1999a).

A more intrinsic explanation for gravel-sand transitions is the following. The gradual downstream fining in gravel bed rivers leads to a gradual increase in the sand content. If the sand content exceeds a critical value, the gravel framework (pavement) is broken. The bed sediment becomes concentrated in gravely and sandy patches. The gravel patches are virtually immobile, but the sand patches quickly move downstream. The sediment transport thus will quickly increase. This causes bed degradation, which in turn leads to a sudden decrease in bed slope. The slope reduction precludes the downstream transport of gravel, because of the decrease in bed shear stress. So a feedback mechanism develops that enhances the abruptness of the gravel-sand transition. Because the gravel-sand transition in many rivers is associated with a sudden change in bed slope, this *feedback* mechanism is probably important for the formation of gravel-sand transitions in many rivers. According to Sambrook Smith and Ferguson (1996) an additional feedback mechanism is present: the development of patchiness causes a decrease of the bed roughness, which leads to a further reduction in bed shear stress, making the gravel more immobile.

Many gravel-sand transitions slowly move downstream. Some are stable, however. Parker & Cui (1998) demonstrated that two mechanisms can cause an ‘arrested gravel front’: abrasion and basin subsidence (or alternatively sea level rise). A third mechanism that can probably have the same effect is the continuous supply of sand at confluences.

6. Modelling downstream fining

In the preceding chapters three processes were described which determine the rate of downstream fining: abrasion, selective transport and sediment addition, extraction and redistribution. This chapter provides ways to describe these processes and their influence on downstream fining mathematically. Both empirical (§ 6.1) and physical (§ 6.2) downstream fining models are described. In a short synthesis (§ 6.3) attention will be given to the capability of the different types of downstream fining models to simulate and predict downstream fining in natural rivers. The synthesis will also evaluate to what extent downstream fining models can be used to examine the relative importance of the three processes.

6.1 Empirical models

Empirical downstream fining models can be subdivided in exponential abrasion models (§ 6.1.1), exponential downstream fining models (§ 6.1.2) and other empirical downstream fining models (§ 6.1.3)

6.1.1 Exponential abrasion models

The first downstream fining models considered abrasion to be the primary downstream fining mechanism. An exponential form was adopted to describe the downstream decline in abrasion rates. The several factors causing this decline (see paragraph 2.5) were replaced by one parameter, the grain weight, resulting in the next formula:

$$\frac{dw}{dL} = -k_{a,w} w \qquad 6.1$$

where w the particle weight [N], L the distance in downstream direction [m] and $k_{a,w}$ the weight-reduction abrasion coefficient [m^{-1}]. After integration this formula becomes the well-known Sternberg abrasion law (Sternberg 1875):

$$w = w_0 e^{-k_{a,w} L} \qquad 6.2$$

The subscript *0* denotes the value at L=0. Generally this relation is written as (*e.g.* Krumbein 1937):

$$D = D_0 e^{-k_{a,D} L} \quad \text{for single grains, and} \qquad 6.3$$

$$D_x = D_{x,0} e^{-k_{a,D} L} \quad \text{for sediment mixtures} \qquad 6.4$$

with D the grain diameter [m] and D_x a representative mixture grain size, for instance the median diameter or the *D90*. $k_{a,D}$ is the diameter reduction abrasion coefficient [m^{-1}]. Many authors have determined diameter reduction abrasion coefficients, mostly in abrasion experiments. Reported values generally range from 10^{-8} to 10^{-4}, depending on the parameters described in paragraph 2.4 (Morris & Williams 1999b). Only Kodama (1994a) reported values from 10^{-3} to 10^{-2}.

Diameter reduction coefficients can be used to calculate weight reduction coefficients. Because w is proportional to D^3, $k_{a,w}$ usually is taken as $3k_{a,D}$. Mikos (1995) however showed that this is only correct for abrasion of single clasts. When size-diminution of sediment mixtures is studied, $k_{a,w}$ can be larger or smaller than $3k_{a,D}$. For instance, if size classes are relatively broad, abraded particles will fall in the same size class as before abrasion. Median grain size will be incorrectly calculated and $k_{a,D}$ will be underestimated, resulting in $k_{a,w} > 3k_{a,D}$. Alternatively $k_{a,w}$ will be smaller than $3k_{a,D}$ when the entire sediment mixture is analyzed, for $k_{a,D}$ will be overestimated because not only the size of the abraded grains is measured but also the size of the abrasion products. Lewin & Brewer (2002) found that $k_{a,w}$ equals 1.9 to 2.6 times $k_{a,D}$.

6.1.2 Exponential downstream fining models

Short after the development of Sternberg's abrasion model, the influence of selective transport on downstream fining rates was recognized (Daubrée 1879 *op. cit.* Krumbein 1941). When abrasion and selective transport act independently, equation 6.4 can be written as (*e.g.* Church & Kellerhals 1978, Knighton 1980):

$$D_x = D_{x,0} e^{-(k_{a,D}+k_{s,D})L} \qquad 6.5$$

or alternatively as:

$$D_x = D_{x,0} e^{-k_D L} \qquad 6.6$$

with $k_{s,d}$ the diameter reduction coefficient due to selective transport and k_D the overall diameter reduction coefficient, due to the combined effects of abrasion and selective transport. Many authors used equation 6.6 to fit their grain size data (*e.g.* Rice & Church 2001, Surian 2002, Church & Kellerhals 1978, Knighton 1980, 1982, Brierley & Hickin 1985, Dawson 1988, Brewer & Lewin 1993, Kodama 1994b). k_D values found range between 10^{-7} and 10^{-3} (Morris & Williams 1999b) which means that they are (on average) a factor 10 larger than $k_{a,D}$ values, implying that abrasion is less important than selective transport in most rivers, though it can be the primary downstream fining mechanism in some rivers.

7.1.3 Other empirical downstream fining models

For most rivers, exponential models provide a very good description of downstream fining trend. They are inappropriate however for other rivers. This is for instance the case in many upstream river parts. Several explanations have been suggested for this, like tributary influence, lithogical differences, slope discontinuities and lack of grains in a particular grains size class (see Brierley & Hickin 1985 for an overview.

In some cases replacing the exponential models by linear or power models can solve the problems. In other cases the river can be divided in several homogeneous parts, each of which can be described by a different exponential model. This concept was extended to drainage scale by Rice (1994) and Pizzuto (1995). They consider the drainage basin as a network of interdependent channel links, separated by nodal points (the confluences). At nodal points the sediments from the two upstream links are mixed, which results in a discontinuous decrease or increase in grain size, depending on the volume and size-distribution of the sediment mixtures supplied from the upstream links. The downstream decline in grain size within a link is described with an exponential model if non-alluvial sediment inputs are absent. The variation in grain size within a link is described by a stochastical model if the link receives a lot of non-alluvial sediment (Rice 1994).

6.2 Physical models

While empirical downstream fining models are principally a description of observed downstream fining patterns, physical downstream fining models try to explain and simulate downstream fining using physical theories about the downstream fining mechanisms.

Two early physical downstream fining models were made by Rana *et al.* (1973) and Troutman (1980). Rana *et al.* assumed an exponentially decreasing bed slope and calculated the decrease in grain size necessary to transport a constant water and sediment discharge down the channel. Downstream fining thus is caused by the preferential transport of fine grains. Troutman’s model is based on the Einstein step length theory. According to this theory grains alternate periods of movement (steps) with periods of rest. It is assumed that small grains have a larger average step length and a shorter average rest period than coarse grains. Small grains are thus preferentially transported downstream.

Both models have in common that sediment continuity is not satisfied, which puts a severe restriction on the applicability of these models. Since 1980 more realistic downstream fining models were developed, all based on the continuity equation for sediment. The models differ, however, in the exact downstream fining mechanisms that have been incorporated.

In the following, three categories of downstream fining models will be discussed: models solely based on selective transport processes (§ 6.2.1), models based on selective transport and abrasion processes (§ 6.2.2), and models including the effects of local sediment addition, extraction and redistribution processes (§ 6.2.3).

6.2.1 Models solely based on selective transport processes

The sediment continuity equation

Downstream fining models based on selective transport processes all use the Hirano active layer concept (*e.g.* Hirano 1971 *op. cit.* Blom 2003) to calculate the sediment exchange between bed load and bed. This involves the division of the river bed into a top layer and a non-moving homogeneous substrate (see figure 27). In case of net aggradation a sediment flux occurs from the active layer to the substrate through a rise in the interface between the active layer and the substrate. In case of net degradation, the bed surface and the interface between the active layer and the substrate will lower, which induces a sediment flux from the substrate to the active layer. Sediment from the substrate thus cannot be entrained by the flow directly, only after being entrained into the active layer. The one-dimensional continuity equation then reads:

$$(1-\lambda)\frac{\partial(F_i L_a)}{\partial t}+(1-\lambda)\frac{E_i\partial z_0}{\partial t}=-\frac{\partial(q_T p_i)}{\partial x} \qquad 6.7$$

λ denotes the bed porosity [-], L_a the thickness of the active layer [m], z_0 the elevation of the interface between the active layer and the substrate [m], F_i the volume fraction content of size fraction *i* in the active layer [-], E_i the volume fraction content of size fraction *i* of the material that is exchanged between the active layer and the substrate [-], p_i the volume fraction content of size fraction *i* in the transported material [-] and q_T the volume of sediment transport per unit width and time [m^2/s].

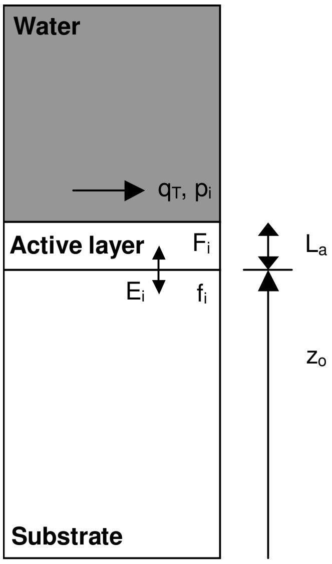

Figure 27. The Hirano active layer concept

Solving the sediment continuity equation requires 1. Specification of the active layer thickness L_a. 2. Specification of E_i. 3. Selection of an appropriate sediment transport formula. 4. Calculation of the flow parameters to be used in the sediment transport formula and 5. Specification of boundary conditions and initial conditions. These factors will be shortly addressed hereafter.

The active layer thickness can be calculated as function of the sediment size (*e.g.* Hoey & Ferguson 1994, Cui *et al.* 1996), the bed form height (*e.g.* Deigaard 1980, Armanini & Di Silvio 1988) or the bed shear stress (*e.g.* Van Niekerk *et al.* 1992).

Determination of the composition of the material exchanged between the substrate and the active layer (E_i) is relatively straightforward in case of degradation. Because a stream then incorporates substrate material into the active layer, E_i must be equal to f_i, the volume fraction content of fraction *i* in the substrate. For the case of aggradation, the appropriate choice is not as clear (Parker 1991). If both the rate of sediment transport and sediment deposition are slow, it is likely that the deposited material may be mixed with the active layer before releasing the portion thicker than L_a to the substrate. In this case: $E_i = F_i$. Another possibility is that the deposited bed load is immediately worked down below the active layer, for example at the leeside of dunes, where deposited grains are immediately covered with other grains. If this were true: $E_i = p_i$. Because it is probable that E_i will be somewhere in between F_i and p_i, most authors calculate E_i as the weighted average of F_i and p_i (*e.g.* Toro-Escobar *et al.* 1996, Hoey & Ferguson 1994, Cui *et al.* 1996).

The selection of an appropriate sediment transport formula is the most crucial point in solving the continuity equation and predicting the downstream fining pattern. Downstream fining models based on selective transport primarily distinguish themselves from each other in the way selective transport is incorporated in the continuity equation (see below).

Calculation of the flow parameters to be used in the sediment transport model usually takes place by solving the St. Venant shallow water equations. Normally it is assumed that the water flow is steady, so that all derivatives of time can be omitted. This involves a considerable numerical simplification in that the flow equations are decoupled in time from the bed evolution equations. According to Cui *et al.* (1996) this does not seriously harm the results.

Boundary conditions that have to be specified are the water level at the downstream end of the river and the water and sediment input at its upstream end. The models require furthermore specification of the initial bed profile and bed composition.

Bed load transport

In a few models it is assumed that downstream fining only results from size-selectivity during bed-load transport. This is for instance the case in the model made by Paola *et al.* (1992a), which was meant to study vertical grain-size changes in an aggrading alluvial basin. The model uses the Meyer-Peter & Müller (1948) equation to calculate the total bed load transport. Hiding and exposure effects are not included. It is simply assumed that each grain size is deposited until it is exhausted, at which point deposition of the next size begins, and so forth until the end of the deposit is reached. In order to include hydrologic regime in a simple way, an intermittency is introduced such that a constant flow is maintained for only a fraction of the year; for the rest of the year the channel is assumed to be dry. This is called *the intermittency concept.* In order to describe avulsion and resulting floodplain reworking, any deposited sediment is distributed across the whole floodplain, rather than restricted to the channel itself.

Examples of bed-load-based downstream fining models that include hiding exposure effects are the models by Hoey & Ferguson (1994) and Cui *et al.* (1996). In both models, the bed load transport of is calculated using Parker's (1990) transport relation, which includes a hiding exposure coefficient of 0.90, resulting in slightly selective transport. Both models neglect the bed load transport of sand.

Bed load and suspended load transport

To predict downstream fining in rivers with a considerable amount of suspended sediments, Deigaard (1980) developed a model, which includes both bed load transport and suspended load transport. Bed load transport is calculated using a straight generalisation of the Engelund & Fredsoe (1976) formula, with no accounting for hiding and exposure effects. Suspended sediment transport is calculated as the depth-integration of the product of the local flow velocity and sediment concentration. The vertical distribution of the sediment concentration is determined with a variant of the Rouse equation in combination with the Engelund & Fredsoe (1976) relation for near-bed sediment concentration. Deigaard's model is restricted to sand size sediments.

After Deigaard's (1980) model no new downstream fining models have been developed which include suspended sediment transport. However, there are some sediment routing models, that – though not explicitly meant to simulate downstream fining – are capable of predicting downstream fining. Examples are the models made by Armanini & Di Silvio (1988), Van Niekerk *et al.* (1992, Vogel *et al.*1992) and Cui & Parker (1999).

The model by Armanini & Di Silvio (1988) is based on the (simplified) bed load transport and suspended load transport formulae of Van Rijn (1984a,b). Both formulae were adapted for sediment mixtures, by including a correction for hiding-exposure effects. It is in fact meant to predict downstream fining in sand-bed rivers. Characteristic for Armanini & Di Silvio's (1988) model is the lag effect in the transport of suspended material. It is assumed that the actual suspended load transport only slowly reacts to a change in the equilibrium transport (as predicted with the Van Rijn's formula). The bed load transport is assumed to react immediately to changes in the equilibrium bed load transport as predicted with Van Rijn's formula. Another feature of Armanini & Di Silvio's model is the incorporation in the continuity equation of sediment storage in the water column.

Van Niekerk *et al.* (1992) calculate the suspended sediment transport like Deigaard (1980) as the depth-integration of the product of the local flow velocity and sediment concentration, but instead of using the Rouse equation to determine the vertical distribution of sediment concentration, the full convection-diffusion-equation is used (e.g. Voogt *et al.* 1991, Van Rijn 1984b). The near-bed sediment concentration, which is needed to solve the convection-diffusion equation, is determined using a relation proposed by Bridge & Dominic (1984). To calculate bed-load transport, the bed load formula of Bridge & Dominic (1984) formula is adopted, in combination with a hiding exposure correction. The model is suitable for sand and gravel rivers, and incorporates the effects of density differences on the transport rates.

Cui & Parker (1999) developed a sediment routing model for a sand bed stream. Total bed material transport of sand is calculated using the Brownlie (1981) relation. The Rouse equation is used to determine the vertical variation in sediment concentration. The model is based on a daily discharge regime and includes floodplain sedimentation and tributary influence (see below).

The effects of armour layers

The effects of armour layers have not yet been included explicitly in downstream fining models. This is partly because the mechanisms of armouring are not fully understood and partly because incorporation of armouring processes requires computations with small time steps, which costs a lot of time. However, some efforts have been made to model the development of armour layer. One especially noteworthy model is the static armour layer model made by Borah *et al.* (1982). The model predicts that all grains with a critical shear stress smaller than the actual shear stress are removed from the bed surface. This exposes underlying fine grains, which are also eroded. The bed surface now only contains grains that cannot be moved, thus a static armour layer has formed. In this situation, no erosion can occur until the shear stress becomes equal to the critical shear stress of the smallest size fraction present in the armour coat.

The effects of patchiness

The effects of patchiness on the degree of size-selectivity have been incorporated in a downstream fining model by Paola & Seal (1995). Their model is restricted to bed load transport and uses a simplified form of the continuity equation. The bed load transport of a grain size fraction *i* (q_i) is calculated as:

$$q_i = q_T f_i J_i \qquad 6.8$$

with q_T the total bed load transport, , f_i the volume fraction content of the *i*th size fraction in the bed and J_i a relative mobility function. For a situation with two patch types, J_i can be written as:

$$J_i = \frac{q_{T,a} p_a f_{i,a} + q_{T,b} p_b f_{i,b}}{q_T f_i} \qquad 6.9$$

$q_{T,a}$ denotes the transport rate of uniform material with a grain size equal to the median grain size of patch *a*, p_a is the fraction of the total bed surface covered with patches of type *a* and $f_{i,a}$ is the volume fraction content of grain size fraction *i* in the sediment mixture of patch *a*. $q_{T,b}$, p_b and $f_{i,b}$ are defined analogously for patches of type *b*. It is assumed that the sediment mixture within patches exhibits equal mobility, and the total transport rate of a patch was calculated with the Meyer-Peter & Müller (1948) formula.

The effects of sorting in bed forms

The influence of sorting processes in bed forms on downstream fining has not yet been incorporated in downstream fining models. In recent years, however, significant progression was made in the physical description of the sorting processes (*e.g.* Ribberink 1987, Parker *et al.* 2000 & Blom 2003).

Ribberink (1987, *op. cit.* Blom 2003) realised that in deep bed form troughs, fine grains are picked up from the substrate, while coarse grains settle. He therefore introduced a sediment exchange mechanism related to deep bed form troughs, using an additional layer between the active layer and the substrate. His model, however, doesn't account for sorting processes at the lee side of dunes.

This became only possible after Parker *et al.* (2000 *op. cit.* Blom 2003) found a way to leave the discrete representation of the river bed. The sediment concentration of size fraction *i* at each desired bed level *z* can be calculated according to Parker *et al.* as:

$$C_i = (1-\lambda) P_s F_{z,i} \qquad 6.10$$

with $F_{z,i}$ the volume fraction content of size fraction *i* at elevation *z* and P_s the probability density function of bed surface elevations, indicating the probability that the bed surface elevation is higher than *z*. The sediment continuity equation now reads:

$$\frac{\partial C_i}{\partial t} = D_i - E_i \qquad \rightarrow \qquad (1-\lambda) P_s \frac{\partial F_{z,i}}{\partial t} + (1-\lambda) F_{z,i} \frac{\partial P_s}{\partial t} = D_i - E_i \qquad 6.11$$

with D_i the deposition density of size fraction *i*, such that $D_i dxdz$ is the volume of sediment of size fraction *i* that is deposited per unit width and time in a bed element at elevation *z*. E_i is the entrainment density of size fraction *i,* and defined similarly.

From equation 6.11 it becomes clear that the vertical sorting profile ($F_{z,i}/dz$) can be calculated when the probability density function of surface elevations P_s and the deposition and entrainment densities (D_i and

E_i) are known. Blom (2003) provided a way to calculate D_i and E_i based on the Einstein step length theory and a empirical lee side sorting function.

6.2.2 Models including abrasion and selective transport processes

The first downstream fining model that incorporates abrasion in a physical way was made by Parker (1991). He added a term A to the right hand side of equation 6.7, which denotes the mass loss through abrasion. It is assumed that abrasion only takes place due to binary collisions between moving bed load and stationary bed particles. Both clasts involved in a collision are supposed to experience a small mass loss, in the form of fine silt particles that are immediately washed away as wash load. The mass loss trough abrasion (A) thus can be calculated by multiplication of the number of grain collisions with the mass loss caused by a single grain collision. The number of grain collisions is deduced from the sediment transport rate and the average saltation distance, while the mass loss during a single grain collision is calculated using an laboratory abrasion coefficient that depends on grain size and grain lithology.

The model only simulates the abrasion processes grinding and cracking. Processes like splitting, chipping, crushing, or abrasion in place are not simulated. The models also doesn't account for the influence of mixture bimodality, grain roundness, grain shape, wheatering and the amount of moving particles on the abrasion rates. Other characteristics of Parker's (1991) model are: a continuous grain size distribution, incorporation of the effects of tectonic subsidence or rise, ignorance of grain sizes smaller than 2 millimeters and ignorance of suspended load transport. Bed load transport is calculated through Parker's (1990) bed load relation, which included hiding exposure effects.

In 1998 a new downstream fining model capable of simulating abrasion was presented by Parker & Cui (1998, Cui & Parker 1998). Primary purpose of this model was to prove that a stable gravel-sand transition can be caused by either tectonic subsidence or abrasion. The formulation of the abrasion process is comparable to that in Parker (1991), but now it is assumed that gravel grains smaller than 8 mm are unstable as was postulated by Yatsu (1955). When the size of a grain becomes equal to 8 mm during the abrasion process, the grain falls apart into several sand grains. These sand grains are not affected by abrasion anymore according to the model.

The model differs also at some other points significantly from Parker's (1991) model. In order to include hydrologic regime in a simple way, the intermittency concept is used. In order to describe avulsion and resulting floodplain reworking, any deposited sediment is distributed across the whole floodplain, rather than restricted to the channel itself. Bed load transport of sand-size grains is not ignored anymore. While transport of gravel is again calculated with Parker's (1990) relation, Engelund & Hansen's (1972) formula is adopted for the bed load transport of sand.

6.2.3 Models including the effects of sediment addition, extraction and redistribution processes

In chapter four a large number of sediment addition, extraction and redistribution processes have been identified that can influence downstream fining rates: bed erosion and aggradation, tributary confluences, non-alluvial sediment sources, dredging and dumping, floodplain sedimentation, groyne field processes, bank erosion & point bar formation and sediment redistribution at river bifurcations.

Most downstream fining models account automatically for the effects of bed aggradation and erosion, because this term is automatically incorporated in the continuity equation (6.7). Incorporation of the effects of tributary confluences in downstream fining models can easily be done. It only requires a simple description of the mixing of sediment and water at the confluence point (*e.g.* Cui & Parker 1999).

To account for the effects of non-alluvial sediment sources, bank erosion and point bar formation, dredging and dumping, floodplain sedimentation and groyne field processes, it suffices to add a term $q_{lat,i}$ to the right side of the continuity equation (*e.g.* Cui & Parker 1999). Problem is, however, to quantify this term $q_{lat,i}$. Only in case of floodplain sedimentation relations have been developed for $q_{lat,i}$. Cui & Parker (1999), for instance, used a convective model to determine $q_{lat,i}$. They assumed that, whenever the floodplain is inundated, a fixed percentage of the suspended load in the water column above bankfull height, is deposited in the floodplain. This implies that the composition of the deposited material is equal to the composition of the suspended load above bankfull height.

The effects of sediment redistribution at river bifurcations have not yet been included in downstream fining models. These processes cannot be dynamically incorporated in a one-dimensional model of water and sediment flow.

6.3 Synthesis

Empirical (exponential) downstream fining models have proved to be very useful for the description of downstream fining patterns in river reaches in which abrasion and selective transport are the major factors causing downstream fining. These are either short homogeneous river reaches in between bifurcation points and confluences (sometimes called sedimentary links) or river reaches of considerable length in which the disturbing influence of processes at river bifurcations or river confluences averages out (see Rice 1998, 1999, Dawson 1988, Surian 2002, Werrity 1992).

Empirical downstream fining models are also very useful to determine the relative importance of abrasion and selective transport. The only thing needed is a comparison of laboratory determined abrasion coefficients ($k_{a,D}$-values) with overall diameter reduction coefficients (k_D-values), determined using field grain size samples. From such comparisons it has become clear that k_D-values are on average a factor 10 larger than $k_{a,D}$-values, implying that abrasion is less important than selective transport in most rivers.

A more detailed picture of the relative importance of abrasion and selective transport can be obtained from Morris & Williams (1999a,b). In a study of more than 100 rivers, they found two relationships:

$$k_D = \Lambda^{-1} \qquad \text{for } 7 < \Lambda < 1{,}770{,}000 \qquad 6.12$$

$$\varepsilon = 1.1\Lambda^{-1} \qquad \text{for } 1 < \Lambda < 1{,}770{,}000 \qquad 6.13$$

with Λ [m] the river length and ε [m^{-1}] the coefficient of concavity, indicating the rate at which the river bed gradient decreases in downstream direction. Combining equations 6.12 and 6.13 results in: $k_D \sim \varepsilon$, which implies that the downstream fining rate causally depends on the decrease in river bed gradient. This is reasonable because the rate at which the river bed gradient decreases determines the rate at which the bed shear stress decreases and therefore the degree of transport selectivity. The relationship between k_D –values and river length thus only is indirect: longer rivers have a smaller rate of river bed gradient decrease and therefore the transport process is less selective in these rivers, resulting in smaller k_d-values. This suggests that the influence of abrasion is relatively large in longer rivers (the low k_D-values of these rivers are comparable to those determined in abrasion experiments), while its influence on downstream fining in short river reaches is negligible.

A disadvantage of empirical downstream fining models is that they are not able to simulate the effects of local sorting processes (*e.g.* armouring), local sediment additions and extractions (e.g. tributary influence) and sediment distribution at river bifurcations properly. Physical downstream fining models are based on the sediment continuity equation and provide the opportunity to incorporate these effects in the model.

The present generation of physical downstream fining models incorporates the effects of selective transport, patchiness, bed erosion and aggradation, tributary confluences, dumping and dredging and floodplain sedimentation in a satisfactory way. The available models do not have proper formulations for abrasion, armouring, bed form sorting, non-alluvial sediment additions, groyne field processes, bank erosion/point bar formation and sediment redistribution at river bifurcations. For bed form sorting, recently promising formulations have been proposed, so in future the effects of sorting in bed forms can be incorporated in downstream fining models.

Difficulty with physical downstream fining models is that they are rather complex, while they all contain some parameters that have to be determined by calibration. The relative importance of two downstream fining processes thus cannot be determined by comparing the degree to which two different models are able to simulate observed downstream fining patterns. Through calibration a model based on selective bed load transport can be adjusted to predict the same downstream fining pattern as a model based on selective suspended load transport, or a model based on abrasion. This does not alter the fact, however, that physical downstream fining models can be very useful to determine the relative importance of the several processes influencing downstream fining. Therefore sensitivity analyses should be carried out, in which the parameters determining one downstream fining process are changed, while the parameters indicating the strength of the other processes are held constant.

7. Conclusions

7.1 Abrasion

Abrasion experiments

- In most cases, field studies do not provide reliable estimates of the abrasion rate.
- Tumbling barrels and abrasion tanks used in laboratory abrasion experiments vary widely in properties. The resulting differences in abrasion intensity and dominant abrasion process make comparison of the experiment findings difficult.

Dominant abrasion processes

- The dominant abrasion processes in the first kilometres of gravel bed rivers are chipping and splitting. Further downstream grinding and cracking are most important.
- The dominant abrasion process in sand bed rivers is grinding, though some chipping can occur during the first kilometres of transport.
- Little is known about the dominant abrasion mechanism in mixed sand-gravel bed rivers. There are some indications that splitting and crushing are highly important.

Abrasion rate

- The abrasion rate depends on the size, lithology, roundness, shape and velocity of the transported grains and also on the mixture composition, the river bed grain size, the amount of weathering, the number of moving particles and the dominant abrasion process.
- Grain size and lithology are the most important among these variables. Naturally occurring variations in grain size and lithology cause differences in abrasion rates of a factor 1000 and 100, respectively. The other variables only cause variations in abrasion rates of a factor 10 or less under natural conditions.
- It is still impossible to predict the exact abrasion rate at a given location in a river.
- In many rivers, however, some of the factors determining the abrasion rate are interrelated, amplifying each other's influence and causing a strong decline in abrasion rates in downstream direction. The continuous supply of fresh, angular, coarse rock fragments into steep-bedded mountain streams, causes high abrasion rates, primarily by splitting and chipping. This results in a strong decrease in grain size, a decline in the presence of weakly durable lithologies, an increase in grain roundness and a change towards less effective abrasion mechanisms (cracking and grinding). In combination with the decrease in bed slope and the gradual transition to sand beds, this leads to the strong decline in abrasion rates in downstream direction.
- Despite the general decrease of abrasion rates in downstream direction, it is possible that during the transition from a gravel bed to a sand bed locally higher abrasion rates prevail, due to severe crushing of fine grains between the coarser gravel particles.

7.2 Selective transport

- Grain-size selective transport can be established through selective entrainment, selective movement and selective deposition, but also by local sorting processes like dune sorting, armouring and patchiness.

Selective entrainment

- The commonly used criterion for the beginning of motion ($\tau > \tau_c$) is only an absolute lower limit for entrainment when applied to single grains.
- The common use of instantaneous τ-values in combination with time-averaged τ_c-values in incipient motion studies is incorrect.
- To obtain selective entrainment, $\tau_{c,i}$ should be larger for coarse grains than for fine grains, but equally important is the requisite that τ should lie in between the $\tau_{c,i}$ for the coarsest grains and the $\tau_{c,i}$ for the finest grains for a considerable part of the year.
- $\tau_{c\ i}$ is not the average critical shear stress of all grains in a size fraction, but the shear stress at which the first grains in a size fraction start to move.
- Empirical studies to determine the degree of selective entrainment have suggested strongly selective entrainment in some rivers, but equal entrainment mobility in others.

- This can be partly explained by differences in the bimodality and median grain size of the sediment mixture, and by differences in the structure of the river bed. Equally important, however, are differences in the experimental methods. Important aspects are the method to determine the critical bed shear stress, the range of D_i/D_{50} used, the definition of beginning of motion, the choice for D_{50} and the way corrections are made for specific environmental conditions.
- Unimodal and weakly bimodal sediments exhibit severe hiding-exposure effects, making all grain sizes virtually equally mobile. Strongly bimodal sediments are characterised by strongly size-selective entrainment.
- The entrainment of grain size fractions smaller than the median grain size is often grain-size independent, while grain size fractions much coarser than the median grain size exhibit selective entrainment.
- This is correctly predicted by Egiazaroff's curve, but empirical relations cannot reproduce this, because they usually, incorrectly, assume a power relation between the critical bed shear stress and the grain size.
- Usually the beginning of motion is defined as a small, but measurable, reference transport rate. Strikingly, this reference transport is not always equal for all grain size fractions.
- To determine the degree of selective entrainment, the median grain size of the *bed surface* should be used during the derivation of the hiding-exposure coefficient. If the median grain size of the *subsurface* is used, indirectly also the effects of vertical sorting processes, like armouring, are studied.

- There is a range of shear stresses in which fine fractions are fully mobilized, while coarse size fractions are only partially transported. This is a form of selective entrainment that has hardly been studied, but probably is more important than the exhaustively studied selective entrainment during initial motion (to which all preceding conclusions refer).

Selective movement

- Information about the selectivity of bed load movement over sandy, dune-covered beds is very scarce.
- Information about the selectivity of bed load movement over plane gravel beds has to be derived from experiments in which the grain velocity, the mean step length or the travel distance during a flood were determined.
- The experiments suggest that coarse bed load grains have a larger velocity of travel and a larger step length than fine grains, but also that this is counteracted by a smaller pick-up rate. The distance travelled during a flood decreases with grain size, but this is probably due to the fact that coarse grains are only mobile during a part of the event. Therefore it can be surmised that the process of bed load movement over plane gravel beds is only weakly selective, or even non-selective.
- If a river is characterized by both bed load transport and suspended load transport, the grain movement process will always be size-selective, because the velocity of suspended load transport is much higher than the velocity of bed load transport.
- Criteria for the beginning of suspension for uniform sediment differ widely, because they point to different probabilities of suspension. All criteria, however, predict a strong increase in critical shear stress with grain size. In non-uniform sediment mixtures coarse grains will still become suspended at higher shear stresses than fine grains, but the increase in critical shear stress with grain size will be less than in uniform sediments, due to hiding-(exposure) effects.
- The size-selectivity due to the division between bed load transport and suspended load transport will increase at increasing discharge, because the percent of the bed grains that is suspended increases much faster with shear stress for fine grains than for coarse grains.
- The movement of grains in the suspended load layer itself is also a size-selective process. Coarse grains are only present in the lowermost part of the suspended load layer, where the fluid velocity is low, while fine grains are present throughout the suspended load layer, thus having a larger average movement rate.

Selective deposition

- Coarse grains usually are deposited at higher shear stresses than small grains, but the critical shear stress at which a grain will settle is somewhat smaller than the critical shear stress at which it was entrained.
- In case of a low bed roughness, coarse grains will settle at a lower shear stress than fine grains. This is because coarse grains will not encounter a pore that is large enough to settle in.

Selectivity due to local sorting processes

- The formation of armour layers always involves size-selective transport
- Sediment entrainment is a non-selective process in case of a stable armour layer. Selectivity, however, can arise when fine material – in the form of sand waves or dunes - moves over the stable armour layer.
- Pavements can cause a situation of equal mobility because of the overrepresentation of coarse grains at the bed surface, making them more available to the flow.
- Even if equal mobility is satisfied locally, the total sediment transport process can be size-selective when the bed is divided in slow moving patches of coarse material and fast moving patches of fine material. This is called the patchiness hypothesis.
- Dune sorting concentrates coarse grains in deep bed layers (fig. 2), which are only mobile at high discharges. Coarse grains thus will generally be underrepresented in the bed load, causing a distinct difference between transport rates of coarse and fine material.

Overall degree of size-selectivity

- The overall degree of size-selectivity due to the combination of selective entrainment, selective movement, selective deposition and selectivity due to sorting processes, can be determined by comparing the size composition of the transport material with the size composition of the subsurface bed material. This is only meaningful for rivers without suspended load transport.
- Usually a coarsening of the bed load at increasing bed shear stresses is observed, indicating a tendency toward equal mobility at high bed shear stresses, but a strong degree of size-selectivity at low bed shear stresses.
- Averaged over a year, small grains seem to be equally mobile, while coarse grains exhibit a strong degree of selective transport. Some authors however, also found equal mobility for coarse grains.
- When considering spatial variations in size-selectivity it is found that sediment transport is selective in upstream river reaches, but non-selective in downstream reaches.

Relation between selective transport and downstream fining

- All four forms of selective transport can cause a rapid downstream fining pattern to develop in the early stages of a river.
- Selective entrainment and selective deposition produce downstream fining patterns that can be seen as a (semi-)equilibrium situation.
- Selective movement caused by the division between bed load and suspended load transport also results in a stable downstream fining pattern, but only in sand bed river reaches.
- Downstream fining patterns solely caused by selective bed load movement are a non-equilibrium phenomenon.
- Dune sorting is the only sorting process that can cause downstream fining by itself. Patchiness and armouring only affect the degree of selective entrainment, selective movement and selective deposition.

7.3 Sediment addition, extraction and redistribution

- The effects of abrasion and selective transport can be completely obscured by sediment addition, extraction and redistribution processes. Examples are: tributary confluences, non-alluvial sediment sources, bed degradation and aggradation, groyne field processes, dumping and dredging, bank erosion and point bar formation, floodplain sedimentation and sediment distribution at river bifurcations.
- Most tributaries do not significantly affect the main stream downstream fining trend. The remaining tributaries usually cause an increase in main stream grain size, because coarse sediment entering a river with fine bed material is often directly deposited, while fine sediment entering a coarse river is immediately washed away.
- Non-alluvial sediment sources have a very diffuse character and can preclude a systematic diminution of sediment grain size in river head waters.
- Point bar formation leads to an extraction of coarse material (the channel lag) from the main channel in aggradational sites.
- Floodplain sedimentation leads to an extraction of fine sediments from the main river. Because only a part of the deposited sediments stems from suspension of river bed material, the influence on the main stream grain size is relatively low.
- Groyne field processes probably do not result in a net sediment transfer from to the main river.

- Dumping influences the main river downstream fining trend when the supplied material has a different size composition or different abrasion durability than the material in the main river. Dredging especially influences the main stream grain size by inducing and fortifying bed degradation.
- Bed erosion can significantly affect downstream fining. It usually leads to a coarsening of the bed.
- At river bifurcations a very sudden change in downstream fining trend can occur, because a river branch bifurcating from the main channel in the outer bend will generally receive a coarser sediment load than the river branch splitting from the main channel in the inner bend, due to the process of bend sorting. Bend sorting in sand bed rivers results from the balance between the inward directed drag force and the outward directed gravity. Bend sorting in gravel bed rivers is caused by grain size adaptation to spatial shear stress variations.
- The relative importance of these processes changes in downstream direction. In head water streams especially non-alluvial sediment sources are important, while in downstream areas bed degradation, pointbar formation, floodplain sedimentation and sediment redistribution at river bifurcations are important. Processes which can affect downstream fining over the entire river reach are dredging, aggradation, dumping and tributary confluences.
- Determination of the exact combined influence of these factors requires the construction of a detailed, fraction-wise sediment balance, in which apart from the addition, extraction and redistribution processes also in-channel processes as selective transport and abrasion are quantified.

7.4 Gravel-sand transitions

- Gravel-sand transitions can be caused by preferential abrasion of fine gravel, by large amounts of tributary sand supply and by the sudden deposition of suspended sand
- Another plausible explanation is that the pavement suddenly breaks up when the sand content of the bed exceeds a critical value, after which patches are formed. The patchiness retards the transport of coarse grains, but causes a rapid transport of fine grains. This causes bed degradation, which in turn leads to a sudden decrease in bed slope. The slope reduction further hinders the downstream transport of gravel, amplifying the effects of patchiness.

7.5 Downstream fining models

- Empirical (exponential) downstream fining models are especially suitable for the description of downstream fining patterns in river reaches in which sediment addition, extraction and redistribution processes only have a minor effect.
- Empirical downstream fining models can help to determine the relative importance of abrasion and selective transport.
- Diameter reduction coefficients are on average a factor 10 larger than laboratory-determined abrasion coefficients, indicating that abrasion is much less important than selective transport in most rivers.
- The diameter reduction coefficient decreases with river length. This is because longer river have a lower slope, and thus a lesser degree of selective transport. The relative importance of abrasion thus must be larger in longer rivers.
- Present physical downstream fining models are all based on the sediment continuity equation. Most of them use the Hirano-active layer concept to calculate the sediment exchange between bed load and bed. Differences between the models result from the particular choice for the active layer thickness, from the way sediment transport and water flow are computed, from the specification of boundary conditions and initial conditions and also from the way the composition of the material that is exchanged between the active layer and the substrate is computed.
- The present generation of physical downstream fining models incorporates the effects of selective transport, patchiness, bed erosion and aggradation, tributary confluences, dumping and dredging and floodplain sedimentation in a satisfactory way.
- The available models do not have proper formulations for abrasion, armouring, bed form sorting, non-alluvial sediment additions, groyne field processes and bank erosion/point bar formation.
- For bed form sorting, recently promising formulations have been proposed, so in future the effects of sorting in bed forms can be incorporated in downstream fining models.
- From physical downstream fining models information can be obtained about the relative importance of different downstream fining processes by performing sensitivity analyses.

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Appendix 1: Hiding-exposure coefficients

Author	Method[2]	River	D50 [1] [mm]	Di /D50 range [-]	t value [-]
Komar (1987)[5]	LG	Oak Creek	63*	0.0	-0.43
	LG	Oak Creek	63*	0.4 - 6	-0.53
	LG	Great Eggleshope Beck	62*	0.6 - 10	-0.68
	LG	Great Eggleshope Beck	62*	0.6 - 10	-0.64
	LG	West Solent (tidal currents)	16*	1 - 5	-0.71
Komar & Carling (1991)[5]	LG	Oak Creek	63*		-0.64
	LG	Great Eggleshope Beck	62*		-0.82
	LG	Oak Creek	20**		-0.64
	LG	Great Eggleshope Beck	27**		-0.82
Lepp *et al.* (1993)[5]	LG		91*		-0.40
	LG		84*		-0.50
	LG		114*		-0.47
Wathen *et al.* (1995)	LG	Allt Dubhaig	21*		-0.70
Gladkow *et al.(* 2000)		Donau		0-0.4	-1.00
		Donau		1.0-8	-0.90
Kuhnle (1992)	RT	Goodwin Creek	8.3**	0.04-5	-0.81
Kuhnle (1993)	RT	Flume	0.5**		-1.02
	RT	Flume	5.6**		-1.10
	RT	Flume	0.5**	0.45-2.5	-1.08
	RT	Flume	0.5**	3.6 - 20.4	-0.32
	RT	Flume	0.6**	0.37 - 3	-1.03
	RT	Flume	0.6**	4.2 - 11.9	-0.55
	RT	Flume	1.0**	0.22 - 1.2	-1.04
	RT	Flume	1.0**	1.8 - 7.1	-0.42
	RT	Goodwin Creek	8.3**	0.04 - 0.34	-0.94
	RT	Goodwin Creek	8.3**	0.67 - 5.4	-0.81
Wathen *et al.* (1995)	RT	Allt Dubhaig	13*		-0.90
Day (1980)[3]	RT (variant)	Flume	1.8	1-7	-0.29
	RT (variant)	Flume	0.4	1-7	-0.53
Day (1980)[5]	RT (variant)	Flume	1.8**		-0.81
	RT (variant)	Flume	1.6**		-0.95
Parker *et. al.* (1982)	RT	Oak Creek	20**	0.045-4.2	-0.98
Batalla & Martin-Vide (2001)	RT	Arbucies river	2.2**	0.15-5.45	-0.92
Wilcock (1987)[5]	RT	Flume	1.8**		-1.00
	RT	Flume	1.8**		-0.97
	RT	Flume	0.7**		-0.98
	RT	Flume	5.3**		-1.10
Wilcock (1992)[5]	RT	Flume	2.6**		-1.04
	RT	Flume	2.0**	0.27 - 0.39	-1.25
	RT	Flume	2.0**	2.1 - 3.1	-1.14
	RT	Flume	0.8**	0.7 - 1.0	-0.73
	RT	Flume	0.8**	5.8 - 8.3	-1.17
Wilcock (1993)	RT	Flume	1.9**		-0.28
Wilcock & McArdell (1993)[5]	RT	Flume	5.3**		-0.45
Misri *et al.* (1984)[5]	RT	Flume	2.4**		-1
	RT	Flume	3.8**		-0.95
	RT	Flume	4.0**		-0.92
Dhamotharan	RT	Flume	2.16**		-1.1
Milhous (1973)[5]	RT	Oak Creek	19.5**		-0.98

[1] The bed layer which D50 has been used is indicated with asterisks. * : bed surface, ** bed sub-surface or laboratory grain size distribution.
[2] LG = largest grain method, RT = reference transport method, VO = visual observation method.
[3] Op.cit. Komar 1997.
[4] Including East Fork River and Snake river.
[5] Op. cit. Buffington & Montgomery (1997).

Author	**Method**[2]	**River**	**D50** [1] **[mm]**	**Di /D50 range [-]**	**t value [-]**
Andrews (1983)	LG	Clearwater River[4]	74**	0.3-4.2	-0.87
Andrews & Erman (1986)[5]	LG	Sagehen Creek	58*		-1.07
	LG	Sagehen Creek	30**		-1.07
Ashworth *et al.* (1992)[5]	LG		21*		-0.69
Batalla & Martin-Vide (2001)	LG	Arbucies river	2.2**	1.5-20	-0.68
Carling (1983)[5]	LG	Great Eggleshope Beck	62*		-0.46
	LG	Great Eggleshope Beck	77*		-0.42
Ferguson (1994)[5]	LG	Roaring River	72*		-0.87
	LG	Roaring River	77*		-0.88
	LG	Roaring River	140*?		-0.89
	LG	Roaring River	106*		-0.78
	LG	Roaring River	75*		-0.69
Ferguson et al. (1989)[5]	LG	White River	73*		-0.88
Hammond *et al.* (1984)[5]	LG	West Solent (tidal currents)	16*		-0.60
Komar (1987)	LG	Oak Creek	20	0.4-6	-0.68
	LG	Great Eggleshope Beck	20	0.6-10	-0.68
	LG	West Solent (tidal currents)	7.5	1-5	-0.71
Petit (1994)[5]	VO	Flume	12.8*		-0.66
	VO	Flume	24.2*		-0.73
	VO	Flume	39.2*		-0.81
	VO	Flume	19.6*		-0.68

Appendix 2 Symbols

a	A constant [N/m^2]
D	Grain size [m]
D_*.	Dimensionless grain size [-]
D_x	Representative mixture grain size [m]
D_{50}	Median grain size of the entire mixture [m]
D_{avg}	Average grain size of the entire mixture [m]
D_{65}	65th percentile of the mixture grain size distribution [m]
D_{90}	90th percentile of the mixture grain size distribution [m]
D_i	Average grain size of size fraction *i* [m]
$D_{roughness}$	Size of roughness grains [m]
E_i	Volume fraction content of size fraction *i* of the material that is exchanged between the active layer and the substrate [-]
F_i	Volume fraction content of size fraction *i* in the active layer [-]
f_i	Volume fraction content of the *i*th size fraction in the bed (subsurface) [-]
$f_{i,bl}$	Volume fraction content of the i^{th} grain size fraction in the bed-load [-]
$f_{i,a}$	Volume fraction content of grain size fraction *i* in the sediment mixture of patch *a* [-]
$f_{i,b}$	Volume fraction content of grain size fraction *i* in the sediment mixture of patch *b* [-]
g	Acceleration due to gravity [m/s^2]
$k_{a,D}$	Diameter reduction abrasion coefficient [m^{-1}].
$k_{a,w}$	Weight-reduction abrasion coefficient [m^{-1}]
k_D	Overall diameter reduction coefficient [m^{-1}]
$k_{s,d}$	Diameter reduction coefficient due to selective transport [m^{-1}]
J_i	Relative mobility function [-]
L	Distance in downstream direction [m]
L_a	Thickness of the active layer [m]
L_i	Step length [m]
p_i	Volume fraction content of size fraction *i* in the transported material [-]
p_a	Fraction of the total bed surface covered with patches of type *a*.
p_b	Fraction of the total bed surface covered with patches of type *b*.
q_i	Transport rate of size fraction *i* [m^2/s]
q_T	Volume of sediment transport per unit width and time [m^2/s]
q^*_i	Dimensionless transport rate [-]
$q_{lat,i}$	Lateral inflow of sediment [m^2/s]
$q_{T,a}$	Transport rate of uniform material with a grain size equal to the D_{50} of patch a [m^2/s].
$q_{T,b}$	Transport rate of uniform material with a grain size equal to the D_{50} of patch b [m^2/s].
Re_*	Particle Reynolds number [-]
s_i	Distance moved during an flood period by grains of size *i* [m]
s_{D50}	Distance moved during an flood period by grains which size is equal to D_{50} [m]
t	Hiding-exposure coefficient [-]
u^*	Shear velocity [m/s]
w	Particle weight [N]
w_i	Fall velocity of grains of size *i*
W_i	Dimensionless transport rate [-]
z_0	Elevation of the interface between the active layer and the substrate [m]

θ_c	Dimensionless mobility parameter of Shields [-]
$\theta_{c,\ i}$	Dimensionless critical shear stress for size fraction *i* [-]
λ	Bed porosity [-]
ρ	Fluid density [kg/m^3]
ρ_s	Particle density [kg/m^3]
τ	Bed shear stress [N/m^2]
τ_c	Critical bed shear stress [N/m^2]
$\tau_{c,i}$	Critical bed shear stress for grain size fraction *i* [N/m^2]
$\tau_{si,\ non\text{-}uniform}$	Shear stress for the beginning of suspension of size fraction *i* in a non-uniform mixture [m]
$\tau_{si,\ uniform}$	Shear stress for the beginning of suspension of size fraction *i* in an uniform mixture [m]