9 Physical Modelling in Fluvial Geomorphology: Principles, Applications and Unresolved Issues

Jeff Peakall
Department of Earth Sciences and School of Geography, University of Leeds, UK

Phil Ashworth
School of Geography, University of Leeds, UK

Jim Best
Department of Earth Sciences, University of Leeds, UK

ABSTRACT

The relationships between fluvial process and form are often extremely difficult to quantify using conventional field and numerical computational techniques. Physical modelling offers a complementary technique to these methods and may be used to simulate complex processes and feedbacks in many geomorphic phenomena. Depending on the temporal/ spatial scale of a particular research problem, physical models may be either 1: 1 replicas of the field prototype, scale with Froude number only, have distorted scales or serve as unscaled experimental analogues that attempt to reproduce some properties of a prototype. This chapter presents a critique of the underlying principles that determine the degree to which physical models accurately replicate the form and dynamics of natural alluvial systems. Examples are presented of each modelling technique to illustrate both the advantages and inherent limitations of these different approaches and highlight the contribution of physical modelling in the study of fluvial geomorphology and sedimentology.

Three issues are identified for achieving significant progress in the scale modelling of fluvial systems: (i) incorporation into models of variables such as multiple time scales, flood hydrographs, fine-grained sediment, cohesion, surface tension and floodplain

The Scientific Nature of Geomorphology: Proceedings of the 27th Binghamton Symposium in Geomorphology held 27 - 29 September 1996. Edited by Bruce L. Rhoads and Colin E. Thorn.

vegetation which will increase the degree of model realism; (ii) continued development and implementation of a range of measurement techniques; and (iii) detailed model: prototype verification across a range of scales. Whilst these steps will increase significantly the power and attractiveness of scale modelling in the earth sciences, simple analogue models will continue to enable testing of new concepts across the full range of spatial and temporal scales.

INTRODUCTION

Many problems in fluvial geomorphology involve complex, multivariate situations, often at large spatial and temporal scales (see Kirkby, Chapter 10 this volume). These topics have traditionally been addressed through detailed fieldwork combined with theoretical and numerical modelling. Whilst mathematical models have promoted major advances in our understanding of the complex interrelationships involved in sediment production, transfer and deposition in dynamic fluvial environments (cf. Pickup 1988; Ikeda and Parker 1989; Kirkby 1994), they necessarily involve simplifications and use of empirical coefficients derived from limited input data. A complementary technique that has developed in parallel with these computational simulations is physical modelling, which has two principal advantages. First, the formative processes can be observed, usually in a reduced time-frame, within a controlled and manageable laboratory environment. Second, physical models may allow incorporation of variables which are not known a priori and which may have markedly non-linear effects on the resultant dynamics or morphology. However, these advantages are counterbalanced by prototype to model scaling difficulties which result in increasing simplification and abstraction from reality as spatial and temporal scales increase. Additionally, it is clearly important to establish and quantify the influences of processes which may be non-linear in their scaling between model and prototype (e.g. particle settling velocity, see p. 233) and their consequent effect on morphology.

Physical modelling techniques can be classified both by their specificity (degree to which the model replicates a prototype) and the temporal/spatial scale at which they are most applicable (see Figure 9.1). For the smallest spatial and temporal scales, a 1:1 replica of flow and sediment dynamics can be re-created in the laboratory with little or no difference from the natural prototype. These models have, for example, been instrumental in investigating the morphology and controlling variables of bedform generation both in sands and gravels (e.g. Guy et al. 1966; Allen 1982; Southard and Boguchwal 1990a). However, even in these 1:1 models, care must be taken in considering temperature/ viscosity influences (Southard and Boguchwal 1990b), applying such flume results to much deeper natural flows (Williams 1970; Southard 1971), and accounting for the influence of sidewalls on the experimental results (Crickmore 1970; Williams 1970). At large spatio-temporal scales, prototypes must be scaled down both to compress the time scale and allow the model to be accommodated within the constraints of available laboratory space. For the largest prototypes, a true scaled modelling approach becomes untenable and purely 'analogue' models must be employed. However, when viewed in the light of imposed modelling constraints, these large-scale models (e.g. studies of base level controls on fluvial incision) can provide invaluable insights into the behaviour of complex natural phenomena.

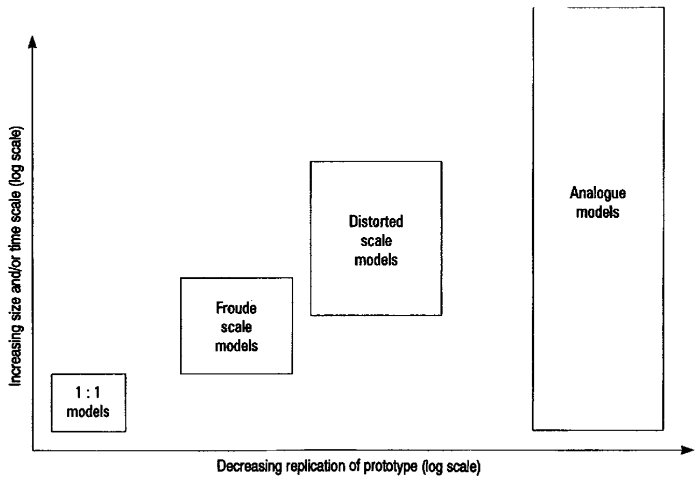

Figure 9.1 Schematic view of the balance between model specificity and spatial/temporal scales for different modelling techniques. There is an overall decrease in the replication of prototype characteristics from 1:1 models through Froude scale models (FSMs), distorted scale models, to analogue models. It should, however, be noted that the modelling of a single parameter (e.g. sediment transport) within a distorted scale model can be more accurate than in an FSM. The spacing of the boxes is schematic, but illustrates two key points. First, there is a significant decrease in replicability when moving from 1:1 to scaled models, and from scaled to analogue models. In contrast, the transition from FSM to distorted scale models is associated with a smaller loss of model replicability. Secondly, the chosen spatial and temporal scales for the 1:1, FSM and distorted models illustrate the relative size at which the modelling techniques are generally used

Early attempts to model fluvial and coastal processes include the pioneering work of Fargue (reported in Zwamborn 1967), Thomson (1879), Reynolds (1887) and Gilbert (1914, 1917). Although Fargue and Reynolds scaled some key controlling variables (e.g. the horizontal and vertical distance and tidal period by Reynolds), it was not until the development of dimensional analysis by Buckingham (1915) that scale modelling techniques in engineering were widely adopted (e.g. ASCE 1942; Murphy 1950). Since these two benchmark publications, numerous physical modelling texts have been published including the influential works of Yalin (1971) on scaling theory, Franco (1978) and Shen (1991) on movable-bed modelling, and Schumm et al. (1987) on a range of analogue modelling techniques.

It is now widely accepted across a range of disciplines that physical modelling offers a number of advantages to the scientist interested in a number of landscape evolution processes (see Hooke 1968; Mosley and Zimpfer 1978; Ashmore 1982; Warburton and Davies, in press). Physical models have been used successfully to investigate various issues in fluvial geomorphology over a range of scales, including:

1. Confluence morphology (e.g. Mosley 1976; Ashmore and Parker 1983; Best 1988; Ashmore 1993);
2. Fluvial sediment transport (e.g. Ashmore 1988, 1991a; Ashworth et al. 1992a; Hoey and Sutherland 1991; Young and Davies 1991; Warburton and Davies 1994a);

3. Bar deposition and migration (e.g. Ashmore 1982; Southard et al. 1984; Lisle et al. 1991; Ashworth 1996);
4. Channel change (e.g. Davies and Lee 1988; Leddy et al. 1993; Ashmore 1991b);
5. Channel pattern development (e.g. Leopold and Wolman 1957; Schumm and Khan 1972; Ashmore 1991b);
6. River response to changing extrinsic variables such as tectonics (e.g. Ouchi 1985; Jin and Schumm 1987), aggradation (e.g. Ashworth et al. 1994; Peakall 1995) and base level (e.g. Wood et al. 1993, 1994; Koss et al. 1994).

With such a great variety of physical modelling applications and their increasing use both in geomorphology and sedimentology, it is now appropriate to evaluate the underlying principles that ensure accurate prototype-model scaling and present an assessment of the degree to which replication of the prototype is achieved in such models.

This chapter briefly reviews the theoretical basis of scale modelling to provide a context for identifying the key issues that must be addressed before scale modelling can achieve its full potential. Examples from several models are used to illustrate both the widespread appeal of physical modelling and the progressive decrease in model replicability with increasing ratio of prototype: model scales (Figure 9. 1). Traditional 1:1 hydraulic flume models, which have been used widely at the smallest scales of geomorphological interest (Figure 9. 1), will not be reviewed here. Examples of such modelling approaches and their application within fluvial environments are contained in Allen (1982), Southard and Boguchwal (1990a) and Best (1996). This chapter instead concentrates on the scales of geomorphological interest ranging from the river channel to the drainage basin. Unresolved issues in the scaling of such systems are discussed, together with an appraisal of future developments which may greatly increase the use and application of scaled physical models within the earth sciences.

CLASSIFICATION OF PHYSICAL MODELS

At the simplest level, physical models of rivers have been traditionally classified on the grounds of specificity, i.e., how closely they replicate a prototype, and by the controlling boundary conditions (Chorley 1967; Schumm et al. 1987). Two types of boundary condition are recognised in the engineering literature: fixed-bed studies, which have nonerodible boundaries and no sediment transport, and movable-bed experiments where the substrate is free to move within a constrained or unconstrained channel. The majority of geomorphological models have movable beds, with either constrained channels (e.g. most bedform studies) or unconstrained channels where the edges of the experimental apparatus serve as the ultimate constraint (e.g. most river/channel network models).

Towards one end of the specificity continuum (Figure 9.1), scale models attempt to represent exactly some, or all, of the key parameters of the system, either from a specific prototype or from general values. Scale models are based on similarity theory, which produces a series of dimensionless parameters that fully characterise the flow. In an idealised situation every variable should be perfectly scaled in the model; however, in the majority of experiments it is not possible to fulfil this requirement. Consequently, the flow Reynolds number is relaxed (see discussion on p. 227) while remaining in the fully turbulent flow regime, but the Froude number is scaled correctly. Relaxation of the flow

Reynolds number allows more flexibility in the model scaling than variation in the Froude number which must be far more tightly constrained. This technique is known as Froude scale modelling (FSM) and has been used successfully in movable-bed modelling of river anabranches and in fixed-bed modelling of flow interaction with artificial structures such as spillways, conduits and breakwaters (e.g. French 1985; Owen 1985). A perfect FSM must achieve geometric, kinematic (motion) and dynamic (force) similarity between model and prototype. FSM studies may model either specific prototypes or scaled versions of a general geomorphic feature. The latter class has been referred to as 'generic Froude scale models' (Church quoted in Ashmore 1991a, b; Ashworth et al. 1994; Warburton and Davies, in press).

For the study of large-scale geomorphological features, such as estuaries and major rivers (e.g. Novak and Cabelka 1981; Klaassen 1991), model geometry may be distorted by increasing the vertical to horizontal scaling ratio, which enables small models to be built or large prototypes to be studied. To achieve precise modelling of sediment transport, a supplementary slope is often added and changes are made to the velocity and discharge scales (e.g. Franco 1978). It is not possible to fully predict the magnitudes of these various adjustments and therefore a process of verification is used, whereby variables are systematically altered until the model reproduces changes observed in the prototype. In this chapter, these models are referred to as 'distorted' models, but it should be noted that similarity of Froude number is also generally achieved in these experiments.

At the other end of the specificity continuum, models can be considered as small landforms in their own right and have been referred to as 'similarity of process' models (Hooke 1968). These models must obey gross scaling relationships and reproduce certain features of the prototype, but since the model is not scaled from either a specific prototype or from generic data, none of the model processes which can be quantified may be applied directly to specific field examples. In this chapter, unscaled, similarity of process studies are referred to as 'analogue' models. The term 'analogue' (Chorley 1967) has been used previously to refer to models that reproduce certain features of a natural system even though the driving forces, processes, materials and geometries may differ from the original. Analogue models are most applicable to the largest prototype:model scales where even distorted scale models cannot be applied successfully (Figure 9.1).

PRINCIPLES OF SCALE MODELLING

An overview of the underlying principles of scale modelling and a definition of key variables is necessary in order to discuss some of the different approaches that have been adopted. Since these scaling laws and their derivation are comprehensively reviewed by Yalin (1971), Langhaar (1980) and covered in detail in many other texts (e.g. Henderson 1966; French 1985; Chadwick and Morfett 1986), only a brief review is given here.

Basic Scaling Laws

Two examples of open-channel flow are used to derive, by dimensional analysis, the most important scaling parameters for physical models. The first and simplest example is that of fixed-bed modelling since only the flow and boundary parameters need to be considered.

The second example of movable-bed modelling is complicated by the additional consideration of sediment transport.

Fixed-bed modelling (case without mobile sediment)

In order to use dimensional analysis, the quantities that control a given system must first be selected and expressed in terms of their fundamental units. For open channel flow with a fixed bed, these controlling variables are usually taken as (Yalin 1971):

- properties of the fluid - the dynamic viscosity (μ) and density (ρ)
- boundary conditions of the channel, normally hydraulic radius *(R)* and surface roughness (k_s)
- bed slope (S)
- average downstream velocity (U) and
- gravitational constant (g)

Three governing variables must be chosen to obtain a solution from these seven variables (μ, ρ, R, k_s, S, U and g) using dimensional analysis. These principal variables are generally taken as ρ, R and U since they generate two key flow parameters, the Froude and flow Reynolds numbers. The method generates n - 3 dimensionless terms where n is the number of variables and 3 is the number of governing variables. These are referred to as 'pi' (Π) terms. In the example considered here, four terms are produced:

$$\Pi_1 = \frac{\rho R U}{\mu} = Re \tag{1}$$

$$\Pi_2 = \frac{U}{\sqrt{gR}} = Fr \tag{2}$$

$$\Pi_3 = \frac{k_s}{R} \tag{3}$$

$$\Pi_4 = S \tag{4}$$

The four Π terms represent the flow Reynolds number (Π_1), the Froude number (Π_2), the relative roughness (Π_3) and the channel bed slope (Π_4). If the ratio between prototype and model is kept identical for all four of these terms, the model would be an exact representation of the prototype. However, this situation is rarely attainable in hydraulic modelling as illustrated by considering the Froude and Reynolds numbers. Since water is used in most experimental studies, the density and viscosity of the fluid are the same in the model ($_m$) and prototype ($_p$), assuming a constant temperature, and therefore the Reynolds number can be rearranged to give

$$U_p R_p = U_m R_m \tag{5}$$

and therefore

$$\lambda_u = \frac{U_m}{U_p} = \frac{R_p}{R_m} \tag{6}$$

where λ_u is the scaling ratio of velocity.

In the case of the Froude number, since acceleration due to gravity, *g,* remains constant for both model and prototype, then

$$\frac{U_{\mathrm{m}}^{2}}{\mathrm{U}_{\mathrm{p}}^{2}}=\frac{R_{\mathrm{m}}}{R_{\mathrm{p}}} \tag{7}$$

and therefore

$$\lambda_{\mathrm{u}}=\frac{U_{\mathrm{m}}}{U_{\mathrm{p}}}=\left(\frac{R_{\mathrm{m}}}{R_{\mathrm{p}}}\right)^{0.5} \tag{8}$$

Equations (6) and (8) can only be resolved if R_{m} is equal to R_{p}. Thus, the flow Reynolds number is commonly relaxed, with the proviso that in the case of open channel flows it remains within the fully turbulent flow regime ($Re > 500$). *As* noted previously, this form of modelling is referred to as Froude scale modelling (FSM).

Movable-bed modelling (case with mobile sediment)

If a movable rather than fixed bed is considered, the flow can be considered as a two-phase flow with both fluid and particles. The following set of parameters is used to describe these flows: μ, ρ, S, R, g and two parameters which describe the sediment, ρ_{s} (the sediment density) and D (the characteristic grain size of the sediment). Some of these variables can be replaced by other dependent parameters. For example, the shear velocity $U_* = (gRS)^{0.5}$ can replace S and the immersed specific weight of grains in the fluid $\gamma_{\mathrm{s}} = g(\rho_{\mathrm{s}} - \rho)$ can replace g giving μ, ρ, R, D, ρ_{s}, U_* and γ_{s}, These variables also produce n - 3 or 4Π terms

$$\Pi_1=\frac{R}{D} \tag{9}$$

$$\Pi_2=\frac{\rho_{\mathrm{s}}}{\rho} \tag{10}$$

$$\Pi_3=\frac{\rho U_* D}{\mu}=Re_* \tag{11}$$

$$\Pi_4=\frac{\rho U_*^2}{\gamma_{\mathrm{s}} D} \tag{12}$$

The Π_1 and Π_2 terms represent relative roughness of the sediment and relative density respectively, while the term Π_3 is the grain Reynolds number (Re_*), which is a measure of the roughness of the bed relative to the thickness of the viscous sub-layer. Equation (12) expresses the Shields relationship which is normally rearranged as

$$\tau_*=\frac{\tau}{(\rho_{\mathrm{s}}-\rho)gD} \tag{13}$$

where τ is the bed shear stress responsible for initiating sediment transport for a particular grain size, D, and τ_* is the dimensionless shear stress. Together, τ_* and Re_* form the axes of the Shields entrainment diagram (Figure 9.2). The scatter of points on the Shields diagram may be used to define an entrainment threshold known as the critical shear stress, τ_c^*, above which a flow is capable of transporting sediment. The Shields diagram also shows τ_c^* becoming constant (approximately 0.056) at high values of Re_*. Recent debate on the critical threshold value of Re_*, where the flow may be deemed to be fully rough and turbulent with the minimal effect of viscous forces (Re_{*crit}), is discussed on p. 228, but, it should be noted that this value occurs in the range $Re_* = 5$ to > 70. It has been proposed that Re_{*crit} may be used to define the optimal length scale, λ_L, for modelling (Yalin 1971):

$$\lambda_L = \left(\frac{Re_{*crit}}{Re_{*p}} \right)^{2/3} \tag{14}$$

where Re_{*p} refers to the prototype. When modelling a two-phase flow, the specific dimensionless properties of the fluid, Re and Fr (equations (1) and (2)), must also be satisfied.

UNRESOLVED ISSUES IN FROUDE SCALE MODELLING

Flow and Hydraulic Constraints

Grain Reynolds number Re_*

The principal criterion for choosing the optimal length scale for an FSM study is a function of the ratio between the prototype Re_* and the minimum Re_* for a fully rough flow field (equation (14)). There is, however, still debate on the interpolation and interpretation of the original Shields diagram (Kennedy 1995). Rouse (1939) added a threshold line to a form of the Shields diagram and later established a critical value of $Re_* = 400$ (Rouse 1950). Subsequent workers have redrawn the line so that it is an asymptote with constant τ_c^* at $Re_* > 350$ (Richards 1982) and 1000 (Henderson 1966; Novak and

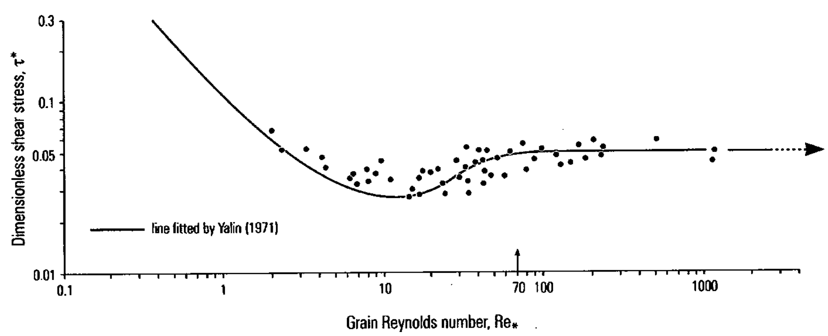

Figure 9.2 The Shields curve as plotted by Yalin (1971, Figure 6.2, p. 154). Note the high degree of scatter amongst the data points which makes the interpolation of the trend line highly subjective. The flow becomes fully rough (i.e. the Shields line becomes horizontal) at a value of 70 using this interpolation. After Ashworth et al. (1994) and reproduced by permission of John Wiley and Sons Ltd, from *Process Models and Theoretical Geomorphology,* edited by M.J. Kirkby, copyright 1994, John Wiley and Sons Ltd

Cabelka 1981). Earlier work by Nikuradse (1933) on pipe-flow boundary conditions produced a similar, but much better constrained plot to that of Shields, with the fully rough category being defined as $Re_* > 70$ (Schlichting 1968). Yalin's (1971) version of the Shields diagram (Figure 9.2) is also an asymptote with constant τ_c^* at $Re_* > 70$ (see Yalin 1971, Figure 3.3, p. 58) and this value has been adopted in the majority of recent FSM studies (e.g. Ashmore 1982; Davies and Lee 1988; Young and Davies 1990; Ashmore 1991a, b; Hoey and Sutherland 1991; Warburton and Davies 1994a).

Parker (1979) divorced the concepts of boundary roughness and constant τ_c^* by arguing that the flow becomes hydraulically rough if $Re_* > 15$. Ashworth et al. (1994) also note that the scatter in the Shields diagram allows for either a 'dip' or a horizontal line to be plotted from values greater than $Re_* = 5$ and argue that the minimum model grain Reynolds number should be 15 (Ashworth et al. 1994, p. 119). Jaeggi (1986) advocates that the minimum value of Re_* that can be used in models is 5 since ripple formation occurs with lower grain Reynolds number. However, Jaeggi (1986) also notes that Re_* values in the transitional field (i.e. before a constant τ_c^* is reached) incorrectly reproduce the initial sediment entrainment conditions.

Prototype verifications of FSMs rarely include Re_* data from anabranches or parts of channels that contain fine-grained sediment or shallow water depths (e.g. backwaters, lateral bars, bank and bartops, and reactivated abandoned channels) although such locations may be expected to contain smooth or transitional Re_* (see calculations in Ashworth and Best, in press). The influence of transitional roughness on the morphological and sediment transport characteristics of many models (Jaeggi 1986), as well as more careful consideration of the range of Re_* found in the field, is clearly a central issue which must be addressed in future FSM studies.

Transitional and supercritical flow

Many braided river FSM studies produce large trains of standing waves indicating areas of supercritical flow (see Figure 9.3; Table 9.1). Field studies also indicate that braided channels may possess Froude numbers greater than unity (e.g. Fahnestock 1963; Williams and Rust 1969; Boothroyd and Ashley 1975; Bristow and Best 1993), but these conditions appear to be less temporally and spatially extensive in the field than in many models. Transitional ($500 > Re > 2000$), supercritical ($Fr > 1$) flow conditions are also evident in many FSM studies (e.g. Ashmore 1982, 1993; Ashworth et al. 1994; Peakall 1995) as highlighted by the presence of oblique rhomboidal standing waves and associated lowrelief bedforms (see Figure 9.4, after Karcz and Kersey 1980). This situation suggests that the present scaling ratios based on Re_* may need revision, since model Froude numbers may be too high and flow Reynolds numbers too low. There are at least two possible explanations for this discrepancy.

1 .The high Froude numbers in the model may be attributable to incorrect scaling of the bed roughness which may considerably influence the velocity distribution. A predominance of supercritical flow in the model may be explained by velocities being too high because there is less skin roughness when compared to the field prototype. The absence of the very coarsest fractions of the grain size distribution in the model ($> D_{95}$) may be a contributing factor as may the use of rounded/subrounded sand as representative of the field sediment which will lead to a marked drop in flow

Figure 9.3 View of a train of standing waves along the thalweg of the main 'active' channel in a 1: 20 braided gravel-bed river FSM study (Ashworth et al. 1992b). Flow is from top to bottom and the braidplain is approximately 1 m wide. Avulsion of the main channel has left an abandoned channel complex along the right bank and flow is concentrated into one dominant channel which is eroding the outer left bank at the beginning of full braidplain development

Table 9.1 Compilation of key hydraulic variables from recent FSM studies

Paper	*We*	*Fr*	*Re*	Re_*
Ashmore (1991a)	ND	0.56-0.93	920-4760	36-103
Ashmore (1991b)	11-65[a]	0.91-1.30	1893-6870	89-140
Ashworth et al. (1994)	ND	0.43-0.6[b]	1885-2632[b]	18-56[b]
Ashworth (1996)	8-132[a]	0.64-1.89	3580-14500	178-586
Hoey and Sutherland (1991)	5[a,c]	0.44-1.54	ND	32-141
Warburton and Davies (1994a, b)	1-16[d]	0.36-1.20[d]	2230-3400[c]	42-106[d]
Young and Davies (1990)	ND	0.30-0.60	ND	76-115[e]

[a]For papers that do not quote water temperature, variables are calculated assuming a value of 15 °C.
[b]Values calculated using average maximum depths and average velocities.
[c]Initial condition only.
[d]Range is calculated based on two standard deviations from the mean.
[e]From Young (1989).
For Re_* calculations, the D_{50} is used by Ashmore (1991a); the D_{90} by Young and Davies (1990), Ashmore (1991b), Ashworth et al. (1994) and Warburton and Davies (1994a, b); and the $D_{100}/1.8$ by Hoey and Sutherland (1991).
ND - No data quoted.
We= Weber number (see equation 18).

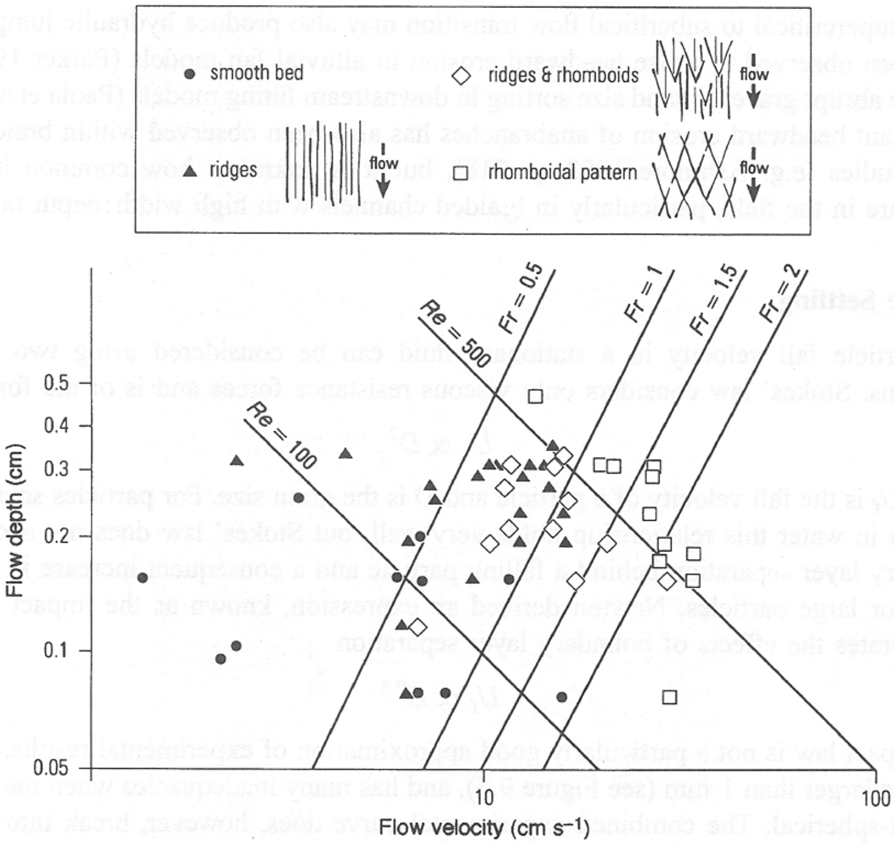

Figure 9.4 Relationship between flow Reynolds number *(Re),* Froude number (*Fr*) and the characteristic low-relief bedforms often found in physical models of braided rivers. Note the predominance of ridges and rhomboidal bedforms at transitional flow Reynolds number ($500 < Re > 2000$) and supercritical ($Fr > 1$) flow. Plot is redrawn from Karcz and Kersey (1980)

resistance and particle interlocking (see Church et al. 1991 and discussion below concerning sediment transport). Clearly, other factors such as discharge and slope may also be involved in this *Re, Fr* and Re_* relationship.

2. The choice of a critical grain Reynolds number may be inappropriate since both Nikuradse's (1933) work on relative roughness in pipes and the Shields curve use sediment with a uniform grain size distribution. Their plots therefore have Re_* numbers calculated using a unimodal grain size distribution whilst the majority of FSM experiments calculate Re_* using the D_{90} of a heterogeneous grain size distribution (see Table *9.1).*

The presence of low particle Reynolds numbers ($Re_* < 5$) in an FSM may lead to the formation of ripples which may not have scaled equivalents in the field prototype (Jaeggi 1986). Changes in morphology of these small, smooth-boundary bedforms in fluctuating local flow depths may also lead to the presence of standing waves and supercritical flow bedforms in the FSM Jaeggi (1986) suggests the need to coarsen the bed material in an FSM to avoid formation of these bedforms which, although possible, may be contradictory to other model study objectives if the fine tail of the grain size distribution is of interest.

The supercritical to subcritical flow transition may also produce hydraulic jumps which have been observed to cause headward erosion in alluvial fan models (Parker 1996) and produce abrupt gravel to sand size sorting in downstream fining models (Paola et al. 1992). Significant headward erosion of anabranches has also been observed within braided river FSM studies (e.g. Ashmore 1982, p. 218), but it is unknown how common hydraulic jumps are in the field, particularly in braided channels with high width: depth ratios.

Particle Settling

The particle fall velocity in a stationary fluid can be considered using two different equations. Stokes' law considers only viscous resistance forces and is of the form

$$U_{\mathrm{f}} \propto D^2 \tag{15}$$

where U_{f} is the fall velocity of a particle and D is the grain size. For particles smaller than 0.1 mm in water this relationship holds very well, but Stokes' law does not account for boundary layer separation behind a falling particle and a consequent increase in the fluid drag. For large particles, Newton derived an expression, known as the impact law, that incorporates the effects of boundary layer separation

$$U_{\mathrm{f}} \propto D^{0.5} \tag{16}$$

The impact law is not a particularly good approximation of experimental results, even for particles larger than 1 mm (see Figure 9.5), and has many inadequacies when the particles are non-spherical. The combined experimental curve does, however, break into two distinct linear segments, one characterised by Stokes' law and the other broadly delineated by the impact law. When both model and prototype grain sizes fall exclusively within either of these two areas, a linear scaling ratio can be applied. However, if the prototype grain size is in the 'impact region' and the model is in the field of Stokes' law, then the function is nonlinear and consequently cannot be perfectly scaled. The main result of this nonlinearity is that the fall velocities of the particles relative to the downstream velocity are much slower in the model than the prototype, although the particle time constant (the ratio of the particle response time to the characteristic eddy turnover time, see Elghobashi, 1994) and relationship between particle size and turbulence in the model and prototype must also be taken into account. Saltating particles in the model may consequently have larger hop lengths and heights than the geometric scale ratio would suggest. This limitation could be overcome by altering the grain-size distribution as suggested by Jaeggi (1986) for initiation of sediment movement, but this solution would also affect the mode of sediment transport.

Bedload Transport and Deposition

A scale ratio for sediment transport rate in models with fully turbulent flow, $(\lambda_t)_{\mathrm{s}}$ can be derived by dimensional analysis (Yalin 1963, 1971),

$$(\lambda_t)_{\mathrm{s}} = (\lambda_L)^{1.5} \tag{17}$$

However, Yalin (1971) has demonstrated that if Re_* is below the critical threshold for a fully turbulent boundary and the fluid and temperature are the same in the prototype and

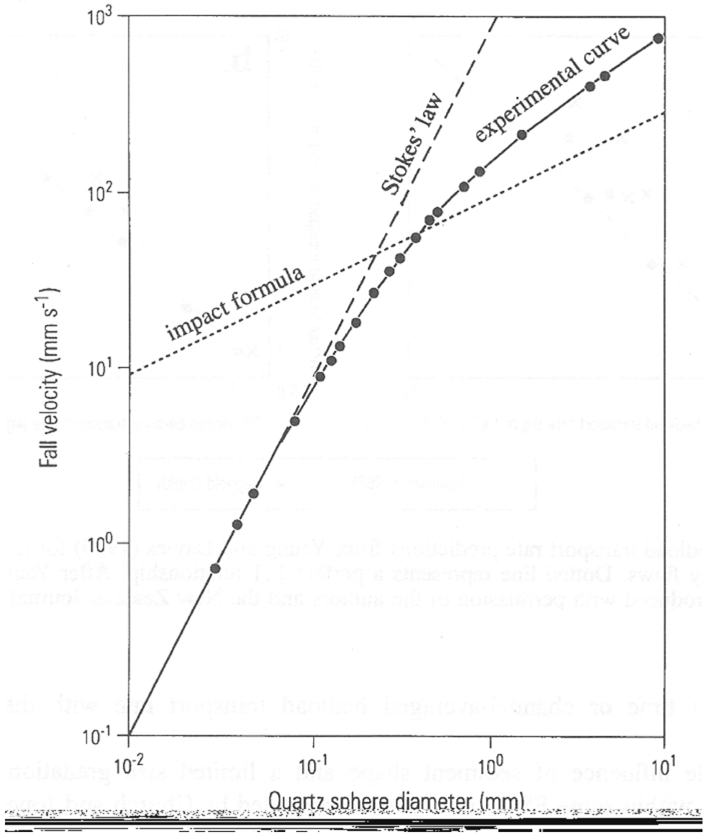

Figure 9.5 Graph of fall velocity as a function of grain diameter, for water at 20 °C, plotted against the predictions of Stokes' law and the impact formula. From Leeder (1982) with the permission of the author using data from Gibbs et al. (1971)

model, then it is impossible to achieve dynamic similarity of sediment transport. The majority of FSM studies fall within the range of suggested critical grain Reynolds numbers (see above; Table 9.1) and therefore may compromise sediment transport similarity.

Several studies have compared the observed bedload sediment transport from an FSM with established transport equations (e.g. Ashmore 1988; Hoey and Sutherland 1989; Young and Davies 1990, 1991; Warburton and Davies 1994a). Young and Davies (1990) compared the empirically based equations of Schoklitsch (1962) and Bagnold, (1980) with their flume data and found a very strong agreement (see Figure 9.6). The Bagnold (1980) equation had the strongest correlation with an average under-prediction of 18% for steady flows and just 1% for unsteady flows. Ashmore (1988) and Hoey and Sutherland (1989) also demonstrated that the Bagnold (1980) formula was in good agreement with model transport rates. The formation of bedload pulses or waves in flumes has also been studied by several authors (e.g. Ashmore 1988; Kuhnle and Southard 1988; Young and Davies 1990, 1991; Hoey and Sutherland 1991; Warburton and Davies 1994a) and the associated short-term variations in sediment transport rates may account for much of the scatter in

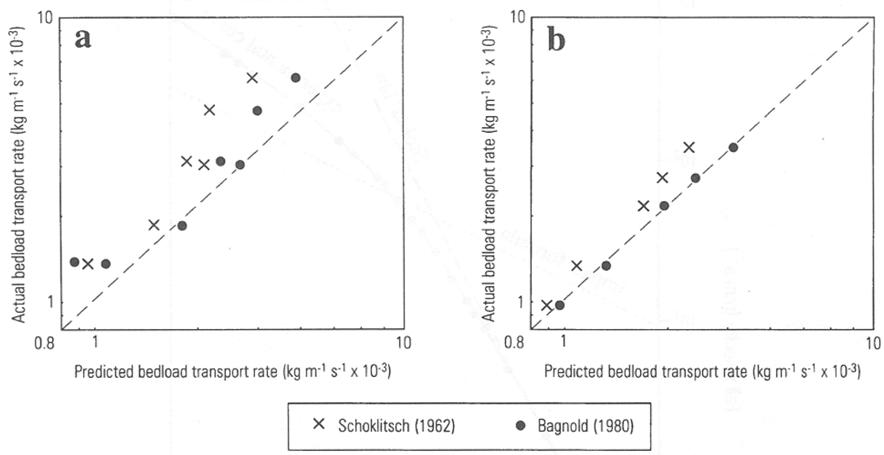

Figure 9.6 Bedload transport rate predictions from Young and Davies (1990) for (a) steady flows, and (b) unsteady flows. Dotted line represents a perfect 1:1 relationship. After Young and Davies (1990), and reproduced with permission of the authors and the New Zealand Journal of Hydrology

correlations of time or channel-averaged bedload transport rate with discharge (e.g. Figure 9.6).

The possible influence of sediment shape and a limited size gradation on bedload transport rates within some FSM studies has been noted by Church and Jones (1982) and Church et al. (1991). Most gravel bed rivers are composed of angular clasts with a very large size range whilst many models use subangular to well-rounded sand grains and do not model the very largest grains. These differences may explain the preferential mobility in many models of the largest grains which tend to roll rapidly across finer-grained substrates ('overpassing') and are preferentially deposited as accreting avalanche faces at bartails and on bartops (e.g. Leopold and Wolman 1957; Ashmore 1982).

Water Surface Tension

Surface tension is the tensile force per unit length (N m^{-1}) acting at the fluid surface. The tensile force results from the difference between the internal molecular forces of a liquid and the forces between liquid molecules and an adjacent surface, and varies as a function of temperature. In rivers, the effects of surface tension are largely insignificant (Dingman 1984, p. 85) although biofilms may help stabilise the sediment surface, but, if a model has too large a vertical scale and consequently too small a flow depth, surface tension can be important. The addition of a surface tension term, a, into the previous dimensional analysis gives a $\prod$ term known as the Weber number (*We*)

$$We = \frac{\rho U^2 h}{\sigma} \tag{18}$$

where *h* is the average flow depth.

The Weber number represents the ratio between the inertial and surface tension forces. The velocity and time scales for perfect scaling of surface tension effects (ASCE 1942) are

$$\lambda_u = \left(\frac{\lambda_\sigma}{\lambda_h \lambda_\rho} \right)^{0.5} \tag{19}$$

$$\lambda_T = \left(\frac{\lambda_h^3 \lambda_e}{\lambda_\sigma} \right)^{0.5} \tag{20}$$

In the case of an FSM, where the fluid density, ρ, is kept the same in both model and prototype and where λ_u is controlled by correctly scaling the Froude number and relaxing the flow Reynolds number, it is not possible to scale the Weber number correctly. However, a similar argument can be made for the Weber number to that used for the relaxation of the flow Reynolds number, namely that so long as the surface tension effects are insignificant then exact scaling is unnecessary. Unfortunately, there is no current consensus on the critical *We* value where surface tension begins to strongly influence sediment transport and deposition, although suggested values range from 10 to 120 (see discussion in Peakall and Warburton, in press). Most small river experiments have Weber numbers that fall within or marginally below the suggested range of critical values (Table 9.1) which suggests that a degree of surface tension induced distortion may have been added to the models. There is therefore a clear need for FSM studies to calculate and publish Weber numbers for a range of channel geometries.

Cohesion and Vegetation

Cohesion

Clay minerals are cohesive due to a combination of two forces, the weak van der Waals' forces which all matter is subject to, and ionic bonds which form through the process of cation exchange between clay minerals. These intermolecular forces act as a major constraint on scale modelling, because coarse sand and gravel in the prototype can only be scaled down as far as silt sizes within the model, without adding cohesion. The cohesive forces of clay are also scale independent so that inclusion of a proportion of clay in the model will lead to unrealistic rates of erosion and channel change. This was demonstrated in the analogue meander model of Schumm and Khan (1972) where the addition of a low concentration of clay caused channel erosion to cease. Consequently, the difficulty in modelling fine-grained sediment limits the length, or geometric scale, of the model and can lead to a truncation of the grain-size distribution. However, it has been shown that it is possible to use inert silica flour as fine as 1 μm as a substitute for prototype silt/fine sand grain sizes (e.g. Parker et al. 1987; Garcia 1993; Leddy et al. 1993; Ashworth et al. 1994). Some of this very fine material may travel solely in suspension or be repeatedly deposited and re-entrained. The use of fine-grained silica flour also leads to significant capillary forces which helps simulate prototype cohesion. This property is desirable in many fluvial models because several geometric variables, such as sinuosity and hydraulic geometry, may change with the degree of cohesion (Schumm 1960).

Vegetation

Although vegetation plays an important role in strengthening the banks and floodplain (e.g. Zimmerman et al. 1967; Smith 1976), particularly in coarse-grained rivers where inter-particle cohesion is largely unimportant, there are very few examples of physical models that incorporate the effects of vegetation. Recent experimental work has documented the interaction between within-channel vegetation and flow structure (e.g. Ikeda and Kanazawa 1995; Tsujimoto 1996) but the significance and magnitude of these effects have yet to be considered within physical scale models. Marsden (1981) experimented with different densities of toothpicks and planted wheat, rape, cress, lawn and budgie seed, before successfully growing mustard on the floodplain of an analogue braided model. The results demonstrate an optimal planting density for maximum floodplain accretion, and Marsden (1981) recommended that the approach be extended to larger models that could incorporate scale effects.

Scaling of Time

One of the primary objectives of physical modelling is to change the rate of the formative processes, thus permitting study of landform evolution over long prototype time periods. Two different approaches to the modelling of time are possible, one based on dimensional analysis and the other on magnitude-frequency analysis.

Dimensional analysis of time scales

The time scale for mean flow velocity $(\lambda_t)_u$ is given by dimensional analysis as $(\lambda_t)_u = (\lambda_L)^{0.5}$. This scale differs from the previously derived time scale for sediment transport, $(\lambda_t)_s$, which has a scale ratio of $(\lambda_L)^{1.5}$ (equation (17)). Similarly, the fall velocity of a particle as characterised by Stokes' law, has a time scale of $(\lambda_t)_{ug} = (\lambda_L)^{-1}$. Yalin (1971) also notes a series of other time scales relevant to scale modelling of river channels

$$(\lambda_t)_y = (\lambda_L)^2 \tag{21}$$

$$(\lambda_t)_x = (\lambda_L)^{0.5} \tag{22}$$

$$(\lambda_t)_m = (\lambda_L)^{-1} \tag{23}$$

where $(\lambda_t)_y$ is vertical erosion/accretion; $(\lambda_t)_x$ is the downstream displacement of individual sediment grains; and $(\lambda_t)_m$ is the grain motion during saltation in either the horizontal or vertical dimensions. Vertical bed surface change is therefore the fastest time scale operating in the model relative to the prototype (see Figure 9.7), followed by sediment transport rate, the displacement of sediment or fluid in the downstream direction, particle fall velocity and the individual motion of grains during saltation:

$$(\lambda_t)_y < (\lambda_t)_s < (\lambda_t)_x\ (\lambda_t)_u < (\lambda_t)_m(\lambda_t)_{ug} \tag{24}$$

These different time scales may cause confusion when trying to interpret experimental results. For example, there has been debate as to whether short-term fluctuations in bedload rates should be scaled in terms of either total sediment transport rate, $(\lambda_L)^{1.5}$, or downstream displacement of grains, $(\lambda_L)^{0.5}$ (Ashmore 1988; Young and Davies 1991).

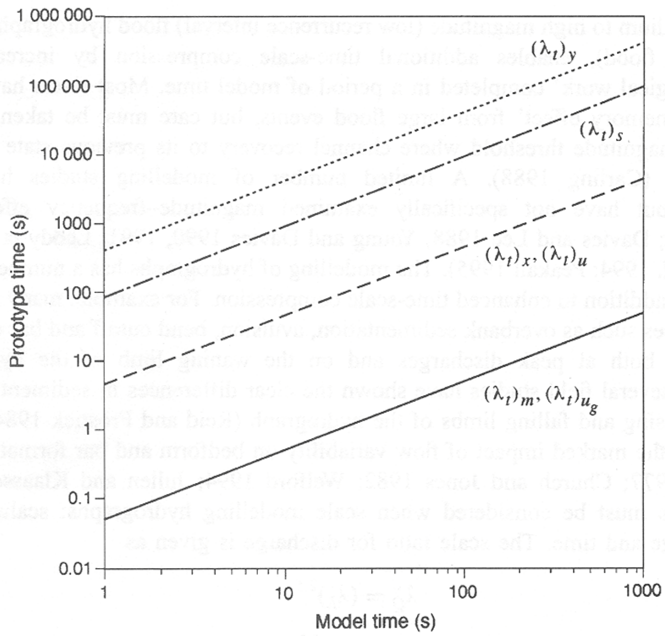

Figure 9.7 Model:prototype time scales for different processes in a 1:20 FSM $(\lambda_t)_y$, $(\lambda_t)_s$, $(\lambda_t)_u$, $(\lambda_t)_m$,and $(\lambda_t)_{ug}$, are the time scales for vertical erosion/accretion, sediment transport, downstream displacement of individual sediment grains, flow velocity, grain motion during saltation and particle fall velocity respectively (see text for more details). Some processes such as vertical erosion are much faster in the model than in the prototype whilst others, such as fall velocity and grain motion during saltation, are slower

This issue may have important ramifications for the modelling of alluvial architecture. For example, since vertical erosion and deposition are much faster than horizontal accretion, there may be distortion in the size of scours and overbank splays preserved in aggrading river models. Initial FSM work on braided river aggradation (Ashworth et al. 1994; Peakall 1995) suggests that the influence of multiple time scales is limited, but detailed experiments are still required to resolve this issue. The question of time scaling within hydraulic models and the period required for 'equilibrium' to be reached in the initial stages of experiments is clearly a subject that warrants further attention, especially if such models are to be used to investigate long-term alluvial channel behaviour.

Hydrograph scaling

A complementary approach to modelling time within physical models is through use of the geomorphological concept of event magnitude-frequency. Wolman and Miller (1960) suggested that there is a good correlation between the 'dominant' discharge (i.e. that which does most 'geomorphological work' in terms of sediment transport) and the bankfull discharge. Most physical models use a constant discharge which approximates to bankfull (e.g. Leopold and Wolman 1957; Ashmore 1982). The sequential simulation of a number of medium to high magnitude (low recurrence interval) flood hydrographs (e.g. 50 or 500 year flood), enables additional time-scale compression by increasing the 'geomorphological work' completed in a period of model time. Most rivers have a relatively short 'memory effect' from large flood events, but care must be taken to avoid exceeding a magnitude threshold where channel recovery to its previous state is imperceptibly slow (Carling 1988). A limited number of modelling studies have used hydrographs, but have not specifically examined magnitude-frequency effects (e.g. Anastasi 1984; Davies and Lee 1988; Young and Davies 1990, 1991; Leddy et al. 1993; Ashworth et al. 1994; Peakall 1995). The modelling of hydrographs has a number of other advantages in addition to enhanced time-scale compression. For example, many important fluvial processes such as overbank sedimentation, avulsion, bend cutoff and bar dissection usually occur both at peak discharges and on the waning limb of the hydrograph. Additionally, several field studies have shown the clear differences in sediment transport between the rising and falling limbs of the hydrograph (Reid and Frostick 1984; Reid et al. 1984) and the marked impact of flow variability on bedform and bar formation (Hein and Walker 1977; Church and Jones 1982; Welford 1994; Julien and Klaassen 1995).

Two aspects must be considered when scale modelling hydrographs: scaling of the water discharge and time. The scale ratio for discharge is given as

$$\lambda_Q = (\lambda_L)^{2.5} \tag{25}$$

In contrast, the time scale for the fluid is $(\lambda_L)^{0.5}$ and model hydrographs are therefore flatter than their prototype equivalents (Leddy 1993).

Aggradation and Alluvial Architecture

Most scale modelling studies have concentrated on describing the two-dimensional planform and surface geomorphology of alluvial channels. However, there is an increasing demand for extending this work into three dimensions by modelling subsidence or aggradation and therefore preserving the alluvial architecture through time. This issue is even more pressing with the recent development of three-dimensional 'process/geometry-based' alluvial architecture models (e.g. Webb 1994, 1995; Mackey and Bridge 1995) which require calibration and testing.

Logistical constraints prevent the use of most flumes for subsidence/aggradation experiments but recent work by Ashworth and Best (1994), Ashworth et al. (1994) and Peakall (1995) show that it is possible to promote aggradation and basin-wide sedimentary fill of scaled braided channels. Using a constant aggradation rate, repeated flood hydrographs and an FSM of a gravel-braided river, Ashworth and Best (1994) and Peakall (1995) showed that the preserved alluvial architecture closely resembles that seen in field outcrop and core. Figure 9.8 shows a cross-stream section through such a preserved deposit, which clearly delineates a number of key sedimentary niches of different grain size. Amalgamation of geometric and spatial information for each niche class in successive sections at closely spaced intervals, permits reconstruction of the subsurface sedimentology and, when combined with surface morphological data, the full three-dimensional alluvial architecture. Recent experiments (Ashworth and Best 1994; Peakall 1995) have started to quantify the impact of allocyclic controls (e.g. a change in

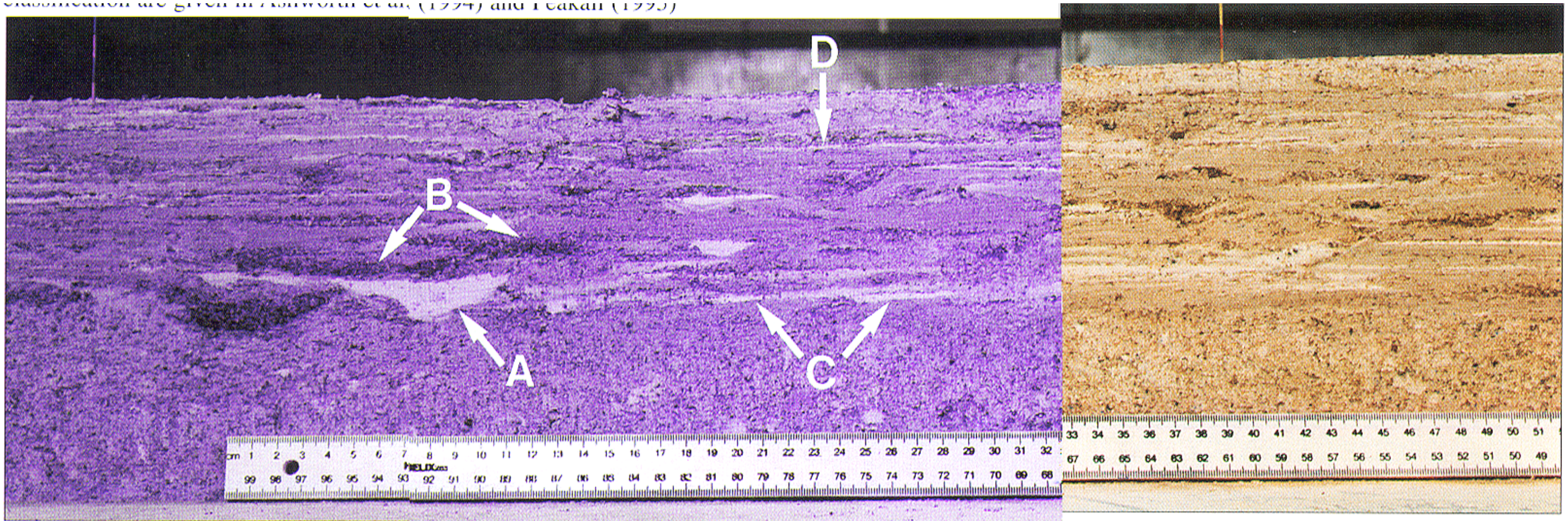

Figure 9.8 Preserved alluvial architecture from a 1:20 braided river FSM (Peakall 1995). The section is
bottom 8 cm of sediment is the original unsorted channel bed sediment. Areas of white silica flour depositio
darker coarse sand is equivalent to gravel. Note the clear differentiation of sedimentary niches with sharp ero
abandoned channel plugs (A), bar core (B) and lee deposits (C), and thin overbank and bartop splays (D).
classification are given in Ashworth et al. (1994) and Peakall (1995).

perpendicular to the mean flow direction and the
n represent fine sand/silt in the prototype whilst the
osion surfaces – common depositional units include
More details of the modelling rationale and niche

aggradation rate and the imposition of different magnitudes of lateral tectonic tilt) on fluvial deposition. Future work will consider changing hydrograph type and base level control.

Although the methodology for scale modelling of aggradation is still being developed, there is tremendous potential for answering the 'what if' scenario by systematically changing an auto or allocyclic control on alluvial deposition. The main drawback with experiments concerning aggradation is that, even with a compression of time, it is impossible to reconstruct basin sedimentation rates over geological time periods (e.g. 10^3-10^6 years). However, it may be the case that short-term, 'instantaneous' erosion/deposition events dominate the preserved alluvial record so that the gradual, long-term, basin-wide subsidence rate is less important for the preservation of *individual* depositional niches (Ashworth and Best 1994). Clearly, there is an immediate need for more experiments that employ a range of aggradation rates.

EXAMPLE OF A FROUDE SCALE MODEL

Braided gravel bed rivers are some of the most difficult environments to study in the field since the majority of planform change occurs during flood when the flow is highly turbulent and turbid, thereby limiting observation of near-bed processes. The pioneering work of Ashmore (1982, 1988, 1991a, b, 1993) was the first to highlight the potential of the FSM approach for understanding and quantifying complex and dynamic braided fluvial environments. Two aspects of this work are described here: channel confluence kinetics and the development of braiding. Both illustrate the power and potential of a Froude scale modelling approach.

Channel junctions are key nodes within braided networks and form the critical areas of flow convergence/divergence that are instrumental in the process of braiding. Past scaled models using both fixed (Mosley 1976; Best 1987, 1988) and mobile banks (Mosley 1976), have considered the details of the confluence zone in terms of the flow dynamics and sediment transport pathways. Additionally, fieldwork has provided invaluable insights into the processes operative at these sites (e.g. Roy and Roy 1988; Ashmore et al. 1992; Biron et al. 1993). However, the importance of channel junctions at larger spatial and temporal scales could not be addressed by these studies. Ashmore (1993), however, considers the kinetics of channel junctions in an FSM and presents models for the migration of channel junctions and their influence on downstream sedimentation. An example from this work demonstrates destruction of a post-confluence medial bar through downstream migration of the upstream junction (Figure 9.9). Such qualitative and semi-quantitative studies may be used to propose generalised models of confluence zone migration (Figure 9.10).

Ashmore's work on confluence dynamics links to the broader issue of bar and channel pattern development. In a sequence of papers, Ashmore (1982, 1991b, 1993) has successfully used an FSM to classify the main mechanisms of braiding, explain the processes controlling bar formation and down-bar fining, and relate the internal generation of bedload pulses to channel change and bar dissection/creation. One of the most influential papers (Ashmore 1991b) unambiguously defined the causes of braiding both at the initial stage of a single channel (e.g. Ashmore 1991b, Figure 3, p. 330) and when a fully braided

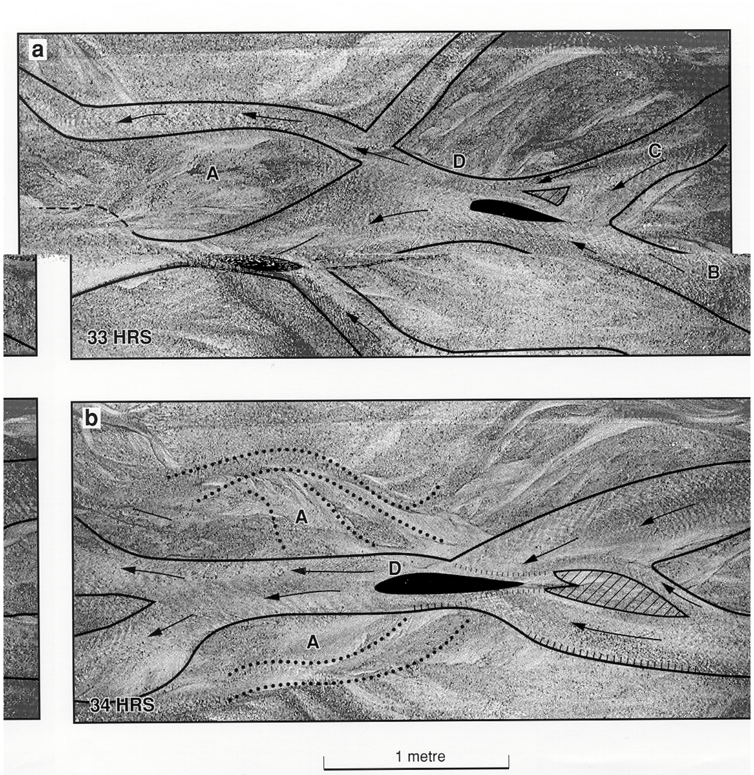

Figure 9.9 Medial bar destruction caused by longitudinal translation and change in total discharge of an upstream confluence. (a) The medial bar complex (A) formed by rapid progradation of sediment in the left confluent channel (B) and subsequent channel bifurcation. Waning of the flow in the right confluent channel (C) led to the formation of a new confluence (D) further downstream. (b) Later, lateral migration of the confluent channels caused the confluence zone (D) to migrate downstream, eventually triggering an avulsion across the centre of the medial bar, leaving two isolated remnants (A). Figure and interpretation reproduced from Ashmore (1993) with permission of the author and the Geological Society of London

network has developed (see Figure 9.11). The main braiding mechanisms identified were through deposition of a central bar, chute cut-off of point bars, conversion of a single transverse unit bar to a mid-channel bar and dissection of multiple bars. The chute cut-off mechanism was the most common process of braiding in Ashmore's experiments (Figure 9.11). This process may be very common in single low-sinuosity gravel-bed streams (cf. Lewin 1976; Carson 1986) and is the dominant transformation process from an initial single channel with alternate bars to a braided network that occurs at the beginning of most scale modelling experiments of braided rivers (Leddy 1993). By relating the different mechanisms of braiding to the local flow conditions (excess shear stress), channel cross-sectional geometry and bedform regime, Ashmore's (1991b) work provided valuable insights into the critical conditions necessary for channel bifurcation and mid-channel bar growth. Ashmore's experiments used a constant water discharge, truncated grain-size

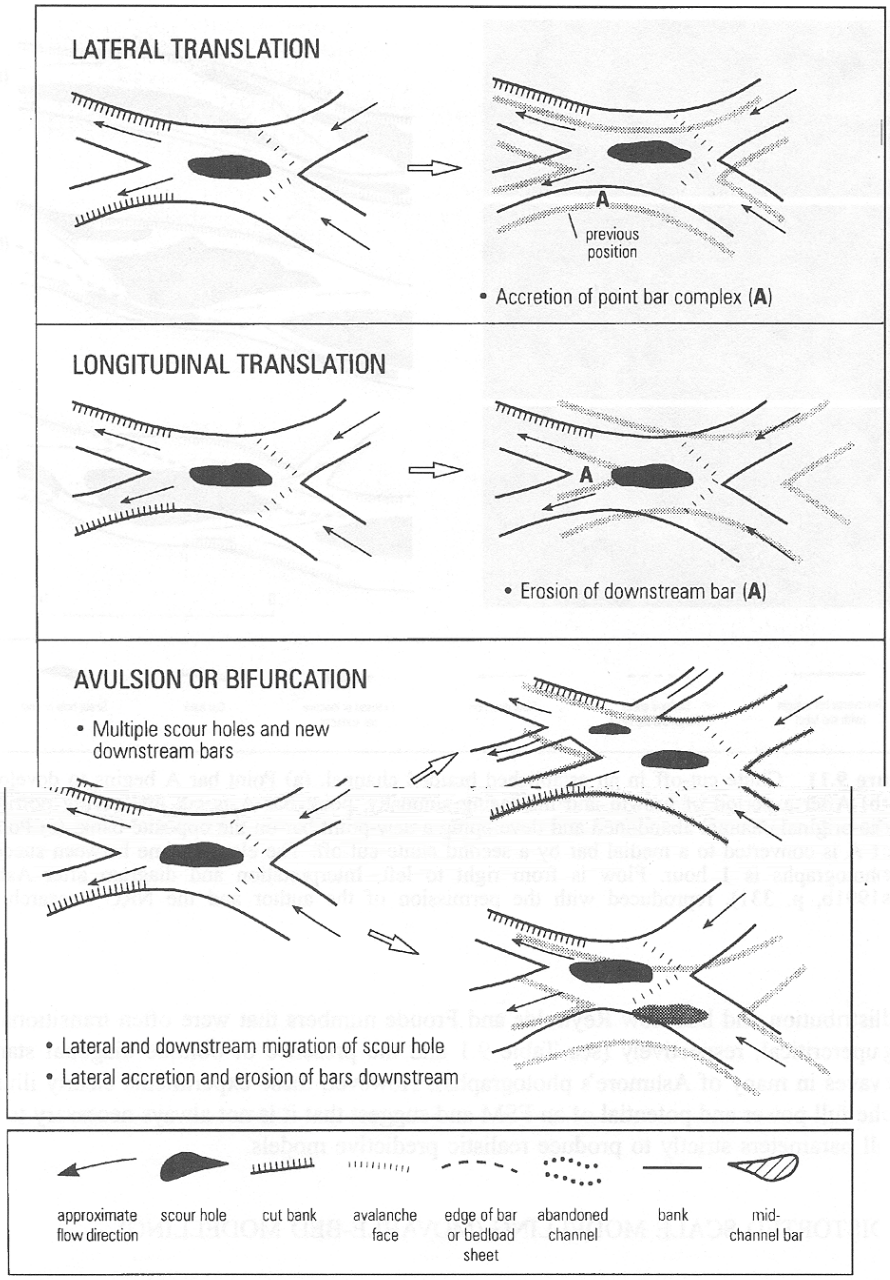

Figure 9.10 Schematic summary of observed modes of confluence movement and sedimentation in response to the migration of confluent anabranches. After Ashmore (1993) and reproduced by permission of the author and the Geological Society of London

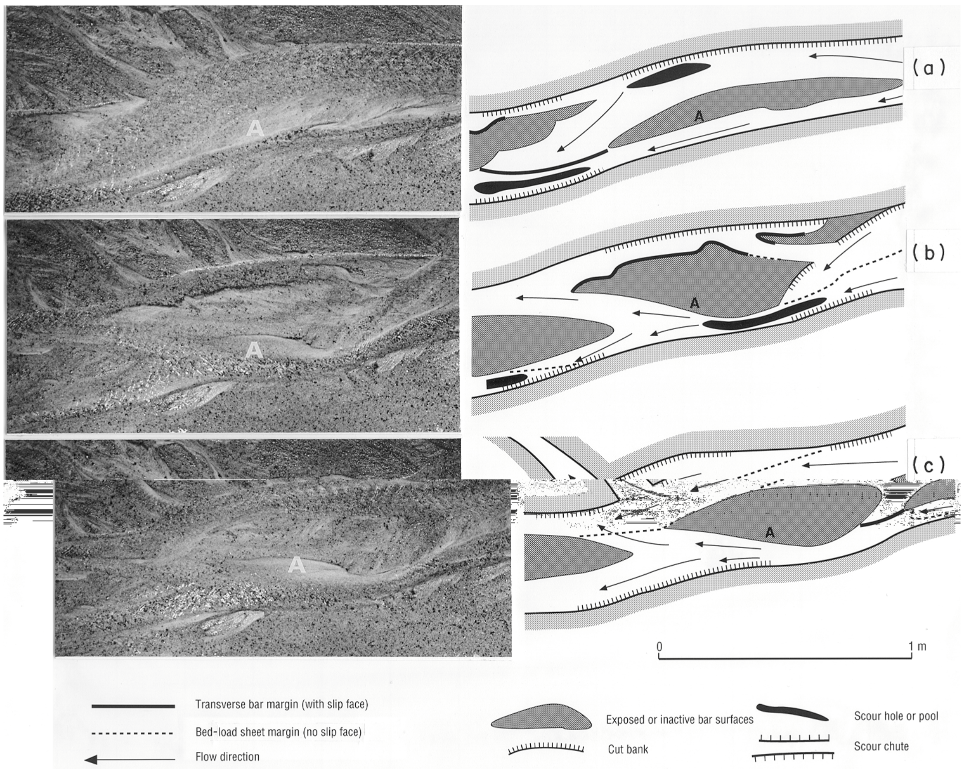

Figure 9.11 Chute cut-off in an established braided channel. (a) Point bar A begins to develop. (b) After a period of growth and increasing sinuosity, point bar A is cut off leaving portions of the original channel abandoned and developing a new point bar on the opposite bank. (c) Point bar at A is converted to a medial bar by a second chute cut-off. The elapsed time between successive photographs is 1 hour. Flow is from right to left. Interpretation and diagram after Ashmore (1991b, p. 331), reproduced with the permission of the author and the NRC Research Press

distribution and had flow Reynolds and Froude numbers that were often transitional and supercritical, respectively (see Table 9.1 and the presence of oblique diagonal standing waves in many of Ashmore's photographs). However, these experiments clearly illustrate the full power and potential of an FSM and suggest that it is not always necessary to scale all parameters strictly to produce realistic predictive models.

DISTORTED SCALE MODELLING (MOVABLE-BED MODELLING)

For the study of large geomorphological scales, fine prototype sediments or precise modelling of sediment transport and deposition, models may have to be distorted. Geometrically distorted models have a small vertical:horizontal scale ratio in order to model large prototypes, whilst maintaining adequate model flow depth. Perhaps the most spectacular example of a distorted geometric model is that of the entire Mississippi basin by the US Corps of Engineers at a horizontal scale of 1:2000 and vertical scale of 1:100, on a 100 ha site (reported in Novak and Cábelka 1981, p. 163). Distortion of the geometric scale is also a prerequisite in many cases where the precise modelling of sediment movement is attempted since this increases model shear stresses. A number of other scales are also commonly adjusted when modelling sediment movement in rivers, including valley slope, grain size, sediment density, flow velocity and discharge. The combination of these scale adjustments is usually referred to as 'movable-bed modelling' in the engineering literature and refers to alteration of more than just the boundary conditions.

One example of movable-bed modelling is McCollum's (1988) study of the Apalachicola River in the southern United States. The model reproduced a 7 km section of the Apalachicola River which suffered from persistent sedimentation, requiring annual dredging to maintain a navigation channel. Horizontal and vertical scales were 1:120 and 1:80 respectively and crushed coal with a specific gravity of 1.3 g cm^{-3} was used to simulate the sandy bed of the Apalachicola, thereby avoiding introduction of cohesion into the model. Model verification consisted of replicating a 12-month prototype discharge record and comparing the observed model changes with hydrographic surveys. The channel slope and discharge ratio were also adjusted to achieve a good model:prototype agreement. After initial verification, a number of channel improvement schemes were tested using a wide range of discharges. Based on these tests, a series of 'L'-shaped dikes were proposed as the most effective method for maintaining the navigation channel.

McCollum's (1988) study illustrates the potential of movable-bed modelling for studying large prototypes, fine-sediment and specific sediment transport problems. There appears to be great potential for utilising the same techniques within fluvial geomorphology, as illustrated by Klaassen (1991) for the Brahmaputra River (Bangladesh) and Davies and Griffiths (in press) for flow-sediment transport relationships in the Waimakariri River (New Zealand).

ANALOGUE MODELLING

Analogue models have been used to study a wide range of fluvial scales from small channels to entire drainage networks. Although Schumm et al. (1987) illustrate the many conceptual and practical advantages this approach offers for geomorphology, some of the small geomorphological scales investigated in the past using analogue models can now be modelled in far more detail using FSM and movable-bed techniques (see Figure 9.1 and earlier discussion). The main advantages of analogue models are speed and simplicity in setting up experiments and the reduced space and budget costs which constrain other modelling approaches. These advantages are illustrated using recent studies of base level change and alluvial fan aggradation.

Base Level Control and Sequence Stratigraphy

The continental shelf/slope system is of such a large size that a scale model would be prohibitively expensive, if not impossible, to construct. In addition, important attributes such as cohesion and vegetation are difficult to incorporate into such models. However, the effect of base level change on these systems has recently been studied using analogue models of- (i) a fan forming in a drainage ditch (Posamentier et al. 1992), (ii) a single channel in a stream table (Wood et al. 1993) and (iii) a drainage network developed in a stream table using a rainfall simulator (Koss et al. 1994). These analogue models have been used to test sequence stratigraphic concepts that are difficult to examine using conventional computer modelling techniques. The analogue models illustrate several important points:

- Depositional systems 'tracts' and the bounding surfaces between them are scale independent (Posamentier et al. 1992; Koss et al. 1994).
- There is a significant lag time between base level fall and coarse-grained sediment reaching the lowstand fan (Wood et al. 1993; Koss et al. 1994).
- A large number of incised valleys form at the shelf/slope interface, only one of which connects to the main river system (Wood et al. 1993; Koss et al. 1994).
- Base level rise is frequently accompanied by significant slumping on the walls of incised valleys (Wood et al. 1993).
- The rate of base level fall and the shelf angle are important controls on sediment deposition and preservation (Wood et al. 1993, 1994).

Whilst these results cannot be directly quantified, they demonstrate one of the major advantages of analogue models in that they can be used to test some of the latest hypotheses concerning large-scale dynamics of sedimentary basins which are extremely difficult to verify by any other technique.

Alluvial Fan Aggradation

Analogue modelling can provide extremely valuable information on general alluvial fan and channel dynamics as shown by Schumm et al. (1987), Bryant et al. (1995) and Whipple et al. (1995). Bryant et al. (1995) tested the hypothesis that the frequency of channel avulsions is linked to the sedimentation rate by forming a simple alluvial fan (sediment cone) and varying the sediment input through time. Whilst the alluvial fan cannot be directly compared with field examples, the experiments clearly demonstrate the general principle that avulsion frequency is directly related to sedimentation rate at all but the highest rates. This conclusion has significant implications for models of alluvial architecture, many of which assume that the avulsion frequency is invariant (e.g. Bridge and Leeder 1979; Mackey and Bridge 1992). More recently, analogue modelling of alluvial fan dynamics has been used in an applied context to simulate the redistribution and aggradation of mine waste tailings from a single point source (Parker 1996).

ADVANCES AND TRENDS IN PHYSICAL MODELLING

The application of physical modelling within fluvial geomorphology has produced major advances in the last two decades, many of which have resulted from, or been produced by, improvements in the methodology by which experiments are conducted. Apart from the solution of scaling issues, future advances in scale modelling will undoubtedly be accompanied by the development of appropriate measurement technology for these models. A clear parallel to this situation is shown by work over the past 30 years concerning bedform generation in scaled hydraulic flume models. This work has progressed from the qualitative description of bedform morphology and their gross fluid dynamic controls (e.g. Simons et al. 1961; Guy et al. 1966), to quantification of sediment transport rates and investigation of bedform response to changing flow conditions (Guy et al. 1966; see Allen 1982) through to studies of the detailed flow, turbulence and sediment dynamics associated with a range of bedforms (e.g. Raudkivi 1966; McLean et al. 1994; Bennett and Best 1995; Nelson et al. 1995). Increases in the quantitative assessment of the processes governing bedform stability have been brought about by changes in instrumentation during this period, which have been partly responsible for the increasing levels of resolution incorporated within numerical models of bedform generation. Perhaps the greatest potential for increasing both the applicability of scale models to larger scales (of the order of the channel width or braidplain scale) and confidence in the degree of prototype agreement, lies in developing new methods of flume experimentation. Several areas appear ripe for development at present.

Quantification of Flow

Most past FSM studies have recorded only the basic attributes of water flow through model fluvial channels. These velocity measurements have usually been obtained by surface float tracing (e.g. Ashworth 1996) or use of pitot tubes (Ashmore 1982, 1996). However, these measurements are often very difficult, if not impossible, to perform in shallow flow depths, or across the width of the channel, and therefore many model studies present flow/hydraulic data predominantly from large channels. Instrumentation in these model channels is extremely difficult yet clearly central to verifying the range of flow conditions within an FSM. The solution to these problems may lie in the application of several available methods of instrumentation for quantifying flow structure:

1. Hot-film or laser Doppler anemometry (e.g. Durst et al. 1987; Tritton 1988) where flow depths and experimental configuration permit. Although hot-film probes may possess calibration difficulties, especially in flows with sediment transport, the sensors often are small enough to be used in shallow depths. Laser Doppler anemometry may provide a non-intrusive methodology for recording flow velocities/turbulence by focusing the beams through the water surface, but will only work in clear liquids with little sediment transport.
2. Ultrasonic Doppler anemometry (e.g. Takeda 1991, 1995) offers the potential for obtaining high-resolution, often multi-point, measurements in opaque fluids although current systems may only be of use in channels with flow depths of the order of several centimetres or greater.
3. Particle tracking and particle image velocimetry (PIV; Linden et al. 1995; Seal et al. 1995) offers great potential to quantify particle velocities (of perhaps fluid and sediment) within the flow and may provide the best method for yielding channel-wide estimates of flow velocity both on the water surface and within the flow through use of neutrally buoyant particles.

Quantification of Topography

Many FSM studies have quantified channel change through continuous recording using video cameras in conjunction with limited topographic surveying/point gauging. One area which would immediately yield valuable information on scale model topography would be the rapid, and automated, quantification of bed heights within the scale model.

The appropriate methodology to accomplish this goal may consist of the application of ultrasonic bed profilers; (e.g. Kuhnle 1993; Best and Ashworth 1994) which can resolve heights down to 0.1 mm. The development of photogrammetric methods (Ashmore, personal communication 1995) or use of laser light sheets (Rice et al. 1988; Römkens et al. 1988) may also yield suitable technology.

Quantification of Sediment Transport

Several studies have documented sediment transport rates within model studies and used these to discuss phenomena such as the presence and importance of bedload pulses within braided channels (e.g. Ashmore 1988). However, prediction of more local channel change (avulsion for example) and development of braid networks requires more detailed quantification of the rates of sediment transport *within* individual model channels. Few studies have sought to address this topic and the introduction of samplers into these flows is fraught with difficulties, not just in the disturbance to the flow field, but in the design and efficiency of the sediment samplers themselves. Apart from use of high-resolution and continuous ultrasonic bed profilers, which may be used to quantify bed height change, the use of PIV techniques to track different size (i.e. colour) grains may yield valuable estimates of transport rates and pathways, possibly of individual size fractions within the sediment load. Refinement of acoustic devices which have been developed to monitor bedload noise (e.g. Thorne et al. 1989; Hardisty 1993; Rouse 1994) could yield another methodology for estimating transport rates, whilst individual particle trajectories in clear flows may be tracked using high-speed video (1000 frames per second, cf. Garcia et al. 1996). Other tracer techniques, perhaps based on thermal imaging of grain paths within the flow, may provide a tool for providing the much-needed quantification of the links between flow, sediment transport and channel change.

Quantification of Sedimentary Architecture

If FSMs can be used to examine the subsurface geometry and internal bedding characteristics of fluvial deposits, great potential exists for obtaining true three-dimensional descriptions of such deposits. Apart from detailed trenching and description of these sediments, use of miniaturised geophysical techniques, such as seismic imaging or resistivity techniques, may provide invaluable tools for quantifying subsurface sedimentary structure and connectivity between key depositional elements.

SUMMARY

The use and application of physical modelling within fluvial geomorphology lies at a crossroad. Work over the past 20 years has yielded considerable advances in our knowledge of many complex fluvial processes and forms and has progressed to incorporate realistic scaling assumptions from the scale of the sediment grain to that of the river channel. Scale models are now powerful tools for testing mathematical models because they can closely approximate the idealised assumptions that underpin many numerical models. Further progress in this field may only be possible if three issues are addressed:

1. Incorporation of more realistic model parameters into FSM studies, such as flood hydrographs, fine-grained sediment and cohesion;
2. Development and implementation of methodology to better quantify flow, sediment transport and morphological change;
3. Additional and more complete testing/verification of FSMs against their prototype conditions.

Although unscaled or 'analogue' models can shed much light on large temporal and spatial scale processes and products, these studies must always be considered in terms of their underlying simplifications and drawbacks and must not be interpreted as true scale models. Such 'analogue' models may increasingly be used to investigate the role of allocyclic factors on sedimentation, such as local tectonic and base level controls, but their departure from true scaling perhaps demands more complete field verification than more rigidly scaled Froude scale models.

ACKNOWLEDGEMENTS

Many of the ideas expressed in this chapter have developed from work sponsored over the past six years by BP Exploration. We are grateful to BP for award of a Ph.D. studentship to Jeff Peakall and grants to establish the scale modelling/aggradation facility at Leeds. This modelling has also been supported by a grant from the Royal Society and more recently funding from NERC (GR9/01640) and ARCO Oil (USA) to Ashworth and Best. Marcelo Garcia and Bruce Rhoads provided helpful suggestions to improve the clarity of this contribution. Peter Ashmore kindly supplied original photographs of his flume experiments for Figures 9.9 and 9.11.

REFERENCES

Allen, J.R.L. 1982. *Sedimentary Structures: Their Character and Physical Basis,* Elsevier, Amsterdam, 539 pp..

Anastasi, G. 1984. Simulazinoe di regime torrentizio su modello fisico a fondo mobile mediante micro-computer, in *Memorie XIX convegno di idraulica e construczioni idrauliche,* Pavia, Italy, 6-8 September 1984, Paper A9, 10 pp.

ASCE 1942. *Hydraulic Models, The American Society of Civil Engineers Manual of Practice,* **25,** American Society of Civil Engineers, New York, 110 pp.

Ashmore, P.E. 1982. Laboratory modelling of gravel braided stream morphology, *Earth Surface Processes and Landforms,* **7***,* 201-225.

Ashmore, P.E. 1988. Bedload transport in braided gravel-bed stream models, *Earth Surface Processes and Landforms,* **13***,* 677-695.

Ashmore, P.E. 1991a. Channel morphology and bed load pulses in braided, gravel-bed streams, *Geografiska Annaler,* **68***,* 361-371.

Ashmore, RE. 1991b. How do gravel-bed rivers braid? *Canadian Journal of Earth Sciences,* **28***,* 326-341.

Ashmore, RE. 1993. Anabranch confluence kinetics and sedimentation processes in gravel-braided streams, in *Braided Rivers,* edited by J.L. Best and C.S. Bristow, Geological Society Special Publications, 75, pp. 129-146.

Ashmore, RE., Ferguson, R.I., Prestegaard, K.L., Ashworth, P.J. and Paola, C. 1992. Secondary flow in anabranch confluences of a braided, gravel-bed stream, *Earth Surface Processes and Landforms,* **17***,* 299-311.

Ashmore, RE. and Parker, G. 1983. Confluence scour in coarse braided streams, *Water Resources Research,* **19***,* 392-402.
Ashworth, P.J. 1996. Mid-channel bar growth and its relationship to local flow strength and direction, *Earth Surface Processes and Landforms,* **21***,* 103-123.
Ashworth, P.J. and Best, J.L. 1994. The scale modelling of braided rivers of the Ivishak Formation, Prudhoe Bay 11: shale geometries and response to differential aggradation rates, *Final BP Project Report,* August 1994, 247 pp.
Ashworth, P.J. and Best, J.L. in press. Discussion of 'the use of hydraulic models in the management of braided gravel-bed rivers' by Warburton, J. and Davies, T.R.H., in *Gravel-bed Rivers in the Environment,* edited by PC. Klingeman, R.L. Beshta, P.D. Komar and J.B. Bradley, Wiley, New York.
Ashworth, P.J., Best, J.L. and Leddy, J.O. 1992b. The scale modelling of braided rivers of the Ivishak Formation, Prudhoe Bay, *Final BP Project Report Phase 2,* September 1992, 76 pp.
Ashworth, P.J., Best, J.L., Leddy, J.O. and Geehan, G.W. 1994. The physical modelling of braided rivers and deposition of fine-grained sediment, in *Process Models and Theoretical Geomorphology,* edited by M.J. Kirkby, Wiley, Chichester, pp. 115-139.
Ashworth, P.J., Ferguson, R.I. and Powell, M.D. 1992a. Bedload transport and sorting in braided channels, in *Dynamics of Gravel-bed Rivers,* edited by P. Billi, R.D. Hey, C.R. Thorne and P Tacconi, Wiley, Chichester, pp. 497-513.
Bagnold, R.A. 1980. An empirical correlation of bedload transport rates in natural rivers, *Proceedings of the Royal Society of London,* **372A,** 453-473.
Bennett, S.J. and Best, J.L. 1995. Mean flow and turbulence structure over fixed, two-dimensional dunes: implications for sediment transport and bedform stability, *Sedimentology,* **42***,* 491-513.
Best, J.L. 1987. Flow dynamics at river channel confluences: implications for sediment transport and bed morphology, in *Recent Developments in Fluvial Sedimentology,* edited by F.G. Ethridge, R.M. Flores and M.D. Harvey, Special Publication of the Society of Economic Palaeontologists and Mineralologists 39, pp. 27-35.
Best, J.L. 1988. Sediment transport and bed morphology at river channel confluences, *Sedimentology,* **35***,* 481-498.
Best, J.L. 1996. The fluid dynamics of small-scale alluvial bedforms, in *Advances in Fluvial Dynamics and Stratigraphy,* edited by P.A. Carling and M. Dawson, Wiley, Chichester, 67-125.
Best, J.L. and Ashworth, P.J. 1994. A high-resolution ultrasonic bed profiler for use in laboratory flumes, *Journal of Sedimentary Research,* **A64***,* 674-675.
Biron, P., de Serres, B., Roy, A.G. and Best, J.L 1993. Shear layer turbulence at an unequal depth channel confluence, in *Turbulence: Perspectives on Flow and Sediment Transport,* edited by N.J. Clifford, J.R. French and J. Hardisty, Wiley, Chichester, pp. 197-213.
Boothroyd, J.C. and Ashley, G.M. 1975. Process, bar morphology and sedimentary structures on braided outwash fans, northeastern Gulf of Alaska, in *Glaciofluvial and Glaciolacustrine Sedimentation,* edited by A.V. Jopling and B.C. McDonald, Society of Economic Paleontologists and Mineralogists Special Publication 23, pp. 193-222.
Bridge, J.S. and Leeder, M.R. 1979. A simulation model of alluvial stratigraphy, *Sedimentology,* **26***,* 617-644.
Bristow, C.S. and Best, J.L. 1993. Braided rivers: perspectives and problems, in *Braided Rivers,* edited by J.L. Best and C.S. Bristow, Geological Society Special Publication 75, pp. 1-11.
Bryant, M., Falk, P. and Paola, C. 1995. Experimental study of avulsion frequency and rate of deposition, *Geology,* **23***,* 365-368.
Buckingham, E. 1915. Model experiments and the forms of empirical equations, *Transactions of the American Society of Mechanical Engineers,* **37***,* 263-292.
Carling, P 1988. The concept of dominant discharge applied to two gravel-bed streams in relation to channel stability thresholds, *Earth Surface Processes and Landforms,* **13***,* 355-367.
Carson, M.A. 1986. Characteristics of high-energy 'meandering' rivers: the Canterbury Plains, New Zealand, *Geological Society of America Bulletin,* **97***,* 886-895.
Chadwick, A.J. and Morfett, J.C. 1986. *Hydraulics in Civil Engineering,* Harper Collins, London, 492 pp.

Chorley, R.J. 1967. Models in geomorphology, in *Models in Geography,* edited by R.J. Chorley and P. Haggett, Methuen, London, pp. 59-96.
Church, M. and Jones, D. 1982. Channel bars in gravel-bed rivers, in *Gravel-bed Rivers,* edited by R.D. Hey, J.C. Bathurst and C.R. Thorne, Wiley, Chichester, pp. 291-338.
Church, M., Wolcott, J.F. and Fletcher, W.K. 1991. A test of equal mobility in fluvial sediment transport: behaviour of the sand fraction, *Water Resources Research,* **27**, 2941-295 1.
Crickmore, M.J. 1970. Effect of flume width on bedform characteristics, *Journal of the Hydraulics Division, Proceedings of the American Society of Civil Engineers,* **96**, 473-496.
Davies, T.R.H. and Griffiths, G.A. in press. Physical model study of stage-discharge relationships in a braided river gorge, *Journal of Hydrology (New Zealand).* **35**, 2.
Davies, T.R.H. and Lee, A.L. 1988. Physical hydraulic modelling of width reduction and bed level change in braided rivers, *Journal of Hydrology (New Zealand),* **27**, 113-127.
Dingman, S.L. 1984. *Fluvial Hydrology, WH.* Freeman, New York, 383 pp.
Durst, F., Melling, A. and Whitelaw, J.H. 1987. *Principles and Practice of Laser-Doppler Anemometry,* 2nd edn, G. Braun, Karlsruhe, 405 pp.
Elghobashi, S. 1994. On predicting particle-laden turbulent flows, *Applied Scientific Research,* **52**, 309-329.
Fahnestock, R.K. 1963. Morphology and hydrology of a glacial stream - White River, Mount Rainier, Washington, *US Geological Survey Professional Paper 422A,* 70 pp.
Franco, J.J. 1978. Guidelines for the design, adjustment and operation of models of the study of river sedimentation problems, Instruction Report H-78-1, *US Waterways Experimental Station,* Vicksburg, Miss., 57 pp.
French, R.H. 1985. *Open-channel Hydraulics,* McGraw-Hill, New York, 739 pp.
Garcia, M.H. 1993. Hydraulic jumps in sediment-driven bottom currents, *Journal of Hydraulic Engineering,* **119**, 1094-1117.
Garcia, M.H., Nino, Y. and Lopez, F. 1996. Laboratory observations of particle entrainment into suspension by turbulent bursting, in *Coherent Flow Structures in Open Channels,* edited by P.J. Ashworth, S.J. Bennett, J.L. Best and S.J. McLelland, Wiley, Chichester, 63-68.
Gibbs, R.J. Mathews, M.D. and Link, D.A. 1971. The relationship between sphere size and settling velocity, *Journal of Sedimentary Petrology,* **41**, 7-18.
Gilbert, G.K. 1914. Transportation of debris by running water, *US Geological Survey Professional Paper 86,* 263 pp.
Gilbert, G.K. 1917. Hydraulic mining debris in the Sierra Nevada, *US Geological Survey Professional Paper 105,* 154 pp.
Guy, H.P., Simons, D.B. and Richardson, E.V.1966. Summary of alluvial channel data from flume experiments, 1956-1961, *US Geological Survey Professional Paper 462-1,* 96 pp.
Hardisty, J. 1993. Monitoring and modelling sediment transport at turbulent frequencies, in *Turbulence: Perspectives on Flow and Sediment Transport,* edited by N.J. Clifford, J.R. French and J. Hardisty, Wiley, Chichester, pp. 35-59.
Hein, F.J. and Walker, R.G. 1977. Bar evolution and development of stratification in the gravelly, braided Kicking Horse River, British Columbia, *Canadian Journal of Earth Sciences,* **14**, 562-570.
Henderson, F.M. 1966. *Open Channel Flow,* Macmillan, New York, 552 pp.
Hoey, T.B. and Sutherland, A.J. 1989. Self formed channels in a laboratory sand tray, *Proceedings, 23rd Congress, International Association for Hydraulic Research,* Ottawa, Canada, pp. 41-48.
Hoey, T.B. and Sutherland, A.J. 1991. Channel morphology and bedload pulses in braided rivers: a laboratory study, *Earth Surface Processes and Landforms,* **16**, 447-462.
Hooke, R.L. 1968. Model geology: prototype and laboratory streams: discussion, *Geological Society of America Bulletin,* **79**, 391-394.
Ikeda, S. and Kanazawa, M. 1995. Organised vortex structures in turbulent flows with flexible water plants, in *Coherent Flow Structures in Open Channels,* Leeds, England, 10-12th April 1995, *Abstract Volume,* 29 pp.
Ikeda, S. and Parker G. (eds). 1989. *River Meandering,* American Geophysical Union, Water Resources Monograph 12, 485 pp.

Jaeggi, M.N.R. 1986. Non distorted models for research on river morphology, *Proceedings of the Symposium on Scale Effects in Modelling Sediment Transport Phenomena,* August 1986, International Association of Hydrological Sciences, pp. 70-84.

Jin, D. and Schumm, S.A. 1987. A new technique for modelling river morphology, in *International Geomorphology, 1986 Part 1,* edited by V. Gardiner, Wiley, Chichester, pp. 681-690.

Julien, P.Y. and Klaassen, G.J. 1995. Sand-dune geometry of large rivers during floods, *Journal of Hydraulic Engineering,* **121**, 657-663.

Karcz, I. and Kersey, D. 1980. Experimental study of free-surface flow instability and bedforms in shallow flows, *Sedimentary Geology,* **27**, 263-300.

Kennedy, J.F. 1995. The Albert Shields story, *Journal of Hydraulic Engineering,* **121**, 766-772.

Kirkby, M.J. (ed.). 1994. *Process Models and Theoretical Geomorphology,* Wiley, Chichester, 417 pp.

Klaassen, G.J. 1991. On the scaling of braided sand-bed rivers, in *Movable Bed Physical Models,* edited by H.W. Shen, Kluwer Academic, New York, pp. 59-72.

Koss, J.E., Ethridge, F.G. and Schumm, S.A. 1994. An experimental study of the effects of base-level change on fluvial, coastal plain and shelf systems, *Journal of Sedimentary Research,* **B64**, 90-98.

Kuhnle, R.A. 1993. Incipient motion of sand-gravel sediment mixtures, *Journal of Hydraulic Engineering,* **119**, 1400-1415.

Kuhnle, R.A. and Southard, J.B. 1988. Bed load transport fluctuations in a gravel bed laboratory channel, *Water Resources Research,* **24**, 247-260.

Langhaar, H.L. 1980. *Dimensional Analysis and Theory of Hydraulic Models,* Robert E. Krieger, Florida, 178 pp.

Leddy, J.O. 1993. Physical scale modelling of braided rivers: avulsion and channel pattern change, M.Phil.thesis, University of Leeds, 130 pp.

Leddy, J.O., Ashworth, P.J. and Best, J.L. 1993. Mechanisms of anabranch avulsion within gravel-bed braided rivers: observations from a scaled physical model, in *Braided Rivers,* edited by J.L. Best and C.S. Bristow, Geological Society Special Publication, 75, pp. 119-127.

Leeder, MR. 1982. *Sedimentology: Process and Product,* Unwin Hyman, London, 344 pp.

Leopold, L.B. and Wolman, M.G. 1957. River channel patterns: braided, meandering and straight, *US Geological Survey Professional Paper 282-B, pp.* 39-85.

Lewin, J. 1976. Initiation of bed forms and meanders in coarse-grained sediment, *Geological Society of America Bulletin,* **87**, 281-285.

Linden, R.E., Boubnov, B.M. and Dalziel, S.B. 1995. Source-sink turbulence in a rotating stratified fluid, *Journal of Fluid Mechanics,* **298**, 81-112.

Lisle, T.E., Ikeda, H. and Iseya, F. 1991. Formation of stationary alternate bars in a steep channel with mixed-size sediment: a flume experiment, *Earth Surface Processes and Landforms,* **16**, 463-469.

McCollum, R.A. 1988. Blountstown Reach, Apalachicola River; movable-bed model study, Technical Report HL-88-17, *US Waterways Experimental Station,* Vicksburg, 39 pp.

Mackey, S.D. and Bridge, J.S. 1992. A revised FORTRAN program to simulate alluvial stratigraphy, *Computers and Geosciences,* **18**, 119-181.

Mackey, S.D. and Bridge, J.S. 1995. Three-dimensional model of alluvial stratigraphy: theory and application, *Journal of Sedimentary Research,* **B65**, 7-31.

McLean, S.R., Nelson, J.M. and Wolfe, S.R. 1994. Turbulence structure over two-dimensional bedforms: implications for sediment transport, *Journal of Geophysical Research,* **99**, 12 729-12 747.

Marsden, N. 1981. Model simulation of effect of vegetation on braided rivers, *Project Report,* Department of Agricultural Engineering, Lincoln College, Canterbury, New Zealand, 68 pp.

Mosley, M.R 1976. An experimental study of channel confluences, *Journal of Geology,* **84**, 535-562.

Mosley, M.P. and Zimpfer, G.L. 1978. Hardware models in geomorphology, *Progress in Physical Geography,* **2**, 438-461.

Murphy, G. 1950. *Similitude in Engineering,* Ronald Press, New York, 302 pp.

Nelson, J.M., Shreve, R.L., McLean, S.R. and Drake, T.G. 1995. Role of near-bed turbulence structure in bed load transport and bed form mechanics, *Water Resources Research,* **31**, 2071-2086.

Nikuradse, J. 1933. Strömungsgesetze in rauhen Rohren, VDI-Forschungsheft, 361. English translations: *NACA Technical Memo. 1292* and Petroleum Engineer (1940) March, 164-166; May, 75, 78, 80, 82; June, 124, 127, 128, 130; July, 38, 40, 42; August, 83, 84 and 87.

Novak, P. and Cábelka, J. 1981. *Models in Hydraulic Engineering,* Pitman, London, 459 pp.

Ouchi, S. 1985. Response of alluvial rivers to slow active tectonic movement, *Geological Society of America Bulletin,* **96**, 504-515.

Owen, M.W. 1985. Ports and harbours, in *Developments in Hydraulic Engineering - 3,* edited by P. Novak, Elsevier, London, pp. 263-311.

Paola, C., Parker, G., Seal, R., Sinha, S.K., Southard, J.B. and Wilcock, P.R. 1992. Downstream fining by selective deposition in a laboratory flume, *Science,* **258**, 1757-1760.

Parker, G. 1979. Hydraulic geometry of active gravel rivers, *Journal of the Hydraulics Division, Proceedings of the American Society of Civil Engineers,* **105**, 1185-1201.

Parker, G. 1996. Some speculations on the relation between channel morphology and channel-scale flow structures, in *Coherent Flow Structures in Open Channels,* edited by P.J. Ashworth, S.J. Bennett, J.L. Best and S.J. McLelland Wiley, Chichester, pp. 423-458.

Parker, G., Garcia, M., Fukushima, Y. and Yu, W. 1987. Experiments on turbidity currents over an erodible bed, *Journal of Hydraulic Research,* **25**, 123-147.

Peakall, J. 1995. The influences of lateral ground-tilting on channel morphology and alluvial architecture, Ph.D. thesis, University of Leeds, 333 pp.

Peakall, J. and Warburton, J. in press. Surface tension in small hydraulic river models - the significance of the Weber number, *Journal of Hydrology (New Zealand),* **35**, 2.

Pickup, G. 1988. Hydrology and sediment models, in *Modelling Geomorphological Systems,* edited by M.G. Anderson, Wiley, Chichester, pp. 153-215.

Posamentier, H.W., Allen, G.P. and James, D.P. 1992. High resolution sequence stratigraphy - the East Coulee Delta, Alberta, *Journal of Sedimentary Petrology,* **62**, 310-317.

Raudkivi, A.J. 1966. Bed forms in alluvial channels, *Journal of Fluid Mechanics,* **26**, 507-514.

Reid, I., Brayshaw, A.C. and Frostick, L.E. 1984. An electromagnetic device for automatic detection of bedload motion and its field applications, *Sedimentology,* **31**, 269-276.

Reid, 1. and Frostick, L.E. 1984. Particle interaction and its effect on the thresholds of initial and final bedload motion in coarse alluvial channels, in *Sedimentology of Gravels and Conglomerates,* edited by E.H. Koster and R.H. Steel, Canadian Society for Petroleum Geology Memoir 10, pp. 61-68.

Reynolds, 0. 1887. On certain laws relating to the regime of rivers and estuaries, and on the possibility of experiments on a small scale. A Report of the British Association, in Reynolds, 0., *Papers on Mechanical and Physical Subjects, Vol. 11,* 1881-1900, Cambridge University Press, 1901, pp. 326-335.

Rice, C.T., Wilson, B.N. and Appleman, M. 1988. Soil topography measurements using image processing techniques, *Computers and Electronics in Agriculture,* **3**, 97-107.

Richards, K.S. 1982. *Rivers, Form and Process in Alluvial Channels,* Methuen, London, 361 pp.

Römkens, M.J.M., Wang, J.Y. and Darden, R.W 1988. A laser microrelief-meter, *Transactions American Society of Agricultural Engineers,* **31**, 408-413.

Rouse, H. 1939. An analysis of sediment transportation in light of fluid turbulence, SCS-TP-25, Sediment Division, *US Department of Agriculture, Soil Conservation Service,* Washington, DC.

Rouse, H. (ed.). 1950. *Engineering Hydraulics,* Wiley, New York, 1039 pp.

Rouse, H.L. 1994. Measurement of bedload gravel transport: the calibration of a self-generated noise system, *Earth Surface Processes and Landforms,* **19**, 789-800.

Roy, A.G. and Roy, R. 1988. Changes in channel size at river confluences with coarse bed material, *Earth Surface Processes and Landforms,* **13**, 77-84.

Schlichting, H. 1968. *Boundary Layer Theory,* 6th edn, McGraw-Hill, New York, 748 pp.

Schoklitsch, A. 1962. *Handbuch des Wasserbaues,* 3rd edn, Springer-Verlag, Vienna, 475 pp.

Schumm, S.A. 1960. The shape of alluvial channels in relation to sediment type, *US Geological Survey Professional Paper 352-B,* 30 pp.

Schumm, S.A. and Khan, H.R. 1972. Experimental study of channel patterns, *Geological Society of America Bulletin,* **83**, 1755-1770.

Schumm, S.A., Mosley, M.P. and Weaver, W.E. 1987. *Experimental Fluvial Geomorphology,* Wiley, New York, 413 pp.

Seal, C.V., Smith, C.R., Akin, O. and Rockwell, D. 1995. Quantitative characteristics of a laminar unsteady necklace vortex system at a rectangular block-flat plate juncture, *Journal of Fluid Mechanics,* **286**, 117-135.

Shen, H.W. (ed.). 1991. *Movable Bed Physical Models,* Kluwer Academic, Boston, 171 pp.

Simons, D.B., Richardson, E.V. and Albertson, M.L. 1961. Flume studies using medium sand (0.45 mm), *US Geological Survey Water Supply Paper 1498-A,* 76 pp.

Smith, D.G. 1976. Effect of vegetation on lateral migration of anastomosed channels of a glacial meltwater river, *Geological Society of America Bulletin,* **87**, 857-860.

Southard, J.B. 1971. Representation of bed configurations in depth-velocity-size diagrams, *Journal of Sedimentary Research,* **41**, 903-915.

Southard, J.B. and Boguchwal, L.A. 1990a. Bed configurations in steady unidirectional water flows, part 2, synthesis of flume data, *Journal of Sedimentary Petrology,* **60**, 658-679.

Southard, J.B. and Boguchwal, L.A. 1990b. Bed configurations in steady unidirectional water flows, part 3, effects of temperature and gravity, *Journal of Sedimentary Petrology,* **60**, 680-686.

Southard, J.B., Smith, N.D. and Kuhnle, R.A. 1984. Chutes and lobes: newly identified elements in braiding in shallow gravelly streams, in *Sedimentology of Gravels and Conglomerates,* edited by E.H. Koster and R.J. Steel, Canadian Society of Petroleum Geology Memoir 10, pp. 51-59.

Takeda, Y. 1991. Development of an ultrasound velocity profile monitor, *Nuclear Engineering and Design,* **126**, 277-284.

Takeda, Y. 1995. Instantaneous velocity profile measurement by ultrasonic doppler method, *International Journal of the Japanese Society of Mechanical Engineers,* **38B**, 8-16.

Thomson, J. 1879. Flow round river bends, *Proceedings of the Institute of Mechanical Engineers,* pp. 456-460.

Thorne, R.D., Williams J.J. and Heathershaw, A.D. 1989. In situ acoustic measurements of marine gravel threshold and transport, *Sedimentology,* **36**, 61-74.

Tritton, D.J. 1988. *Physical Fluid Dynamics,* 2nd edn, Clarendon Press, Oxford, 519 pp.

Tsujimoto, T. 1996. Coherent fluctuations in a vegetated zone of open-channel flow: causes of bedload lateral transport and sorting, in *Coherent Flow Structures in Open Channels,* edited by P.J. Ashworth, S.J. Bennett, J.L. Best and S.J. McLelland, Wiley, Chichester. pp. 375-396.

Warburton, J. and Davies, T.R.H. 1994a. Variability of bedload transport and channel morphology in a braided river hydraulic model, *Earth Surface Processes and Landforms,* **19**, 403-42 1.

Warburton, J. and Davies, T.R.H. 1994b. Variability of bedload transport and channel morphology in a braided river hydraulic model, errata, *Earth Surface Processes and Landforms,* **19**, Issue 8, ii.

Warburton, J. and Davies, T.R.H. in press. The use of hydraulic models in the management of braided gravel-bed rivers, in *Gravel-bed Rivers in the Environment,* edited by PC. Klingeman, R.L. Beschta, R.D. Komar and J.B. Bradley, Wiley, New York.

Webb, E.K. 1994. Simulating the three-dimensional distribution of sediment units in braided-stream deposits, *Journal of Sedimentary Research,* **B64**, 219-231.

Webb, E.K. 1995. Simulation of braided-channel topology and topography, *Water Resources Research,* **31**, 2603-2611.

Welford, M.R. 1994. A field test of Tubino's model of alternate bar formation, *Earth Surface Processes and Landforms,* **19**, 287-297.

Whipple, K.X., Parker, G. and Paola, C. 1995. Experimental study of alluvial fans, in *Proceedings of the* International Joint *Seminar on Reduction of Natural and Environmental Disasters in Water Environment,* edited by J.H. Sonu, K.S. Lee, I.W. Seo and N.G. Bhowmik, Seoul National University, 18-21 July, pp. 282-295.

Williams, G.P. 1970. Flume width and water depth effects in sediment transport experiments, *US Geological Survey Professional Paper 562-H,* 37 pp.

Williams, R.E. and Rust, B.R. 1969. The sedimentology of a braided river, *Journal of Sedimentary Petrology,* **39**, 649-679.

Wolman, M.G. and Miller, J.P. 1960. Magnitude and frequency of forces in geomorphic processes, *Journal of Geology,* **68**, 54-74.

Wood, L.J., Ethridge, F.G. and Schumm, S.A. 1993. The effects of rate of base-level fluctuation on coastal-plain, shelf and slope depositional systems: an experimental approach, in *Sequence Stratigraphy and Facies Associations,* edited by H.W Posamentier, C.P. Summerhayes, B.U. Haq and G.P. Allen, Special Publication of the International Association of Sedimentologists 18, pp. 43-53.

Wood, L.J., Ethridge, F.G. and Schumm, S.A. 1994. An experimental study of the influence of subaqueous shelf angles on coastal plain and shelf deposits, in *Siliciclastic Sequence Stratigraphy: Recent Developments and Applications,* edited by P. Weimer and H.W Posamentier, Association of American Petroleum Geologists Memoir 58, pp. 381-391.

Yalin, M.S. 1963. An expression for bed-load transportation, *Journal of the Hydraulics Division, Proceedings of the American Society of Civil Engineers,* **89***,* 221-250.

Yalin, M.S. 1971. *Theory of Hydraulic Models,* Macmillan, London, 266 pp.

Young, W.J. 1989. Bedload transport in braided gravel-bed rivers, Ph.D. thesis, University of Canterbury, New Zealand, 187 pp.

Young, W.J. and Davies, T.R.H. 1990. Prediction of bedload transport rates in braided rivers: a hydraulic model study, *Journal of Hydrology (New Zealand),* **29***,* 75-92.

Young, W.J. and Davies, T.R.H. 1991. Bedload transport processes in a braided gravel-bed river model, *Earth Surface Processes and Landforms,* **16***,* 499-511.

Zimmerman, R.C., Goodlett, J.C. and Comer, G.H. 1967. The influence of vegetation on channel form of small streams, *Publication of the International Association of Scientific Hydrology,* **75***,* 255-275.

Zwamborn, J.A. 1967. Solution of river problems with movable bed hydraulic models, Symposium paper S.24, Council for Scientific and Industrial Research, Pretoria, South Africa, 40 pp.