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The "unreasonable effectiveness" of stratigraphic and geomorphic experiments

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ABSTRACT

The growth of quantitative analysis and prediction in Earth-surface science has been accompanied by growth in experimental stratigraphy and geomorphology. Experimenters have grown increasingly bold in targeting landscape elements from channel reaches up to the entire erosional networks and depositional basins, often using very small facilities. The experiments produce spatial structure and kinematics that, although imperfect, compare well with natural systems despite differences of spatial scale, time scale, material properties, and number of active processes. Experiments have been particularly useful in studying a wide range of forms of self-organized (autogenic) complexity that occur in morphodynamic systems. Autogenic dynamics creates much of the spatial structure we see in the landscape and in preserved strata, and is strongly associated with sediment storage and release.

The observed consistency between experimental and field systems despite large differences in governing dimensionless numbers is what we mean by "unreasonable effectiveness". We suggest that unreasonable experimental effectiveness arises from natural scale independence. We generalize existing ideas to relate internal similarity, in which a small part of a system is similar to the larger system, to external similarity, in which a small copy of a system is similar to the larger system. We propose that internal similarity implies external similarity, though not the converse. The external similarity of landscape experiments to natural landscapes suggests that natural scale independence may be even more characteristic of morphodynamics than it is of better studied cases such as turbulence. We urge a shift in emphasis in experimental stratigraphy and geomorphology away from classical dynamical scaling and towards a quantitative understanding of the origins and limits of scale independence. Other research areas with strong growth potential in experimental surface dynamics include physical–biotic interactions, cohesive effects, stochastic processes, the interplay of structural and geomorphic self-organization, extraction of quantitative process information from landscape and stratigraphic records, and closer interaction between experimentation and theory.

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1. Introduction

Recent years have seen a bloom of experiments designed to reproduce aspects of the dynamics of natural landscapes at greatly reduced scale. In this paper, we review and summarize results from such experiments, focusing on work since the book by Schumm et al. (1987), which included work prior to the mid 1980s on erosional landscapes and river channel processes. The scope of our review is limited to physical laboratory experiments, emphasizing channels and channel networks. We consider all three mass-flux regimes: erosional, neutral (bypass), and depositional. We exclude, for the sake of manageability, field experiments, and what might be termed "unit-process" experiments, e.g. ones focused on rheology or sediment transport.

The appeal of experiments in stratigraphy and geomorphology is not hard to understand. Experimental landscapes evolve under controlled conditions, so they allow study of steady states and response to changes in a single variable that would be difficult to observe in nature. In addition, a small, self-contained system can be studied and measured comprehensively to a degree that is rarely possible in the field. Finally, experiments greatly speed up time, through two main effects: characteristic time scales typically increase with spatial scale, so that small systems have intrinsic time scales that are shorter than large systems. The second effect is that in the field, morphodynamic evolution is usually intermittent, occurring during floods, storms, or other high-energy events. Experimental systems can be continuously active — for instance, a steady-state experimental stream is in effect permanently in flood.

Beyond these practical reasons, there is an irresistible fascination in watching a small, controlled landscape evolve, creating dynamic patterns that seem to come from out of nowhere. The surface of an experimental landscape 1 m on a side made of fine, noncohesive sediment could easily comprise some 10^{10} particles, all capable of independent motion. Despite decades of research, the laws by which fluids and particles interact in bulk are known only in imperfect and highly empirical forms. So the capacity of particle-fluid experiments to inform and surprise us should not itself be surprising.

Skepticism about experimental geomorphology and stratigraphy arises mostly from concerns about how representative they are of field-scale systems, and issues of scaling form a thread running through all the papers included in this review. Two views of scaling provide bookends for our review of the experimental results. In Section 2, we summarize the principles of classical dynamical scaling. These principles are invoked in various ways in nearly all the papers we review in Section 3. These include experimental studies of erosional-landscape dynamics and response to large-scale perturbations; depositional basins and stratigraphic recording; submarine systems, primarily density-driven flows; and river channels, alluvial fans, and deltas. We will see that although most of the experiments fall far short of being dynamic scale models, they seem to capture the essence of many important processes in natural systems. This is what we mean by "unreasonable effectiveness" (Wigner, 1960). With these results in mind, we revisit the scaling question in Section 4 and argue for a broader view that combines classical scaling with more recent discoveries about self-similarity and scaling in nature. We close the paper (Section 5) with a short synthesis and outlook for the future of experimentation in surface process science.

Questions of scaling arise from the desire to relate experimental observations to the field. But direct comparison with natural systems is not the only way, or necessarily even the best way, to use geomorphic and stratigraphic experiments. Sedimentary geology and geomorphology are moving away from reasoning by analogy and toward reasoning by analysis. In an analogy-based framework, an experiment is just another analog, so there is no alternative to treating it as a scale model. But statistical variability and the myriad possible forcing scenarios that influence landscape evolution make it difficult to apply analogs to new situations confidently. An analytical framework aimed at process and mechanism, though harder to develop, is more flexible and powerful. In an analytical approach, experiments are used to test models and to develop analytical methods. Both applications require data obtained under controlled conditions. To serve these purposes, an experiment need not be a scale model of a natural system. It need only include enough of the relevant dynamics to serve as a plausible test of the theory or technique. Because experiments generally do not capture the full range of complexity of natural systems, they can provide more exact and specific tests of theory than field cases where the variables in play may be numerous and poorly constrained. Of course, a theory that has been tested experimentally could still fail in the field. But it is difficult to imagine circumstances in which a model or technique that fails under controlled conditions would perform reliably in the field. Once tested, a mechanistic theory can then provide quantitative insight about how the system is affected by changes in scale.

Experimentation is a natural part of the broader process of building predictive insight through analysis. But although the growth of analysis does not require that experiments function as scale models of natural systems, the extent to which the research community accepts experimentally derived insight inevitably depends on how well the experimental systems are thought to represent field-scale dynamics. The comparison between experimental and field systems matters. Thus we begin with the most time-tested way of addressing this issue: the methods of classical dynamical scaling, for the most part developed in the engineering community.

2. Dynamical scaling

2.1. Engineering approach

The aim of classical dynamical scaling is to design experiments that are scale models of natural systems. Before the advent of large-scale computing, scale models were the primary basis for many types of engineering design. In these cases, there is a strong practical motivation to design experiments that can be interpreted as exact, quantitative scale models.

2.1.1. Scaling fluid flow

The simplest case of interest here is fluid flow over structures with fixed geometry. The fixed structure is a geometric scale model of the prototype system. The model may be scaled down homogeneously from the prototype, or the length ratios may differ along different coordinate axes. The latter are referred to as distorted models, and usually the sense is of vertical exaggeration, to produce relatively greater flow depths and topographic slopes, and hence greater driving forces, than one would have with homogeneous scaling. The effect is similar to that of a vertically exaggerated geologic section. In either case, measurements made on the model are scaled up for application to the prototype through the use of dimensionless numbers. The power of classical dynamical scaling is that if all the relevant dimensionless numbers are matched between the experiment and prototype, then any measurement made in the experiment can be converted via a simple algebraic transformation to an equivalent field value.

The engineering applications we are interested in involve fluid motion, and thus rely on methods of fluid-dynamical scaling developed for hydraulic and aeronautical modeling. There is an extensive literature on this (Yalin, 1971; Peakall et al., 1996). For our purposes, it suffices to say that classical dynamic scaling involves two basic steps: (1) identifying, from the governing equations if they are known, or from dimensional analysis if the equations are not known, the complete set of dimensionless variables that characterize the system dynamics; and (2) designing the scale model so that all these dimensionless parameters, along with those characterizing the system geometry and initial and boundary conditions, have the same values as in the field prototype. For a system involving fluid flow over boundaries with fixed geometry, the minimum set of dynamical dimensionless variables would be the Froude number Fr and the bulk Reynolds number Re defined as, respectively, $Fr = U/(gL)^{1/2}$ and $Re = UL/\nu$ where U and L are velocity and length scales, respectively, g is gravitational acceleration, and ν kinematic viscosity. For experiments involving density variation, such as turbidity currents, the Froude number is generalized to the so-called densimetric form: $Fr_{\mathrm{d}} = U/(g([\rho_{\mathrm{f}} - \rho_0]/\rho_0)L)^{1/2}$ where ρ_{f} is the density of the flow and ρ_0 is that of the ambient fluid. Other dimensionless numbers come into play if either the prototype or experiment is influenced by physical processes like sediment transport, surface tension, heat flow, planetary rotation, or electromagnetic fields. For very small-scale flows, the Weber number, which measures the importance of surface tension forces relative to fluid inertia, comes into play (Peakall and Warburton, 1996). It is defined as $We = \rho_{\mathrm{f}} U^2 L/\sigma$, where σ is the surface tension.

2.1.2. Scaling sediment transport

A granular bed brings with it new length scales and dynamics that are reflected in additional dimensionless variables. The overall scale range between a transport system and the sediment comprising it is measured by what we will term the granularity, $Gr = L_{\mathrm{s}}/D$ where L_{s} is the length of the system in question and D is a representative grain size. A general index of noncohesive-sediment transport is the dimensionless shear stress $\tau_* = u_*^2/([s-1]gD)$ where u_* is the friction velocity, and s is the sediment specific gravity. The state of suspended transport is measured by the simplified Rouse number $Ro = w_{\mathrm{s}}/u_*$ where w_{s} is the settling velocity. The state of flow around particles near the bed is measured by the particle Reynolds number $Re_{\mathrm{p}} = u_* D/\nu$. Applying the generalized settling velocity formula of Ferguson and Church (2004), the three parameters can be related via $Ro^{-1} = 18(\tau_*/R_*) + a\tau_*^{1/2}$ where $a = [0.75(s-1)]^{1/2}$.

The discussion so far pertains only to noncohesive sediment, i.e. particles for which resistance to movement is scaled to particle weight because surface forces are negligible. The mechanics of cohesive sediment are as yet not sufficiently understood to provide a mechanistic basis for scaling cohesive-sediment dynamics. However, cohesive sediment appears to move through the environment mainly as agglomerates (flocs), which in turn appear to behave non-cohesively (Schieber et al., 2007). Even if floc transport is effectively noncohesive, the problems of initiation of motion and consolidation of cohesive sediments in experiments and the field remain as important research topics.

Values of τ_* and *Ro* typical of field conditions can readily be reproduced in the laboratory, if necessary using light sediment (e.g. plastic, walnut shells, coal) to reduce *s* and w_s. Matching Re_p and *Gr* are more problematic, however, as we will see below.

2.1.3. Limits of classical dynamical scaling

Dynamical scaling offers a rigorous and well defined method for imposing experiment conditions that not only match the prototype system in appearance but also are guaranteed to reproduce it dynamically. Unfortunately, meeting all the conditions for exact dynamical scaling in experiments involving flowing water is nearly impossible, even in engineering problems for which the system geometry is known and static. The biggest obstacle is the bulk Reynolds number. Of all the fluids readily available for experimental use, none has a kinematic viscosity significantly less than that of water. This means that the only way of reaching the high *Re* values typical of natural-scale flows is to use extremely high velocities. This typically makes it impossible to match *Fr*.

In addition, scale models generally cannot reproduce prototype geometry perfectly. Unless the prototype is unusually simple, it is impractical to model the fine scale geometry (e.g. bed roughness) exactly. Usually this is dealt with heuristically, for example, by tuning artificial bed roughness to force model velocities to match (under the appropriate scaling) a set of known values in the prototype.

These obstacles sound worse than they really are. Nearly all of the river and coastal civil infrastructure in the developed world was designed using imperfectly scaled models subject to the problems described above. Evidently, perfect dynamic scaling is not necessary to get useful results and solve practical problems. This is the first in a series of "bumps" against the limits imposed by classical dynamical scaling that we will encounter. We will look now at how engineers have overcome the problem, and in subsequent sections will try to extend these ideas to develop new ways of using experiments in geomorphology and stratigraphy.

The primary workaround that experimental engineers have used for the *Re* problem is an empirical principle called *Reynolds-number independence*. The basic idea is that as long as *Re* is high enough in both the prototype and experimental systems, its exact value does not strongly influence the overall dynamics. At a minimum, "high enough" is taken to mean that the small-scale flow is fully turbulent. The value of *Re* in a turbulent flow controls the ratio of the largest (energy-producing) to the smallest (energy-dissipating) scales of the turbulence. Since the former are controlled by the overall flow geometry, the main effect of increasing *Re* is to reduce the turbulent fine scale. Reynolds-number independence amounts to asserting that the mean flow dynamics is relatively insensitive to the turbulent fine scale. Standard graphs of drag coefficient for objects of a given shape versus *Re* show that over a wide range of *Re* the drag coefficient is constant; this is a good example of Reynolds-number independence. Decades of success with engineering scale models show that it works in practice.

Reynolds-number independence is trickier to apply in sediment–fluid systems. The culprit is the sand-sized and finer sediments that not only constitute the majority of sediment on Earth but also are now known to be dynamically important even where the predominant grain size is gravel (Wilcock and Crowe, 2003). Fine sediments in the field lead to very large values of *Gr*. Moreover the small length scale of sand or silt implies small values of the prototype Re_p, undermining the *Re*-independence argument and requiring an effectively impossible matching of Re_p values across a spectrum of grain sizes. There is a rough analogy between the effect of *Re* on the fluid flow and the effect of fines on the sediment flow — both are tied to the fine scale of the dynamics. Unfortunately the effects of fines on sediment transport are less well understood than those of fine-scale turbulence. The presence of fine sediment in the field creates a second hurdle to imposed dynamical scaling at least as formidable as that created by the low kinematic viscosity of water.

A more subtle issue is that most experiments with sediment, especially landscape experiments, involve the development of self-formed topography — the essence of morphodynamics. In such experiments morphologic features like slope and channel geometry cannot be set directly but are determined internally. Without using low-density sediment (e.g. plastic or walnut-shell), scaling down the depth generally means increasing slope to provide the necessary shear stress. High slopes and shallow depths lead to high (~1) values of *Fr*. If the target system is relatively coarse-grained, field-scale *Fr* values may be in this range, and in these cases Froude scaling, in the form of "generic Froude modeling" (Ashworth et al., 2007) or distorted Froude modeling (Cazanacli et al., 2002), can be used. But natural sand-bed and finer-grained rivers are more commonly subcritical. Subcritical flow with self-formed channels requires either extremely low sediment supply or low-density sediment, and has proved very difficult to reproduce experimentally (Martin, 2007).

2.2. System-scale nondimensional variables

Geomorphic and stratigraphic experiments bring with them additional governing parameters and hence additional dimensionless variables. Most natural geomorphic systems display a self-organized hierarchy of dynamics over the scale range presented by the system (e.g. Werner (1999)). In this context, one can define sets of dimensionless variables, analogous to the classical fluid-dynamics variables discussed above, pertaining to higher levels of the hierarchy. Examples include the depositional-system Peclet number introduced by Swenson et al. (2000) and the landscape Reynolds number proposed by Haff (2007).

If one has a complete set of governing equations, dimensionless numbers can be derived by scaling the terms in the equations. For landscape systems, the equations are still being developed, but one can start creating dimensionless variables using naturally occurring system length and time scales. Obvious length scales range from grain size to the lengths of specific transport domains (e.g. erosional catchment, alluvial fan); in most cases these can be estimated directly from imagery or other plan-view information. Time scales, which begin to bring dynamics into the picture, are much harder to estimate. At the scale of whole depositional systems, the basin diffusional equilibrium time (relaxation time) T_{eq} proposed by Paola et al. (1992a) provides a reference time scale for nondimensionalization. Paola et al. (1992a) showed using a simple numerical model how basin response to forcing (in this case, migration of the fluvial gravel–sand transition) differs qualitatively depending on whether the forcing is slow (i.e. forcing period $T_i > T_{eq}$) or rapid ($T_i < T_{eq}$). Van Heijst et al. (2001) refer to the dimensionless ratio $Br = T_i / T_{eq}$ as the "basin response factor". T_{eq} values for experiments are typically measured in hours, while at field scales T_{eq} can range from thousands to millions of years, depending on system scale and transport efficiency. Postma et al. (2008) have advanced the approach by developing methods using nonlinear diffusion models for systems responding to base-level changes.

In erosional systems, T_{eq} is the time needed to achieve steady state between erosion and uplift, measured in terms of the total erosional relief H_r and uplift rate w_u. A simple estimate based on numerical modeling is $T_{eq} = 3H_r / w_u$ (Howard, 1994). Allen (2008) presents more sophisticated estimates in a recent review of landscape time scales. More importantly, Allen emphasizes that erosional and depositional systems are so strongly coupled that we should be thinking in terms of time scales for the entire system. In an analysis of linked erosional and depositional systems, Beaumont et al. (2000) expanded on the idea of slow and rapid forcing to include the intermediate case where $T_i \approx T_{eq}$, and the additional case of impulsive forcing where T_i

becomes vanishingly short. Several authors (Carretier and Lucazeau, 2005; Densmore et al., 2007; Allen, 2008) have stressed as well that landscape systems are characterized by a number of different time scales; these can lead to complex response on multiple time scales to even simple perturbations. Given that most erosional landscapes are spatial fractals, it also seems possible that they have continuous power–law distributions of response time, over some range of time scales. In this case, it would be critical to know the limits of the range of power–law behavior. Allen (2008) has also introduced the valuable idea of "reactive" versus "buffered" systems, to characterize the degree of system sensitivity to input signals. For the reasons discussed above, this concept is dependent on both the system and the input signals to which it is subjected.

From the point of view of experimental landscape research, intrinsic length and time scales provide a natural way of forming dimensionless numbers that can be used to analyze large-scale similarity between experimental and field systems. For example, if a given process is known to be associated with a characteristic response time scale T_r then a necessary (though not sufficient) condition for large-scale similarity between experimental and field cases is that the time-scale ratio T_i/T_r be the same in the two systems, or that T_i be so short that its exact value does not matter. In addition to their role in similarity analysis, characteristic length and time scales provide natural "measuring sticks" that can be used to put experimental results in the context of field length and time scales. Any of the time scales discussed above (e.g. T_{eq}) can serve as an appropriate reference time scale for large-scale comparisons. Another, not tied to a specific model, is the time required to create a given volume V_s of morphodynamic change, given by $T_{ref}=V_s/Q_s$ where Q_s is the volumetric sediment flux (Van Heijst et al., 2001). A useful set of length and time scales for channelized systems is given by Sheets et al. (2002): h_{ch} and $h_{ch}/\bar{r}$ where h_{ch} is the average channel depth and $\bar{r}$ is the average deposition rate.

The large-scale dimensionless numbers discussed in this section measure similarity at system scales. The classical dimensionless numbers discussed in the previous section measure fine-scale similarity in terms of local fluid and sediment dynamics. If the large-scale dynamics is the main focus, large-scale dimensionless numbers are the appropriate starting point for comparing experiments to the field. The key questions are how the large-scale dynamics is coupled to the fine scales, and how sensitive the former is to the latter. We will return to this point later. For now, we note that mechanistic insight about the coupling of dynamics across scales would be a major step toward understanding the dynamics of multi-scale morphodynamic systems in general, in addition to being a major advance in relating experiments to field-scale systems.

3. "Unreasonable effectiveness" in action: results from stratigraphic and geomorphic experiments

Before looking at experimental results, we review two key ideas that will recur throughout the discussion. The first is morphodynamic steady state, the condition mentioned in the previous section in which the topographic elevation remains constant, under suitable averaging, such that tectonic forcing is balanced by erosion and/or deposition. The steady state is illustrated by the simplest form of Exner mass balance equation that includes tectonic forcing:

$$\frac{\partial \eta}{\partial t} - w_t = \frac{-\partial q_s}{\partial x} \tag{1}$$

where η is the topographic elevation relative to some fixed datum, t is the time, w_t is the vertical speed of tectonic movement of the crust (positive upwards), q_s is the average sediment flux per unit of system width, and x is the downstream distance. The steady state is the case for which $\partial\eta/\partial t=0$. (Strictly speaking this is the topographic steady state identified by Willett and Brandon (2002), which as they point out need not precisely correspond with flux steady state. For present purposes we will ignore this distinction.) Without tectonics, this requires that q_s be constant in x, the "graded" condition defined originally by Mackin (1948). Steady states are equally possible for tectonic uplift and subsidence; in those cases the rate of sediment gain/loss is such as to balance the tectonic term w_t. Note that the simplified mass balance in Eq. (1) leaves out a number of important effects that apply to natural systems, discussed in detail in Paola and Voller (2005). In addition, depending on the overall system configuration, a long-term steady state may not be possible (e.g. Muto et al., 2007). Nonetheless, steady or quasi-steady states remain a useful if idealized reference state.

The second general idea we will need is the distinction between autogenic and allogenic dynamics. *Autogenic* refers to morphologic change that arises from the system's internal dynamics, as opposed to *allogenic* changes that result from external forcing. By definition, autogenic variability includes all variability that occurs once steady state has been reached. In addition, autogenic behavior occurs during, and often interacts in interesting ways with, variable external forcing. Autogenic behavior can be stochastic or deterministic, and occurs over a wide range of space and time scales. One might say informally that autogenic dynamics makes systems behave as if they had a mind of their own. Stochastic autogenic variability is the morphodynamic equivalent of weather — a kind of "morphodynamic turbulence". Like weather, stochastic autogenic behavior appears to be an example of deterministic chaos (Slingerland, 1990): as far as is known, the morphodynamic processes in question are described by deterministic equations, but as they evolve their evolution becomes increasingly difficult to predict exactly, in the same way that the weather predictions lose reliability into the future. With few exceptions (Rubin, 1992; Murray and Paola, 1996), relatively little has been done to analyze autogenic morphodynamics formally for chaotic behavior, or to measure the rate at which predictability decreases in time.

Deterministic autogenic dynamics is a little harder to pin down, because one can easily end up referring to all forms of nonlinearity as autogenic behavior. For now, we note that there is general agreement that "autogenic" includes all forms of behavior that clearly arise through internal processes and thus continue indefinitely under steady-state conditions. Proposed generalizations will be discussed later in this review.

Experimental systems are particularly suited to study both steady states and autogenic dynamics. Natural systems are subject to a wide spectrum of external effects, making steady state difficult to pin down precisely and often blurring the distinction between autogenic and allogenic effects. The capacity of morphodynamic systems to create interestingly complex forms of self-organized behavior – e.g. pattern formation, stochastic variability, and abrupt change – is one of the main features that makes them such attractive targets for experimentation.

3.1. Erosional landscapes

3.1.1. Experimental methods

A wide range of experimental systems have been devised to investigate landscape evolution and the development of self-organized channel networks. This approach is exemplified by the Rainfall Erosion Facility (REF) (Schumm et al., 1987) at Colorado State University. The REF comprises a basin 15 m by 9.2 m equipped with a sprinkler system to apply rainfall. Base-level is typically fixed in these experiments and a preformed topography is eroded through rainsplash, overland flow and/or groundwater sapping (Schumm et al., 1987; Pelletier, 2003).

The years since development of the REF and comparable facilities have seen the creation of the first generation of numerical landscape models capable of reproducing the development of self-organized erosional channel networks. The models used minimal representations of runoff and erosion processes, and are well suited to look at interaction with tectonic forcing (Willgoose et al., 1991; Chase, 1992; Kooi and Beaumont, 1994; Rodriguez-Iturbe and Rinaldo, 1997; Densmore et al., 1998; Pelletier, 1999, 2004). The development of these numerical models has been accompanied by a shift in experimental focus to systems capable of simulating tectonic uplift (or, equivalently, relative base-level fall), enabling exploration of the relation among tectonic uplift, erosion, and climate (i.e. precipitation). The experimental systems apply rainfall via misters to a column of erodible material (Fig. 1). The erodible column is moved relative to base-level either by forcing it upward past a fixed sill (Crave et al., 2000; Bonnet and Crave, 2003; Lague et al., 2003; Babault et al., 2005; Bonnet and Crave, 2006; Turowski et al., 2006; Babault et al., 2007), or by moving one or more outlet points down the column (Hasbargen and Paola, 2000, 2003; Bigi et al., 2006). The experimental systems minimize the influence of groundwater sapping and rainsplash driven diffusion through the use of noncohesive but tightly packed impermeable grains and very small (μm scale) rain droplets; erosive mechanisms are therefore restricted to overland flow, channelized flow and landsliding.

Developing an erosive system with a moving base-level and reasonably uniform rainfall is not easy, so to date tectonic geomorphology experiments have mostly been fairly small (<1 m width). Typical measurements include topography by photogrammetry (e.g. the French ATOS system (Turowski et al., 2006)) or a scanning laser, and sediment output flux via weight or turbidity of output flow (Hasbargen and Paola, 2003).

3.1.2. Steady-state erosional landscapes

One outcome of research on the interaction of erosion and tectonics was the idea introduced above that a landscape subject to temporally constant uplift and rainfall eventually reaches a steady state in which uplift and erosion rates balance. The experimental landscape group at the University of Rennes has used the system

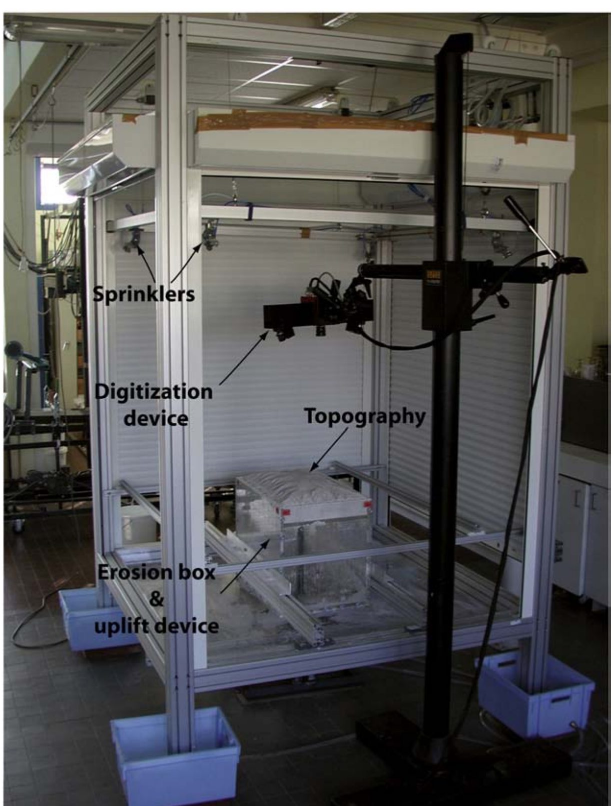

Fig. 1. University of Rennes device for erosion experiments. From Bonnet and Crave (2006).

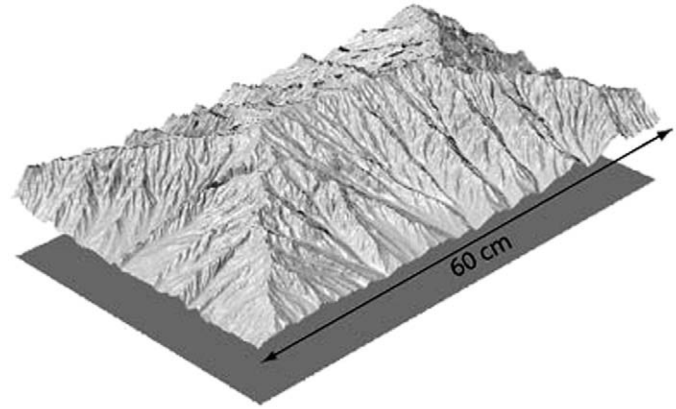

Fig. 2. Shaded-relief DEM of steady-state topography produced using the setup shown in Fig. 1. From Bonnet and Crave (2006).

described in the previous section to study landscape response to different rates (and ratios) of uplift and rainfall. These experiments are summarized in Bonnet and Crave (2006). In the Rennes experiments a constant low rate (mm/h) of relative uplift exposes an erodible substrate along all four edges of their device, leading to the emergence of a multi-catchment linear mountain range with a central drainage divide (Fig. 2).

The Rennes experiments (Crave et al., 2000; Bonnet and Crave, 2003; Lague et al., 2003) also show the strong control of both rainfall rate and uplift rate on total steady-state topographic relief. For a limited range of conditions dependence on rainfall rate seems to be roughly inverse-linear. The trend with uplift rate also appears linear, with the slope depending on material properties. Total relief increases with system size, as one would expect. Turowski et al. (2006) were also able to show, using very high-resolution topographic measurements, systematic reduction in channel width and cross-sectional area with increases in uplift rate in a 1-m scale laboratory experiment. Finally, there is a well defined finite residual relief for zero uplift rate, which the authors interpret as evidence for a threshold shear stress for erosion. This residual relief represents the limit to which erosion can create low-relief landscapes (peneplains).

Experimental slope–area curves fall into three domains, characterized by differing power–law slopes, whose physical origin is not entirely clear. Overall, though, experimental power–law exponents on slope–area relations are significantly lower than most field values. Morphologically, this is expressed as lower concavity in stream profiles in experiments relative to the field. One interpretation of the low exponents and concavity is that the small-scale systems are dynamically similar to debris-flow dominated landscapes in nature (Bonnet and Crave, 2006).

Babault et al. (2005) investigated the effect of piedmont sedimentation on steady-state relief by allowing sediment accumulation over an apron surrounding the uplift region (Fig. 3). Ongoing sedimentation prevents development of steady-state relief, since the reference level for the relief continues to rise. Once a bypass condition develops in the piedmont, the steady-state erosional relief is higher with the piedmont than without it, but relative to the top of the depositional apron it is similar to what it would be without the deposit. Overall the results are consistent with the theoretically based claim by Carretier and Lucazeau (2005) that range-front sedimentation has a strong influence on the evolution of the uplift, but the experimental and numerical model results have apparently not been compared in detail.

For steady input conditions, the steady state produced by nearly all numerical models is static: the balance between uplift and erosion is exact over the whole domain. The numerical topography, which reproduces observed statistical properties of natural erosional topography, is fixed in time. (An important exception is the numerical

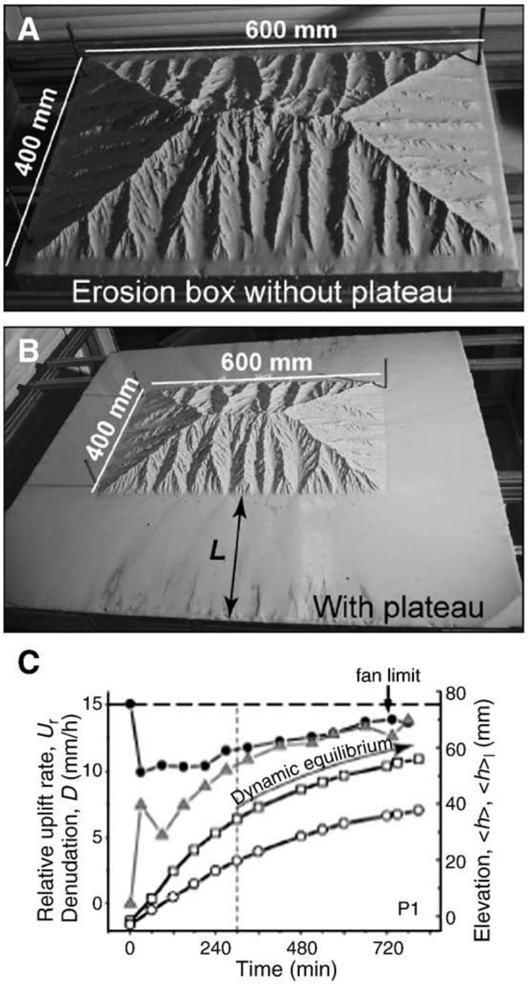

Fig. 3. A tectonically rising block (A), and with addition of a depositional apron (B). Ongoing aggradation if the apron is sufficiently wide prevents development of a topographic steady state in the rising block. From Babault et al. (2005).

model of Densmore et al. (1998), which produced ongoing topographic evolution through landsliding.) Sapozhnikov and Foufoula-Georgiou (1996a) argued on the basis of these "frozen topography" results that the numerical models that produced them could not be considered examples of self-organized criticality, as other workers had proposed. Hasbargen and Paola (2000, 2003) carried out a series of experiments designed to test the idea of static steady-state erosional topography by subjecting an experimental landscape to constant relative uplift and rainfall. Their experiments were designed around the criterion presented in Section 2.2 that attainment of erosional steady state requires erosion through roughly $3H_r$. Thorough evaluation of the nature of a steady-state erosional catchment requires a system capable of substantially exceeding this minimal condition, so the erosion facility used by Hasbargen and Paola was tall relative to its length.

The principal result of a series of experiments using different ratios r_r/w_u ("water/rock ratio"; r_r is the rainfall rate in units of velocity) is that, although an average steady-state condition is readily achieved, it is anything but static (Fig. 4). Rather, it is characterized by vigorous ongoing dynamic modification of the landscape, including major reconfigurations of the drainage network. Processes of change included migration of ridges through asymmetric erosion, landsliding, knickpoint migration, and stream capture. Stream capture in some cases led to rapid, major changes in the drainage pattern, making it an erosional analog of channel avulsion in depositional systems. The autogenic dynamics was closely associated with sediment storage and release — another strong parallel between erosional and depositional systems. A common storage-release mode was the creation of temporary depositional areas that later self-channelized, releasing the stored sediment. Channelization was often triggered by migrating knickpoints. The net effect of these distributed "hot spots" of sediment production turning on and off over the system is a highly variable sediment output at the basin exit (Hasbargen and Paola, 2000, 2003).

The steady-state landscapes produced by Hasbargen and Paola show consistently higher levels of dynamism than those produced by the Rennes group described above. The Rennes group has put a good deal of effort into their rainfall system, and Bonnet and Crave (2006) suggest that the higher level of topographic migration reported by Hasbargen and Paola is caused by variability in precipitation. It would be remarkable if landscapes were that sensitive to random spatial fluctuations in rainfall, but it seems more likely to us that it is associated with the differences in the experimental geometry. In the Rennes group's range-scale experiments catchment exits are free to find their own locations along the uplift margin (Fig. 2), whereas in the single-catchment case of Hasbargen and Paola, the outlet location is pinned. It would be an interesting paradox if relaxing the constraint on a key downstream boundary condition leads to decreased dynamism within the system.

3.1.3. Response of erosional landscapes to change

The ability to control boundary conditions in the experimental landscapes makes them ideal for studying landscape response to imposed changes in external forcing conditions. Thus it is surprising that so much of the recent experimental work on erosional systems has focused on various forms of steady state. However, Babault et al. (2007) extended their study of the effect of piedmont sedimentation discussed above by looking into the effect of adding a piedmont coincident with the cessation of uplift. As in the previous study, piedmont sedimentation in effect resets the base level. The local relief is reduced following a similar time trend but piedmont sedimentation leaves the final mean elevation higher. The results provide an alternative to the common explanation of high-elevation, low-relief surfaces as indicators of tectonic uplift of low-elevation peneplains. Inasmuch as development of a depositional piedmont is a natural part of the evolution of an uplifting mountain range, relief reduction at high elevation could be considered a natural part of orogen development, though there remains a finite limiting relief at which erosion rates cease or drastically slow down.

Bonnet and Crave (2003) used the steady-state landscapes produced in the Rennes facility as a starting point for an investigation of the effect of a change in rainfall on a landscape with constant uplift. The mean elevation and relief increase under a decrease in rainfall (Fig. 5), and vice versa. The results support suggestions based on field work and theoretical modeling that climate change alone could increase the height of mountains; the experiments indicate that the erosional morphology produced by rainfall changes is similar to that resulting from tectonically driven uplift. The latter, however, leads to a permanent increase in sediment production while climatically induced uplift does not (Bonnet and Crave, 2003).

A number of the studies discussed above include data on the transient evolution of their systems toward steady state. Pelletier (2003) focused on the development of drainage basins from an initial condition, using the prototype of modern experimental erosion facilities, the Rainfall Erosion Facility at Colorado State University (Fig. 6). The goal in this case was both to study and constrain the effects of autogenic processes, especially terrace formation, and to compare the evolution of the experimental system with the many numerical models proposed for drainage basin evolution. One important finding was that lateral channel migration, missing from many numerical models, was critical to drainage capture and thus to integration of the drainage network. The initial configuration of the eroded mass significantly influenced the form of the final basin. This sensitivity seems to be at odds with the apparent

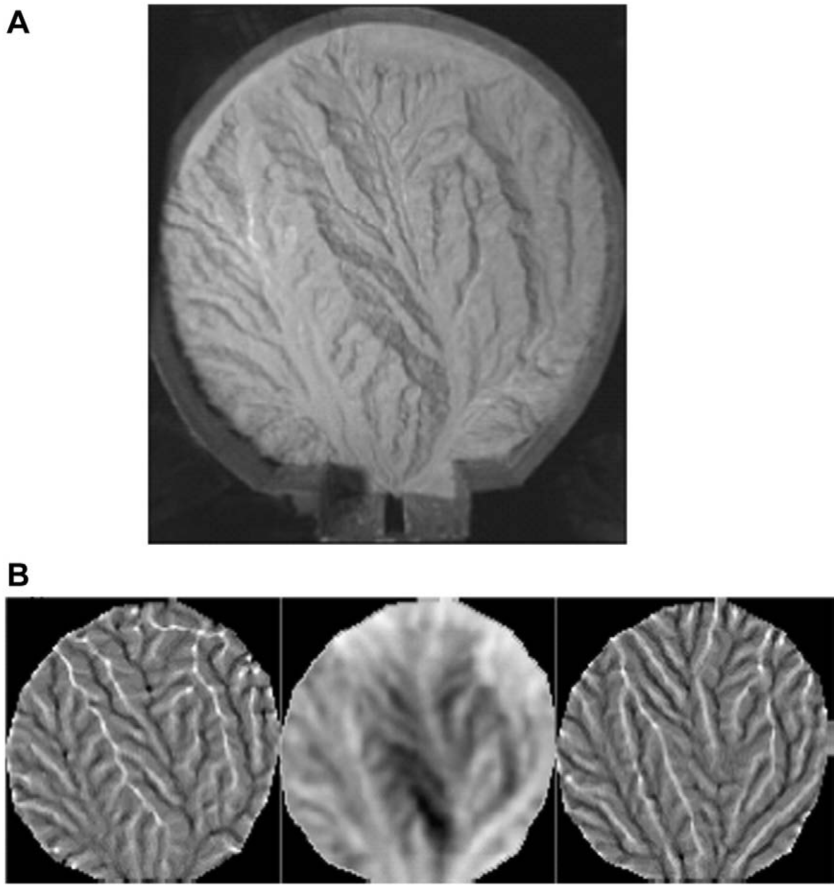

Fig. 4. Dynamic steady-state erosional topography in a system with a single, fixed outlet. (A) surface image, (B) panels showing (left, right) shaded-relief images of steady-state erosional surfaces separated by total erosion of approximately 0.33 relief distances and (center) gray-scale image of the spatial variability of erosion between them. From Hasbargen and Paola (2003).

statistical consistency of fractal drainage networks in natural landscapes despite differences in forcing history and lithology. Perhaps these networks are less generic, and thus contain more information on forcing history, than is currently thought.

A recent major advance is to combine experimental geomorphology and structural geology by applying rainfall to an experiment in which an erodible substrate is subjected to continuous tectonic shortening (Graveleau and Dominguez, 2008). As shown in Fig. 7, the extent to which the experiment recreates natural tectonic and geomorphic patterns is extremely striking. In our view, this experiment is an important first example of what we hope will be a next generation of landscape experiments in which the morphologic and tectonic evolution are coupled and studied at comparable levels of detail.

3.1.4. Summary and next steps: erosional systems

The most striking result of all the experiments carried out on landscape evolution to date is that they work as well as they do, capturing spatial structure and (as far as can be determined from the field) important aspects of the time dynamics as well. This is a good example of "unreasonable effectiveness" in action: in the classical sense the experiments reviewed above cannot be considered to be scale models. The experiments are reduced in scale relative to catchments and orogens by many orders of magnitude. In some cases the fluid flow is not even turbulent, and *We* values for flows of the order of mm deep moving at 1–10 mm/s are much less than 1.

Yet even where whole orogens are being simulated, the experimental drainage patterns are strikingly similar to those seen in natural mountain ranges, as illustrated in the previous section. There is also strong similarity between experimental and field-scale mountains in the structure of individual catchments. Linear mountain ranges, which are common, tend to have a regular spacing ratio of 2 (Hovius, 1996), close to the value of 2.6 observed in experiments such as those carried out by the Rennes group. Regular catchment spacing and geometry are related to the regular (fractal) structure of channel networks which commonly have a length–width ratio of 2; this ratio appears be a function of the common set of relations among stream length, stream order, drainage area, channel slope and confluence angle (Kirchner, 1993; Hovius, 1996; Rodriguez-Iturbe and Rinaldo, 1997). Geometry, of course, is intrinsically scale independent, so evidently a good deal of the self-organized geometry of erosional channel networks is insensitive to the fine details of the processes that create it.

Equally important are features that experimental mountains do not reproduce. Experimental mountain ranges generally create an excessive proportion of high slopes. This implies that the threshold failure slope in the experiments is higher than it is in nature, consistent with experimental studies of mass failure generally. In addition, although slope decreases with catchment area, values of experimental slope–area exponents are generally lower than in the field. Low values of the exponent are thought to represent the steepest portion of a channel where debris flows dominate river incision (Montgomery and Dietrich, 1992; Lague et al., 2003; Stock and Dietrich, 2003). Although debris flows are common in some experiments (Bigi et al., 2006), most of the erosion still appears to be fluvial, so debris flows apparently are not the only cause of the lack of concavity in experimental erosional systems. Evidently the relative contribution of downstream increases in water discharge (estimated via contributing catchment area) to erosion rate is weaker at small scales than in the field. The simplest explanation of this would be some form of scale dependence in the overall erosion law. This is a topic for further work, but we note that Malverti et al. (2008) have recently shown that sediment transport in laminar flows obeys similar

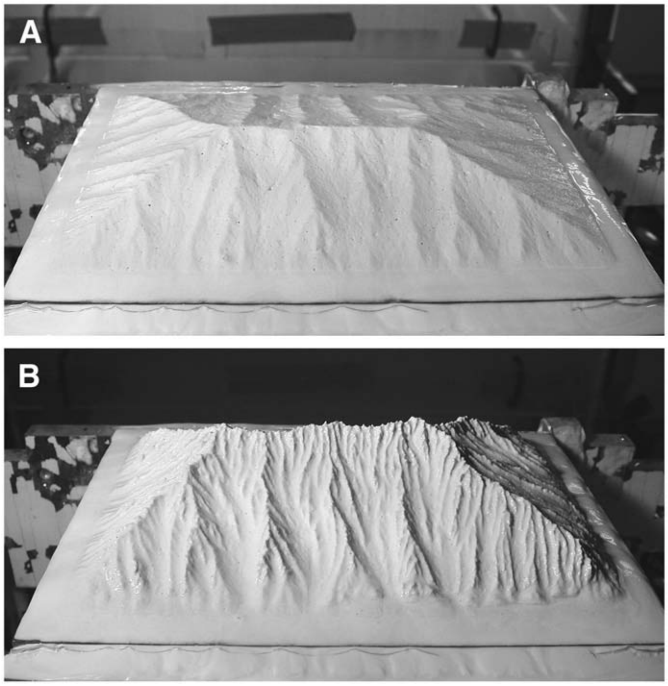

Fig. 5. Images of erosional topography with a constant uplift rate and (A) high and (B) low rainfall rate. From Bonnet and Crave (2003).

laws to those established for full scale turbulent flows. So the origin of the scale dependence in the slope–area relation remains to be explained.

An overall theme emerging from erosional-landscape experiments is the importance and variety of autogenic processes. These range from autogenic terrace and fan formation due to sediment storage and release within an overall erosional system, to stream capture and piracy – the erosional analogs of avulsion and channel switching – and finally to large-scale processes like the relief damping associated with piedmont deposition proposed by Babault et al. (2007). The ongoing autogenic variability of steady-state erosional systems calls for further investigation. The first question, why this behavior does not appear in conventional erosional-landscape models, has several possible answers. Densmore et al. (1998) produced continuing variability by including landslides, which are in effect large, destabilizing flux events, in their landscape model. Willett et al. (2001) showed that horizontal shortening leads to drainage instability as well. Finally, Pelletier (2004) found the simplest solution: he showed that ongoing landscape dynamics under average steady-state conditions could be produced by modifying the conventional landscape erosion models to allow for distributed flow routing, as opposed to routing all water down the path of steepest descent.

The next question is whether autogenic dynamism in erosional landscapes can be observed at field scales. Recent developments in cosmogenic nuclide erosion rate measurements may allow us to measure erosional dynamism in any of the mountain ranges thought to be in steady state (e.g. New Zealand Alps, Taiwan). Divide migration, for example, would be indicated by systematic differences in erosion rate between the two sides of a ridge crest. Another approach would be to use sequences of fill terraces of known age from which quartz samples can be collected. The mean catchment erosion rate at the time of deposition of each terrace can be calculated irrespective of depositional process, allowing a variable erosion history to be reconstructed at the temporal resolution of the terraces.

The discovery that the erosional steady state is statistical rather than exact also reopens the question of whether erosional landscapes show power–law scaling in their kinematics. Power–law kinematics would be a necessary though not sufficient condition for erosional landscapes to show temporal self-similarity in the sense proposed by Sapozhnikov and Foufoula-Georgiou (1997, 1999). It also reopens the possibility of landscapes being self-organized critical systems.

A clear trend in experimental erosional-systems geomorphology is away from steady states and toward investigation of how external changes are recorded in landscapes. The work of Babault et al. (2007) and some of the earlier work compiled in Schumm et al. (1987) serve to highlight the potential of experiments using erosion facilities to shed light on how external changes are propagated and recorded in erosional topography. As we will see in Section 3.2, experiments with depositional systems show that even very simple scenarios of external forcing are often recorded in fascinatingly complex ways. Is the same true of erosional landscapes? How far can we push parallels in the recording process in erosional versus depositional systems? How does the menagerie of local autogenic processes that has emerged from field and experimental research interact with nonlinear large-scale system response to create a record of past history in erosional landscapes? Would a geomorphologist presented with an experimental landscape produced by an unknown sequence of forcing events be able to deduce these events from the form of the landscape alone?

Finally, the work of Graveleau and Dominguez (2008) illustrates the potential for experiments that link structural and geomorphic evolution. Fault–fold systems and river–channel networks are the two most widespread forms of self-organization affecting the Earth's surface, so it is natural to ask how they interact. It is now clear that experiments are an effective way of addressing this. If anything, perhaps they are too effective — the complexity of the co-evolving system, even under the simplified conditions of an experiment, still

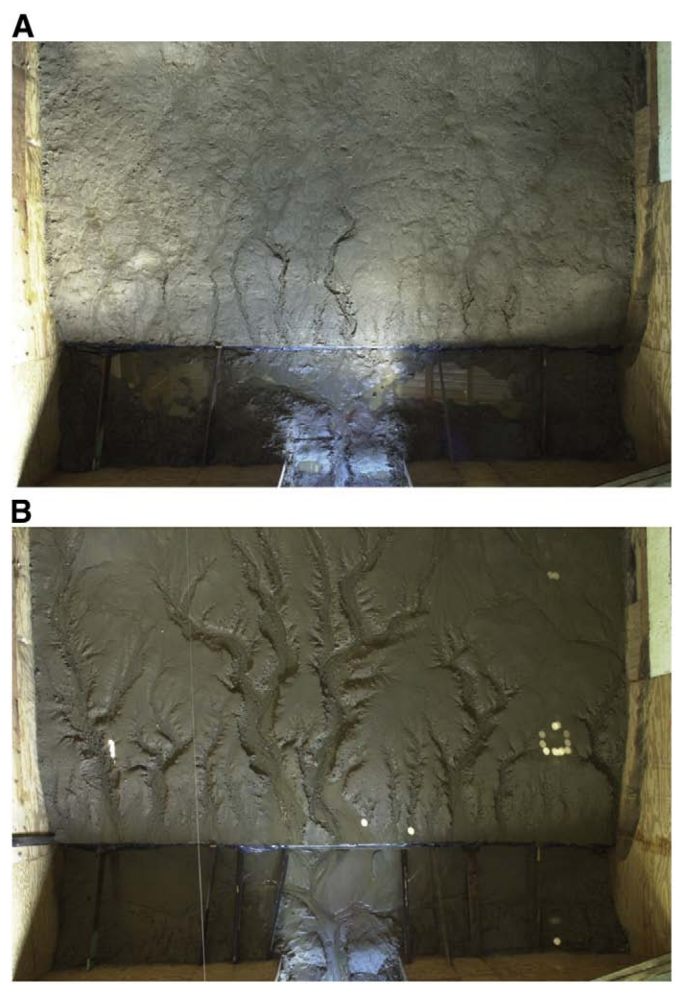

Fig. 6. Early (A; elapsed time = 30 min) and late (B; elapsed time = 10 h) states in the evolution of a drainage network in the Rainfall Erosion Facility. From Pelletier (2003).

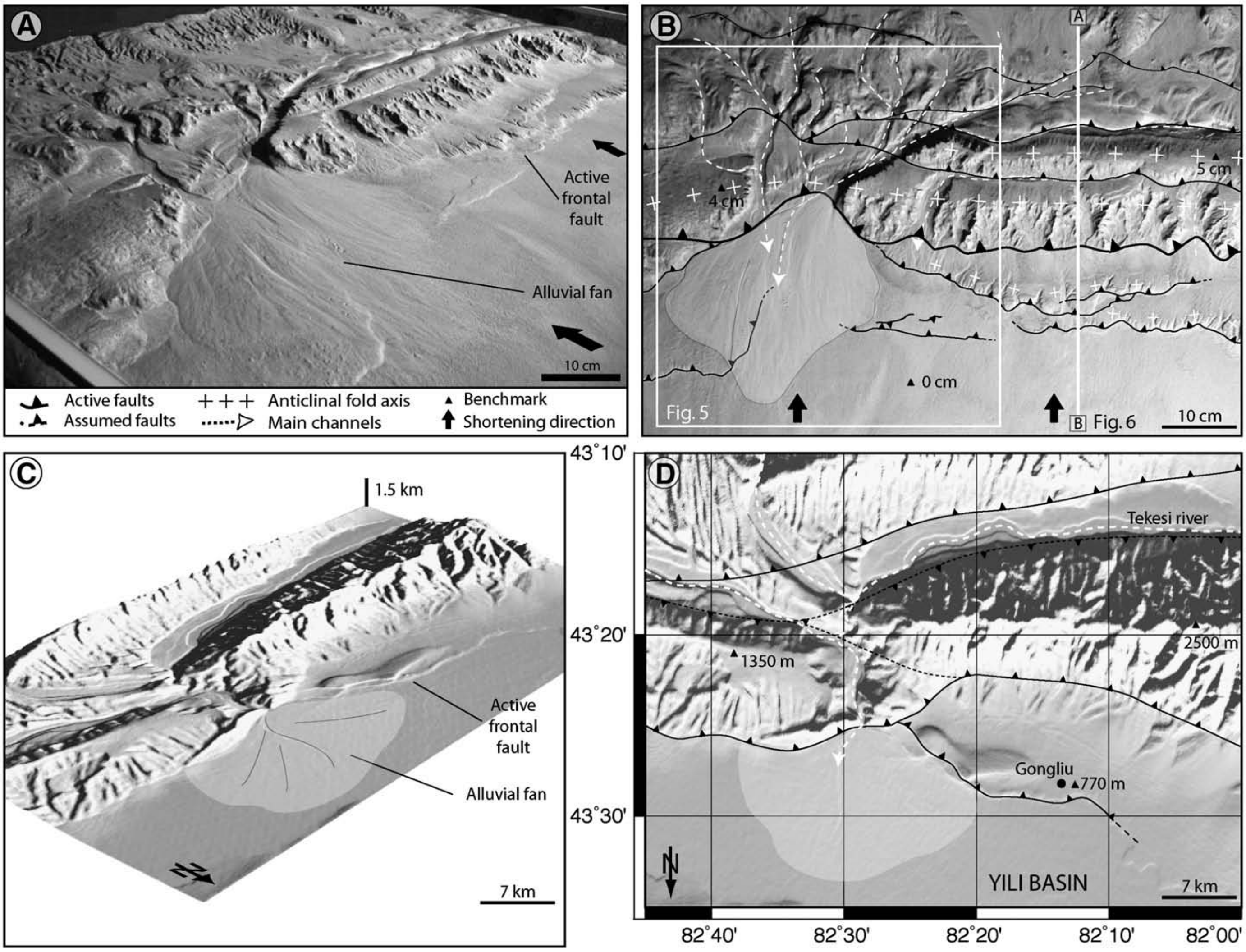

Fig. 7. Comparison of surface configuration of an experiment with structural convergence and rainfall erosion with a field case in Tian Shan, China. (A, B) Oblique and map views with structural sketch, of the experiment. (C, D) Oblique and map views with structural sketch of the Tekesi River flowing to the intramontane Yili basin. From Graveleau and Dominguez (2008).

appears to be beyond the reach of predictive models. In systems such as these where the configuration and behavior appear to depend in poorly understood ways on details of initial conditions, boundary conditions, and material properties, it is not surprising that description remains a major part of interpreting the results. But we must continue to work toward quantitative analysis and comparison with theoretical models as they develop.

3.2. Depositional systems and stratigraphy

In parallel to numerical landscape models, the last 30 years have seen dramatic growth in quantitative models for the evolution of sedimentary systems and the creation of the stratigraphic record (Tetzlaff and Harbaugh, 1989; Cross, 1990; Slingerland et al., 1994; Paola, 2000). This has proceeded in parallel with the development of sequence stratigraphy, a set of methods for partitioning and analyzing stratigraphic sections. Sequence stratigraphy was developed initially in the oil industry to relate stratal patterns to eustatic changes; specifically, it provided a way of analyzing large-scale geometric patterns imaged with seismic reflection in order to predict lithologic (reservoir) characteristics. As an analysis framework, sequence stratigraphy involves a combination of mapping contact geometry (e.g. onlap, offlap, etc.) and process interpretation of key surfaces and packages. The stratigraphic predictions then derive from these interpretations. Sequence stratigraphy and formal stratigraphic modeling have proceeded in parallel; in general, academia has led the way on modeling while industry has focused more on sequence stratigraphy. There has been less fruitful exchange between them than one would expect. Nonetheless, for our purposes the main point is that both approaches can provide testable predictions about preserved strata.

Although the field is the ultimate testing ground for stratigraphic models, there are some significant complications to field testing. Industrial model testing is mainly indirect, by drilling prospects worked up using input from sequence stratigraphy and/or quantitative stratigraphic models. Outcomes of specific plays are often proprietary, as are the details of the methods used to develop them. The outcome of a hydrocarbon play depends on so many additional factors that most cannot be considered to be tests of stratigraphic prediction. Outside of industry, the possibilities for predicting unknown characteristics of sedimentary rocks and then testing the prediction by observation are limited. Model evaluation mainly takes the weaker form of explanation of a known set of observations. In addition, given the expense of collecting high-quality seismic data, the academic community must often make do with outcrop observations, with their limitations in continuity and representativeness. Finally, in all field tests we are faced with the difficulty of independently constraining parameters and boundary conditions.

These reasons motivate the construction of systems by which stratigraphic models can be tested experimentally, under controlled conditions. This requires, at the very least, control of the “stratigraphic trinity” of external forcing recognized as primary controls from field studies: sediment supply, sea level, and subsidence. The first two of these are relatively easy to set up and control experimentally, but subsidence is quite a bit harder. In the field, subsidence varies strongly in space and time, so comprehensive stratigraphic experimentation requires a mechanism for spatially and temporally variable subsidence, independent of sea level and sediment supply.

3.2.1. Experimental methods

The facilities used by stratigraphic experimenters range in scale from very small tanks less than 1 m long and of the order of 0.01 m wide designed to produce 2D stratigraphic panels to large tanks and basins equipped with programmable subsiding floors. In this section we focus on the novel techniques required to include subsidence and ongoing net sedimentation into stratigraphic experiments.

The simplest way to produce net stratigraphic accumulation in a fluvial transport system is to raise base level. Rising base level is equivalent to spatially uniform subsidence with a fixed base level. Experiments using rising base level have been carried out for many years in engineering and earth sciences laboratories worldwide. The Leeds group (Moreton et al., 2002; Ashworth et al., 2004, 2007) introduced a new twist on this idea, inducing deposition by raising the experimental feed point. This creates a local slope excess near the feed point that then propagates the tendency to deposit through the system. Like rising base level, the rising feed point produces the equivalent of spatially constant subsidence.

The first system for producing experimental stratigraphy with controlled, spatially variable subsidence was developed at St Anthony Falls Laboratory beginning in 1996. The experimental stratigraphy basin is called the Experimental EarthScape (XES) system, though it is more often referred to as “Jurassic Tank”. In XES subsidence is produced by filling an experimental basin with granular material and extracting the material through a honeycomb of hexagonal cells in the basin floor (Paola, 2000; Paola et al., 2001) (Fig. 8). The internal friction of the granular material (pea gravel in this case) prevents uncontrolled release and allows the basement to support high lateral subsidence gradients, but its fluid-like properties smooth out the cell boundaries, producing a continuous basement surface. The pea gravel is extracted through the bottoms of the cells by precisely controlled fluid pulses that knock aliquots of gravel out of the cell base and draw the basement surface down. The pulses are continuously calibrated to produce about 0.1 mm of subsidence per pulse. A pulse pattern is fired about every 2 min, so that the maximum subsidence rate is several mm/h. The granular basement is covered with a flexible membrane that stretches and unfolds as the basement deforms. The experimental deposit is developed on top of this membrane by supplying the system with water and sediment from one or more input points and, if desired, manipulating base level.

Under the leadership of George Postma, the research group at Utrecht University has developed the “Eurotank”, which uses a

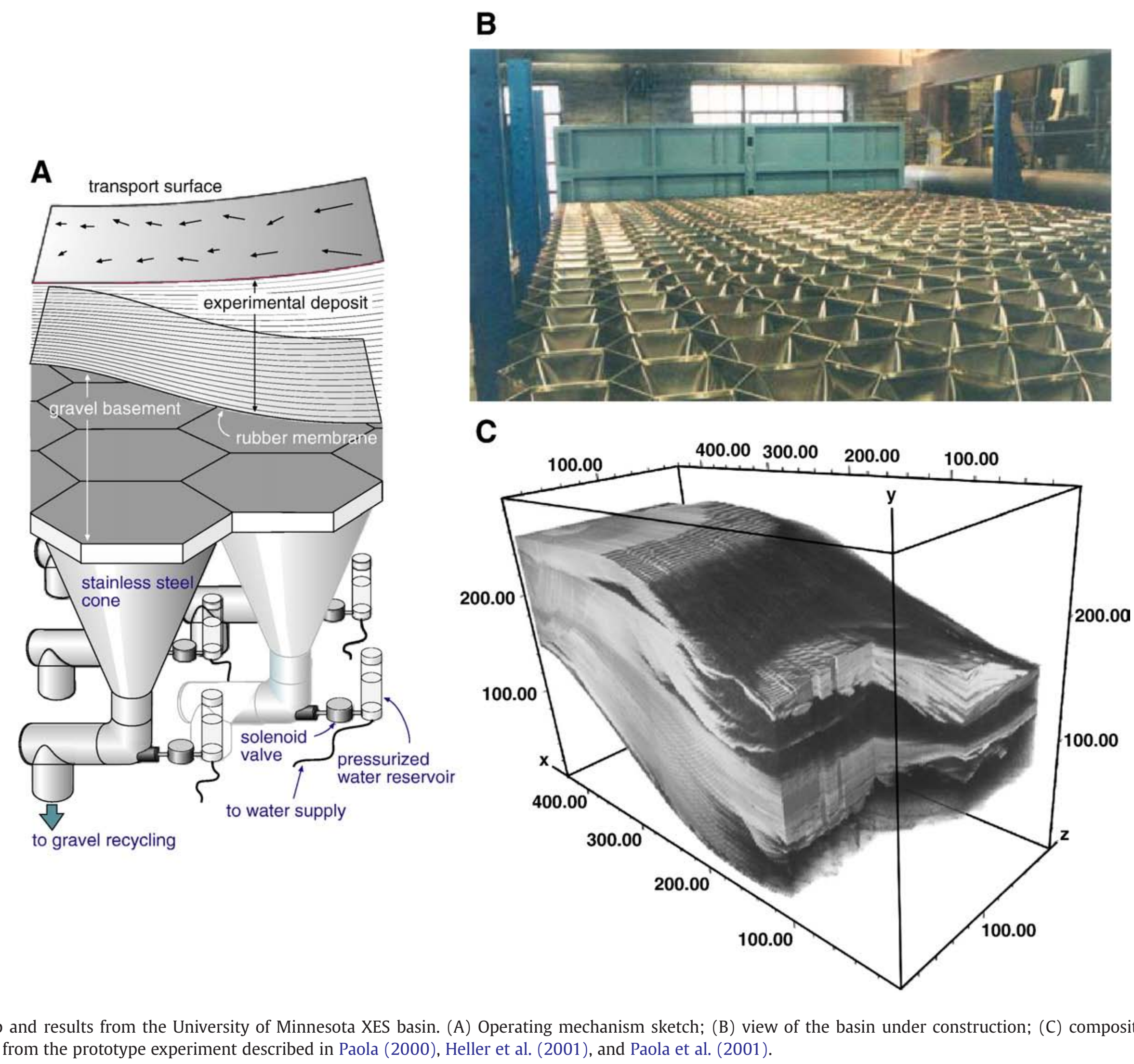

Fig. 8. Setup and results from the University of Minnesota XES basin. (A) Operating mechanism sketch; (B) view of the basin under construction; (C) composite 3D block of stratigraphy from the prototype experiment described in Paola (2000), Heller et al. (2001), and Paola et al. (2001).

honeycomb pattern comparable to that of XES but a different mechanism for vertical offset. The hexagonal units of the Eurotank floor are supported on screws operated by a small robot (Fig. 9). The robot/screw design has the advantage of allowing both uplift and subsidence to be produced directly, whereas in XES relative uplift must be produced by combining differential subsidence with falling base level. The drawback of the robot/screw system is that it is difficult to deform the basement smoothly as sedimentation is occurring.

In all other respects – sediment and water supply, surface imaging and measurement, etc. – subsiding-floor experiments are no different than other geomorphic experiments. Most XES experiments involve a binary mix of quartz sand ($D = 0.1$ mm) and anthracite sand that is more poorly sorted, ranging from 0.1–2 mm. The anthracite is lighter (though mostly coarser) than the quartz, making it the more mobile phase in the two-phase mixture (Fig. 8C). The XES system uses laser-sheet scanning for the subaerial surface and sonar for subaqueous surfaces. Eurotank uses photogrammetry everywhere, requiring that the experiment be drained to measure under water, but providing spatially continuous coverage.

A typical XES experiment involves of the order of 100 total run hours and takes 1–2 months to complete. Eurotank experiments are designed to be shorter and produce smaller stratigraphic sections, which allow more experiments to be run. In both systems, once the run is finished, the deposit is drained slowly, to avoid modification of the surface. Then it is sectioned, usually in the dip (downstream) or strike (cross-stream) direction. The sectioning involves exposing a vertical section of the deposit that is then imaged digitally. Spacing of the vertical faces varies; in XES experiments it has ranged from a few mm to 0.02 m. In XES, cutting is done using a wire cutter, similar to those used to slice cheese, aided by a computer-controlled 3D positioning system. Face imaging in XES uses a unique camera that creates a continuous image with a spatial resolution of about 0.2 mm by assembling small, undistorted individual images (Mullin and Ellis, 2008). The composite image can be as large as desired, has no detectable geometric distortion, is extremely color-faithful, and is insensitive to imperfections in the sediment surface. Collecting these images at relatively fine horizontal spacing produces a highly detailed 3D data cube of the experimental deposit. This can be analyzed using,

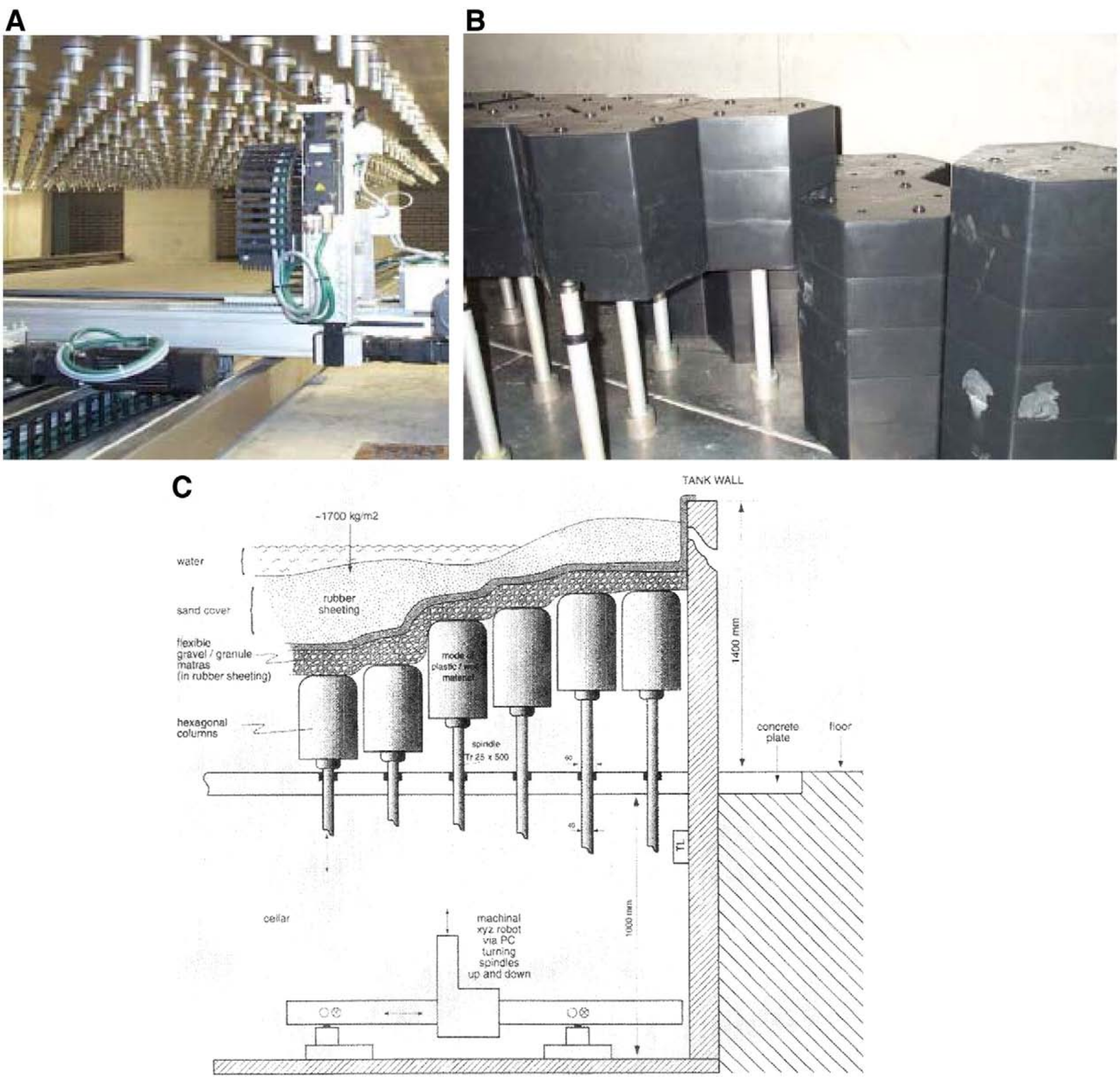

Fig. 9. Setup of the Utrecht University Eurotank basin. (A) Elevated, 0.32 m thick concrete tank floor pierced by 600 spindles that can be moved up and down by robot with a maximal throw of 0.20 m. Each spindle supports hexagonal plastic blocks each 0.10 in height that are mounted on top of each other to pre-shape the substrate relief (B). The substrate formed of the blocks is covered by two elastic rubber sheets, which enclose a layer of granules (C) under the experiment. Photos courtesy George Postma, University of Utrecht (NL).

for instance, 3D seismic software, and/or converted to a synthetic seismic image (Fig. 10).

In both the XES and Eurotank systems, the goal is to reproduce the kinematics but not the dynamics of subsidence. This works well for many purposes, but without feedback between subsidence and sedimentation, effects like isostatic subsidence due to sediment loading are excluded. McClay et al. (1998) studied clinoform sedimentation over a viscous substratum intended to simulate the presence of salt. The result included a substantial and realistic component of isostatic subsidence. This is an approach that deserves more attention — in particular, it would be exciting to combine this method with externally driven deformation along the lines of the XES or Eurotank systems.

3.2.2. Analytical methods

The combination of controlled conditions and the chance to observe evolution of an entire system at reduced time and space scales, makes experiments an ideal platform to develop new types of analyses. In general the main limitation in applying these to field situations is the availability of sufficient data. Even if the data requirements seem impossibly steep, the ability to measure something often brings with it new perspectives on how to think about system dynamics. Here we present a few examples of new forms of analysis that take advantage of the observational and analytical opportunities that experiments provide.

3.2.2.1. Geomorphic surfaces, stratigraphic surfaces, and chronostratigraphic significance. An important advantage of experimental stratigraphy is that it allows us to compare geomorphic (i.e. topographic) and stratigraphic surfaces quantitatively. We compare surfaces via what may be termed the *stratigraphic transformation* of the measured topographic surface η (x, y, t). The transformation has two steps (Strong and Paola, 2008; Martin et al., 2009). Step one is migration of topography to account for subsidence, leading to a new surface $\eta'(x, y, t)$:

$$\eta'(x,y,t) = \eta(x,y,t) - \int_t^{T_f} \sigma(x,y,\tilde{t})\, d\tilde{t} \tag{2}$$

where σ is the local subsidence rate (positive for subsidence, negative for uplift), $\tilde{t}$ is a dummy time variable, and T_f is the time when basin evolution is considered complete. Step two is clipping to account for erosion to produce a final stratigraphic surface η'' $(x, y, \hat{t})$. The clipping operation is defined with reference to the minimum migrated topographic elevation $z_{min} = \min$ $(\eta'$ $(x, y, \tilde{t}))$ for all times later than t, i.e. $t < \tilde{t} \le T_f$. The time of occurrence of the minimum is $\tilde{t}_{min}$. The transformed (migrated and clipped) stratigraphic elevation is defined as η'' $(x, y, \hat{t})$ where:

$$\eta'' = \eta' \text{ and } \hat{t} = t \text{ if } z_{min} \ge \eta' \tag{3a}$$

$$\eta'' = z_{min} \text{ and } \hat{t} = \tilde{t}_{min} \text{ if } z_{min} < \eta'. \tag{3b}$$

A mapable stratigraphic surface is a chronostratigraphic surface if two conditions are met: (1) it must coincide with a single transformed surface η'' $(x, y, \hat{t})$, and (2) that surface must satisfy condition (3a) everywhere over the mapped domain, i.e. it was never reworked. In that case, the stratigraphic surface has a unique age given by $\hat{t}$. More commonly a mapable surface coincides with a surface η'' $(x, y, \hat{t})$ for which $\hat{t}$ is not constant, in which case the surface has an age range given by the range of variation of $\hat{t}$.

3.2.2.2. Mass balance. Strong et al. (2005) analyzed the concept of depositional mass balance, measured via relative sediment extraction, as a fundamental variable controlling stratal architecture. The basic idea is that fractional sediment mass loss to deposition is a more general and useful way to measure proximality than downstream distance. They defined the dimensionless mass-extraction χ for a 1-D depositional system as:

$$\chi(x) = \frac{\int_0^x r(l)\,dl}{q_{s0}} \tag{4}$$

where x is downstream distance, r is rate of deposition, l a dummy variable representing streamwise distance, and q_{s0} is the volumetric sediment supply per unit basin width. Using vertical changes in alluvial architecture in an XES experiment, they found that recasting the data in terms of χ removed most of the change in channel stacking caused by a shoreline transgression due to an increase in updip fluvial slope.

Martin et al. (2009) proposed a quantitative measure of the commonly used idea of the depocenter via the depositional centroid. They used this to explore sediment mass balance by mapping the downstream migration of the centroid, showing that it follows the shoreline closely. Cross-stream centroid migration is controlled by the development and filling of incised valleys. Van Heijst et al. (2001) showed how the quantitative volumetrics of an experimental deposit could be compared directly with a well constrained field example; this work is discussed in the next section.

3.2.3. Stratigraphic effect of base-level cycles

Given the emphasis in classical sequence stratigraphy on eustatic sea level as the main control on sequence evolution, it is not surprising that the most thoroughly explored topic in experimental stratigraphy has been the response of depositional systems to base-level cycles. Most of the experimental work on this problem has used an initial platform-shaped surface and a fixed (nonsubsiding) floor. Early experimental work on sequence stratigraphy, done at Colorado State University using the Rainfall Erosion Facility discussed in Section 3.1, established that the major elements of the sequence model could be produced experimentally (Wood et al., 1993, 1994; Koss et al., 1994). This work is reviewed and summarized in Ethridge et al. (2005). In the CSU studies the sequences were produced by applying base-level cycles to a preformed step morphology representing a continental shelf-coastal plain. The substrate was weakly cohesive, so that a well defined incisional valley formed during base-level fall. The

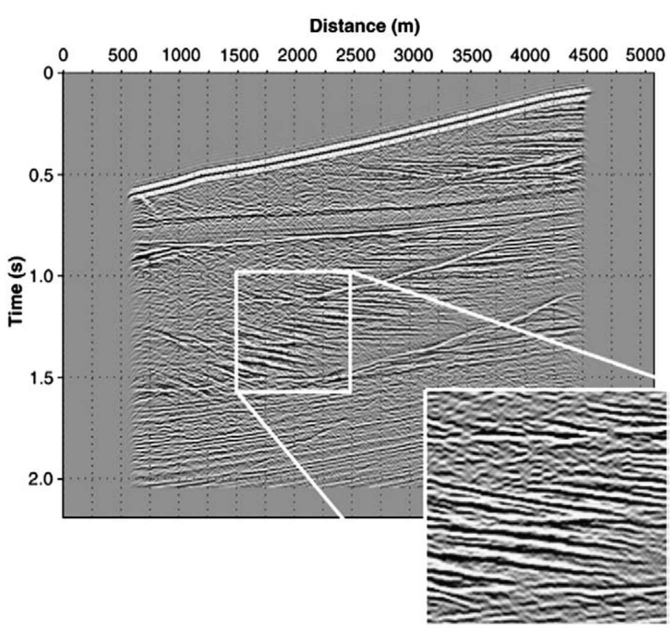

Fig. 10. Synthetic seismogram of part of an experimental dip section (Fig. 8C), produced by L. Pratson (Duke University) by one-way convolution of a source wavelet with seismic properties estimated from the experimental section based on local sand–coal ratio. Distance scales are arbitrary.

experiments included distributed rainfall as a water source, in contrast to most later work in which the water is supplied at one or more point sources, along with sediment. Using rainfall results in a more natural dendritic erosion pattern on the shelf surface exposed during base-level fall. Given the time needed for this erosion pattern to develop, it is not surprising that the experiments nicely illustrate the time lag associated with propagation of the erosional signal upstream. Another important finding, which deserves further work, is that small-amplitude base-level changes can be absorbed through internal adjustments to the channel system (e.g. changes in sinuosity) (Ethridge et al., 2005). Finally, Koss et al. (1994) reiterate an important point made by Posamentier et al. (1992) that we will return to later: the success of experiments like these shows that the main elements of sequence stratigraphy are scale independent over a wide range of scales.

Van Heijst et al. (2001) also studied the response of a fluvial-shelf system to sea-level cycles, emphasizing asymmetric cycles (slow fall, rapid rise). Their work showed that a key autogenic control on development of the shelf-edge deltas is the creation by incision of a complete connecting channel between the river system and shelf-edge delta. The creation of this channel by competition among channels headcutting across the emerging shelf establishes a "connection time" between the fluvial system and the shelf edge. The connection time then sets the time lag between base-level fall and the initiation of incision in the fluvial system updip; until connection, the fluvial system continues to aggrade even though the sea level is falling. The faster the eustatic fall rate, the shorter the connection time. The connection time in turn determines the relative amount of reworked shelf sediment versus fluvial sediment delivered to the shelf edge, since only reworked sediment is available until the connection is established. Based on a comparison with observed knickpoint migration rates in depositional systems, Van Heijst et al. (2001) estimate connection times in large river–shelf systems (e.g. the Mississippi) to be in the range of a few thousand years. The duration of the eustatic fall also determined the total volume of slope fan and lowstand delta deposits, with the volume increasing with increasing duration. The complete system evolution during a single representative cycle is summarized in Fig. 11. It is also worth noting the parallel between the connection-time idea and the knickpoint-controlled time scales discussed by Allen (2008).

Van Heijst et al. (2001) have carried out one of the most thorough quantitative comparisons to date between a stratigraphic experiment and a relatively well documented field case. Here the target was depositional units in and offshore of the Colorado River Delta, and the scaling was carried out in terms of the kinds of large-scale dimensionless parameters discussed in Section 2.2. A quantitative comparison is shown in terms of sediment volume partitioning in Fig. 12. While not perfect, the agreement in terms of scaled sedimentation volumes is surprisingly good, indicating that the large-scale dynamics that control the volumetrics of sediment redistribution during base-level cycles can be reasonably well captured in experiments such as these in which the fine-scale dynamics are unscaled.

Basin-scale patterns of stratigraphic response to base-level change, emphasizing steady rise and fall, have been the focus of a series of experiments carried out at Nagasaki University under the leadership of Tetsuji Muto. This work is remarkable for the simplicity and

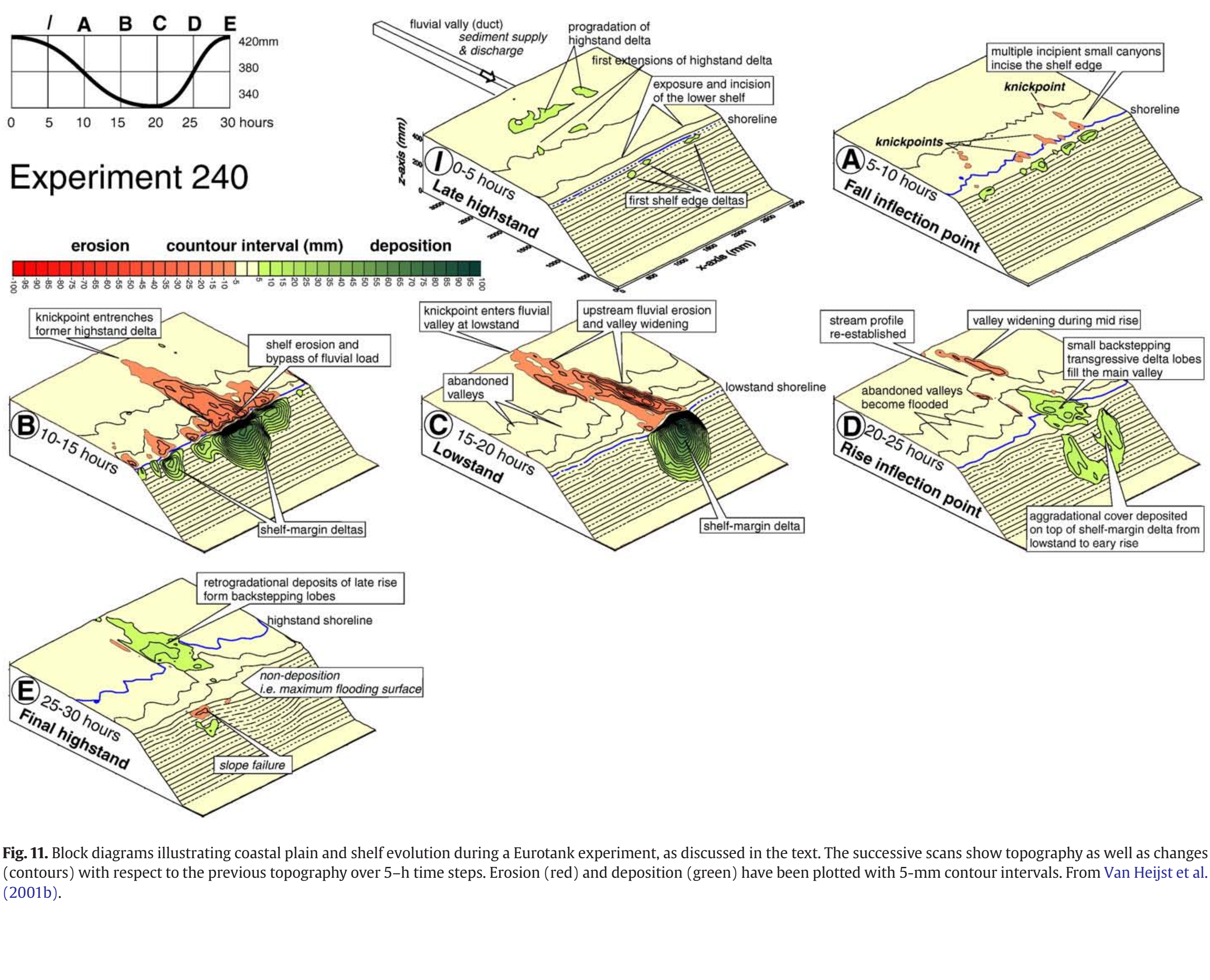

Fig. 11. Block diagrams illustrating coastal plain and shelf evolution during a Eurotank experiment, as discussed in the text. The successive scans show topography as well as changes (contours) with respect to the previous topography over 5–h time steps. Erosion (red) and deposition (green) have been plotted with 5-mm contour intervals. From Van Heijst et al. (2001b).

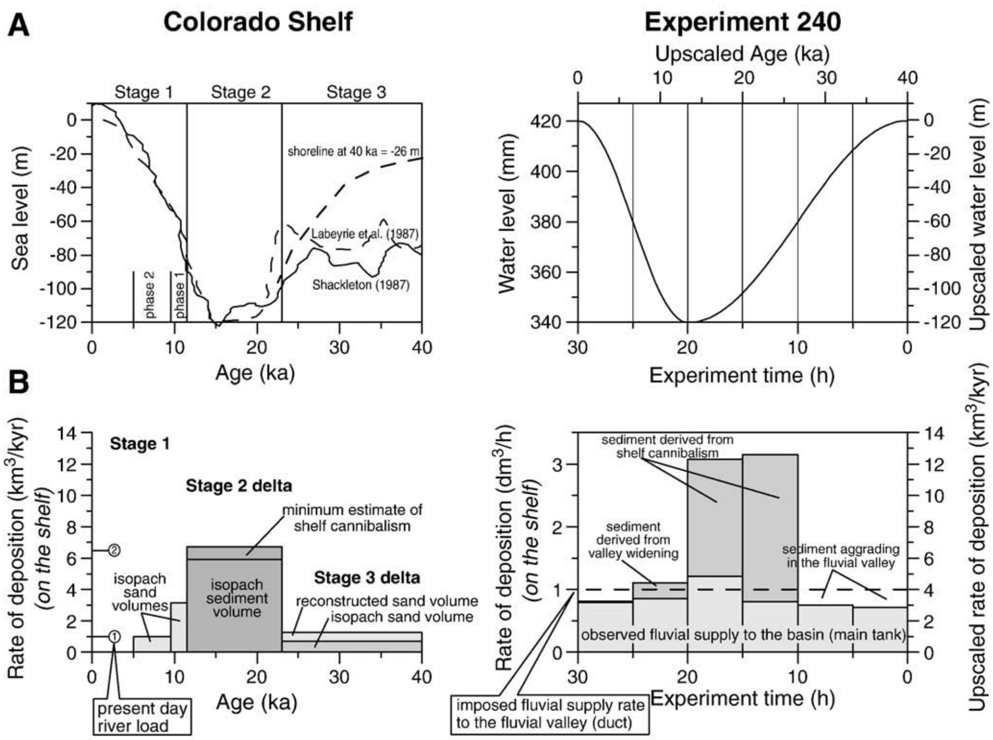

Fig. 12. Results of a quantitative comparison between an experiment with variation in base level (right) and the evolution of the Colorado River–shelf system (left). (A) Sea-level curves. (B) Scaled rate of deposition on the shelf and slope versus scaled time. The dashed line in the right graph indicates the imposed fluvial sediment supply rate during the experiment. The observed fluvial supply to the tank is represented by the light gray shading. Details of the scaling conversions are given in the source: Van Heijst et al. (2001b).

elegance of the experimental design, which serves to highlight all the more strongly the complexity of the stratigraphic results. Muto and his colleagues have run two types of experiments, one in a very narrow tank that eliminates lateral variability, and the other in a wider one that allows channelization, avulsion, and lobe formation. The general theme of the experiments is autogenic dynamics (“autostratigraphy”: (Muto et al., 2007)). Muto and his colleagues use a somewhat broader definition of autogenic dynamics than many researchers. The fundamental step they have taken is to expand the concept of autogenic dynamics from variability unrelated to external forcing to nonlinear system response to steady allogenic forcing. One fundamental point emerging from this work is that even the simplest clinoform building into an unconfined space cannot reach a true steady state despite constant subsidence or base-level rise (Muto, 2001). The foreset must continue to lengthen with time, requiring continual shift in the partitioning of deposition from topset to foreset. This result is one example of how clinoform systems subject to steady forcing can respond in interesting and non-intuitive ways. A second example is that, depending on the initial configuration and sediment supply, the shoreline can prograde during the initial stages of sea-level rise (Muto, 2001). Likewise the fluvial system can continue to aggrade during sea-level fall; in fact, depending on rate of fall, system geometry, and sediment supply, different clinoforms could experience and record the same sea-level fall in very different ways (Swenson and Muto, 2007). Fluvial aggradation during eustatic fall never gives way to incision if the fluvial transport slope is higher than the basin floor slope (Petter and Muto, 2008). Finally, the classical idea of fluvial grade in the sense of Mackin (1948) reappears in an interesting and unexpected new form in the analysis of fluvial response to eustatic fall: the fluvial system on an isolated clinoform reaches grade when the sea level falls as $t^{1/2}$. Even more unintuitively, grade is reached for *any* constant rate of fall for the common case of clinoforms prograding over previously deposited clinoform surfaces with the same surface slope (Muto and Swenson, 2005).

The Nagasaki group's work has also shown transitions in the qualitative form of stratigraphic response to base-level change. One example is termed the “autobreak” by Muto and Steel (2001) (Fig. 13). Autobreak represents a transition from mixed onshore–offshore deposition to purely onshore deposition in a simple clinoform during steady base-level rise. The stratal pattern and mode of shoreline migration change at the autobreak limit (Fig. 13). As illustrated in Fig. 13, the complete evolution of the clinoform starting from a bare tank comprises three stages with qualitatively different modes of shoreline evolution; it would be easy to misinterpret these changes as having an external cause. In the complementary case of steadily falling sea level, starting again from a bare tank, the onset of incision (“autoincision”) is delayed by a time interval that depends on system geometry, rate of fall, and sediment supply (Fig. 14). For the same starting conditions, if the fluvial slope is steeper than the flume-bed (platform slope in nature), the fluvial system aggrades during steady eustatic fall, and the shoreline eventually detaches from the alluvial prism (“autogenic detachment”; Fig. 15) (Petter and Muto, 2008). These changes in system behavior can be thought of as, in effect, phase transitions in the depositional system, analogous to the transformation of liquid water to ice during steady cooling. The essential point is that all of these behaviors and transitions could easily be interpreted as being the results of a change in allogenic conditions when in fact they are due to changes in internal response to steady forcing.

The Nagasaki group's work over the past decade represents a major step in our understanding of how stratigraphic systems can respond to and record external forcing in complex and counterintuitive ways. This idea, which recurs throughout this section, is the depositional analog of Schumm's “complex response” concept in which geomorphic systems respond in complex ways to simple external changes (Muto and Steel, 2004). One especially powerful and appealing aspect of the Nagasaki group's research has been the playing off of results from narrow versus wide experimental tanks. Narrowing the width suppresses stochastic autogenic variability, providing a clear view of

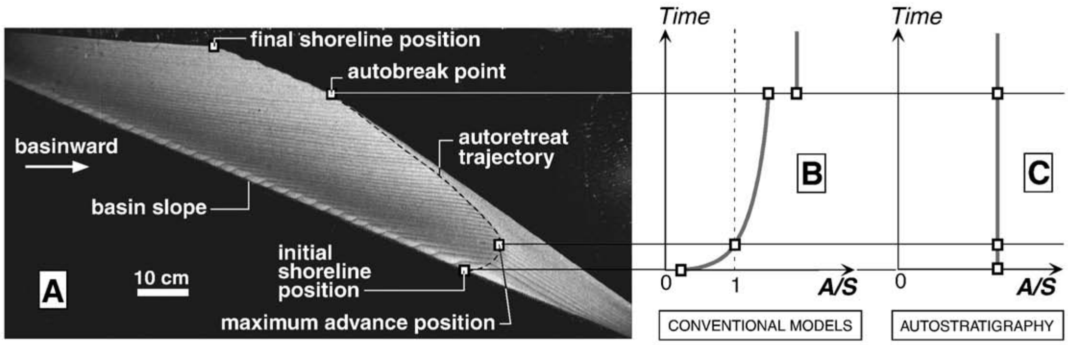

Fig. 13. (A) Evolution of shoreline and clinoform geometry in a 2D tank experiment under conditions of steady sediment supply and steadily rising base level, starting from a bare floor, showing autobreak. (B) shows how the observed variation in the direction of shoreline migration and partitioning of sediment supply between fluvial and offshore deposits could be misinterpreted if shoreline migration is linked directly to changes in the accommodation/supply (A/S) ratio. The correct (steady forcing) interpretation is shown in (C). From Muto et al. (2007).

the strongly nonlinear ways in which simple clinoforms can respond to steady forcing. Relaxing the width constraint adds the stochastic autogenic variation back in and illustrates how the deterministic and stochastic aspects of the system interact. Both the deterministic and stochastic dimensions of clinoform dynamics revealed by this research occur during steady base-level rise or fall. Thus Muto and his colleagues advocate expanding the definition of "autogenic" to include the deterministic, nonlinear dynamics they have described. In our view, this becomes problematic when, as is typically the case, the base-level and other external changes are not steady. Here we will retain a more traditional view and use "autogenic" for internally generated processes occurring on time and length scales that are clearly separated from those of the external forcing, as distinct from complex, nonlinear responses to the forcing. The contributions of Muto and colleagues to autogenic processes as defined this way are discussed in Section 3.2.7.

3.2.3.1. Base-level changes combined with subsidence. The XES 96 experiment (Paola, 2000) was to our knowledge the first to include subsidence as a separate, spatially variable contribution to total accommodation. The experiment was carried out in a prototype version of the XES basin that measured 1.3 m long and 1.0 m wide and had 10 subsidence cells. Two cycles of base-level change were applied, one slow relative to the characteristic equilibrium time for the basin (T_{eq} as defined in Paola et al. (1992a)), and the other rapid, i.e. with period $T_i < T_{eq}$. The system response to the two cycles was significantly

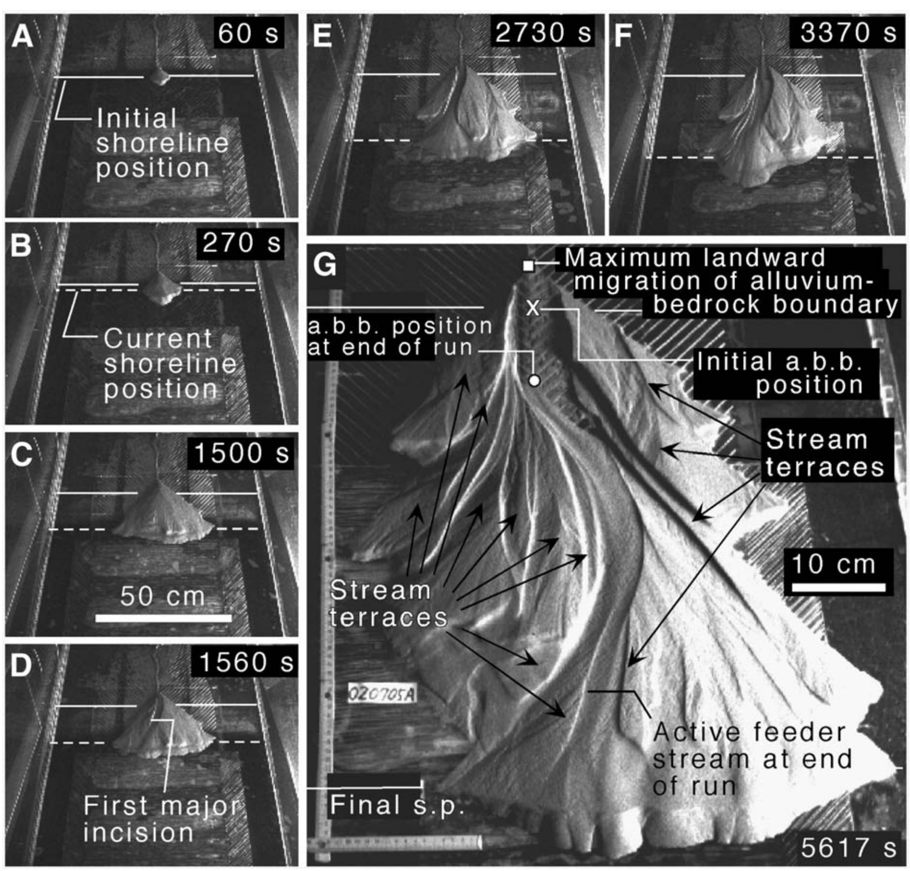

Fig. 14. A fluvial delta evolving under conditions of steady water and sediment supplies, and steady base-level fall. Time from the beginning of the run is indicated in successive images (A–G). The system initially aggrades (A–C). Incision begins at (D), and continues (E–G), producing numerous autogenic terraces and downdip depositional lobes, despite the steady forcing. a.b.b. = alluvium–bedrock boundary. From Muto and Steel (2004).

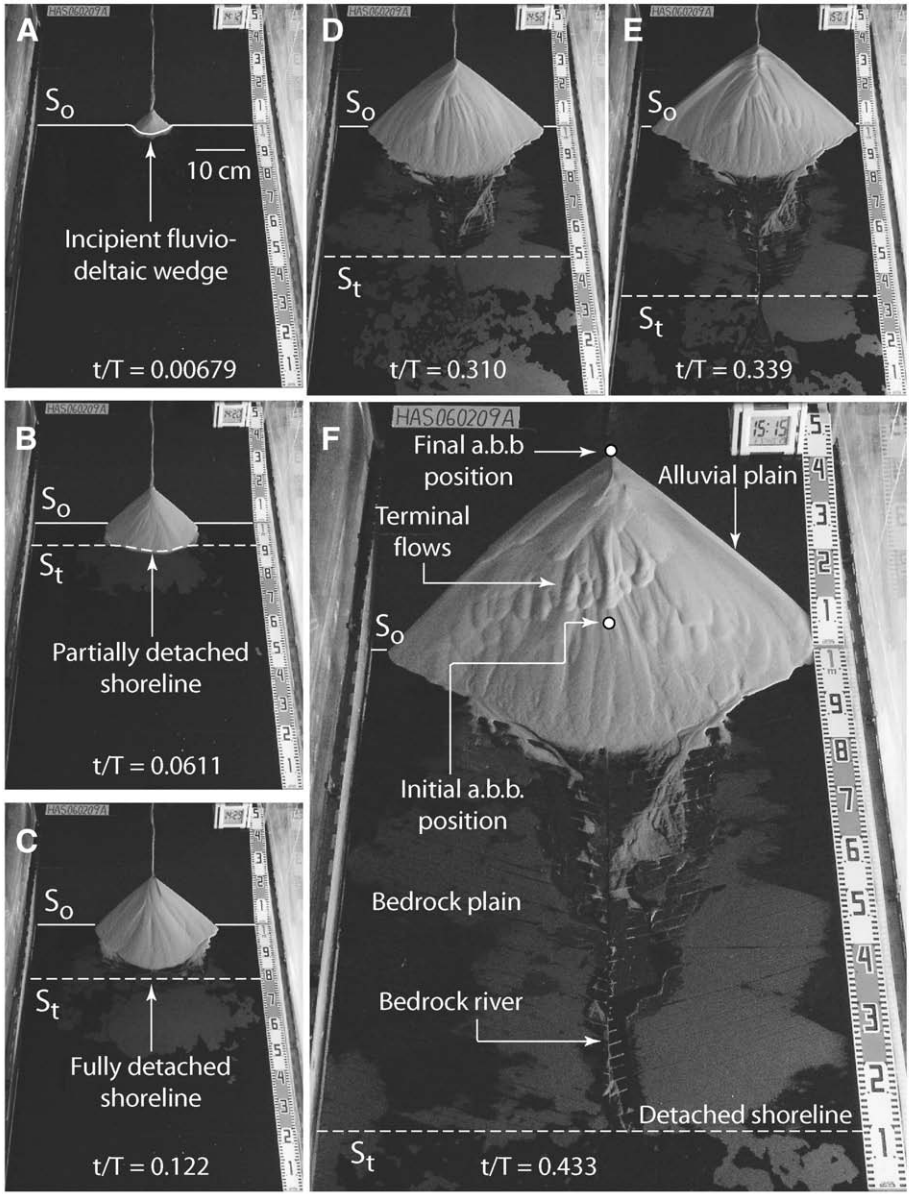

Fig. 15. A case of ongoing fluvial deposition throughout an interval of steady base-level fall. The slope of the flume bed must be greater than the slope of the fluvial surface for this to occur. Shoreline detachment (auto-detachment) is complete in image (C). Time from beginning of run normalized to total run time is indicated in the upper right of each image. S_0 and S_t are positions of the initial shoreline and shoreline at time of the image, respectively. Positions marked with black circles mark the alluvium–bedrock boundary at the upstream termination of the alluvial plain. From Petter and Muto (2008).

different: the rapid cycle produced a well defined incised valley, though it eventually widened to occupy most of the basin width, while the slow cycle produced only a broad erosional surface and a low-angle unconformity surface. During the slow cycle, the fluvial system was able to migrate laterally fast enough relative to the rate of base-level fall to avoid localization and valley formation. In addition, the rapid cycle produced a series of growth faults that extracted sand from the near-shore, creating a series of isolated sand bodies similar to growth-fault associated oil reservoirs (Fig. 8).

Part of the 1996 experimental design was to investigate the model of shoreline response to sea-level cycles proposed by Pitman (1978). Pitman's most interesting finding was that for long-period base-level cycles (i.e. $T>T_{eq}$; note that Pitman used a different but equivalent form of T_{eq}), the shoreline response is phase-shifted relative to the base-level cycles by $\pi/2$, such that the time of maximum transgression coincides with the time of maximum rate of rise, rather than with sea-level highstand. No evidence of this phase lag was found in the experimental data (Paola, 2000; Heller et al., 2001; Paola et al., 2001). In their analysis of shoreline response to sea-level cycles, Swenson et al. (2000) found that the predicted phase lag was an artifact of a key model assumption, that the fluvial system is everywhere erosional.

A larger-scale experiment XES experiment on response to base-level cycles was carried out in 2002 (Kim et al., 2006a,b; Strong and Paola, 2006; Giosan et al., 2008; Strong and Paola, 2008; Martin et al., 2009). Apart from involving a much larger experimental system (3 m by 6 m, with 104 subsidence cells) the other major improvement was a second stage with superimposed base-level cycles. The slow cycle in this experiment lasted for 108 h and the rapid cycle 18 h. Cycle amplitude was set to 110 mm, or about 3 times the maximum channel-scour depth. For comparison, for a glacioeustatic cycle amplitude of 120 m, the equivalent channel depth would be 30 m, comparable to the maximum depth of the modern Mississippi. The second stage involved superposing six of the rapid cycles onto the slow cycle.

The overall stratigraphic record of this experiment is shown in dip section in Fig. 16 and is analyzed in sequence-stratigraphic terms in

Martin et al. (2009). As in the much smaller-scale XES 96 experiment, the slow cycle record is markedly more symmetric than that of the isolated rapid cycle. Its sequence boundary, created via a broad, flat erosion surface with no clear incised valley, is less distinct than the valley-derived sequence boundary of the rapid cycle. The sequence boundary is extremely time transgressive; we discuss this and other sequence-stratigraphic results in the next section.

The record of the six rapid cycles superimposed on the slow cycle shows the strong effect of the slow cycle in determining how the rapid cycles were recorded stratigraphically (Martin et al., 2009). For the three cycles on the falling limb of the slow cycle, each successive erosion surface cuts (on average) deeper than the one before, so that the fluvial records of the three cycles are largely obliterated. What survives, the chance remnants of autogenic dynamics, is so highly amalgamated as to be unrecognizable. The sequence boundary in this case is a "super boundary" developed over the entire three-cycle set, and thus even more time-transgressive than that of the isolated rapid base-level cycle. In fact, the entire three-cycle package appears to represent a single cycle and could easily be interpreted as such. The only means of recognizing the presence of superimposed cycles within the falling slow cycle limb turns out to be marine onlap surfaces, though these can be produced by other means so are not uniquely diagnostic (Martin et al., 2009).

Fig. 16. Uninterpreted dip panel from the XES 02 variable base-level run discussed in the text. Insets show (lower left) detail of updip termination of a fluvial erosion surface, lost in autogenic "noise" as its amplitude diminishes updip, and (upper right) marine downlap and onlap surfaces associated with lateral variation in the location of the incised valley and offshore depositional lobes. Interpretations of this and similar panels are provided in Martin et al. (2009).

An important mass-balance question is the extent to which eustatic cycles produce net offshore transfer of sediment. Given that the processes involved in the depositional-system response to eustatic cycles are highly nonlinear, it is not clear to what extent offshore sediment transfer during relative fall is balanced by onshore sediment trapping during relative rise, over a complete cycle. Kim et al. (2009) introduced the idea of the "eustatic pump" to describe net sediment transfer offshore during eustatic cycles. They used a numerical model calibrated to the results of the XES 02 experiment to show that simple, symmetric eustatic cycles produce at best weak eustatic pumping. Base-level cycles with basin subsidence cause substantially greater net pumping in backtilted basins than in foretilted ones. When sediment supply varies over a base-level cycle, pumping is maximized when the sediment supply maximum occurs during falling stage or lowstand, consistent with the findings of Perlmutter et al. (1998).

A final major issue in the experimental study of basin response to eustatic variation is the age of key surfaces. In the dip direction, Van Heijst et al. (2001) showed that extent of time transgression of the sequence boundary increases for shorter cycle times, so that the age of the unconformity updip correlates with the high stand for sufficiently rapid cycles. Where subsidence is involved, the analytical framework for surface dating discussed above comes into play, allowing us to compare age and time transgression for key sequence stratigraphic surfaces. The sequence boundary typically develops over nearly the entire base-level cycle, especially if the cycles are rapid (Strong and Paola, 2006, 2008; Martin et al., 2009). The time transgression includes both stochastic and systematic components. The former are associated with autogenic dynamics, which we discuss in Section 3.2.7. In strike section, the sequence boundary has the form of a valley, and the edges are generally much younger than the valley center (Strong and Paola, 2006, 2008). Other key sequence-stratigraphic surfaces develop in experiments as well. The transgressive surface of erosion (TSE) is identifiable as in rapid cycles, and the maximum flooding surface (MFS) is well defined in both rapid and slow cycles (Martin et al., 2009) Transgression in the slow cycle involves landward migration of the shoreline accompanied by ongoing deposition, so there is no TSE developed in that case.

3.2.4. Stratigraphic effects of water and sediment supply cycles

Stratigraphic patterns comparable to those produced by eustatic variation can be produced by changes in other forcing parameters, like sediment and water supply. An excellent example is provided by the experiments of Milana and Tietze (2002) on the effects of changes in water supply on alluvial fans. The stratal geometries produced by cycles in water supply have much in common with stratigraphic sequences produced by changes in relative sea level. In particular, increases in water supply produce regional erosional surfaces that, in addition to truncating underlying units, resemble sequence boundaries in that they mark a basinward shift in facies. The sequences could easily be mistaken for base-level driven sequences in the absence of information on dip-section geometry. The situation becomes more complex when the effects of upstream-driven changes in water supply are superimposed on downstream-driven changes in base level (Milana and Tietze, 2007). In these experiments water supply and base level were varied, with constant sediment supply. Depending on the phase relationship between the water supply and base-level cycles, regional unconformities propagating toward the interior from both updip and downdip boundaries are produced, but at different times during the cycle. The major sequence-bounding unconformities in this case are neither time-correlative nor do they separate everywhere older from everywhere younger deposits (Milana and Tietze, 2007). These experimental results are complementary to the theoretical predictions of Perlmutter et al. (1998) on the effects of phase-shifted changes in sediment supply during glacioeustatic sea-level cycles.

3.2.5. Tectonics and sedimentation

The primary motivation for building experimental systems with deformable floors is to investigate the interplay of surface dynamics and imposed tectonic forcing. The key scientific questions arise from the interaction of self-organization of the surface transport systems and the deformation pattern associated with the tectonics. For example, the XES 99 experiment examined the effect of lateral variation in subsidence rate on alluvial channel dynamics and stratigraphic stacking. This was inspired by the model proposed by Alexander and Leeder (1987) in which lateral tilting attracts channels to subsidence maxima. This effect would lead to high cross-sectional areal density (stacking density) of channel deposits in the subsurface over subsidence maxima — for example, on the down-thrown side of normal faults. Hickson et al. (2005), in an analysis of the effects of changing subsidence rate on alluvial architecture using the XES 99 experiment, did not find this to be the case. A lateral doubling of the subsidence rate produced the expected thickening of strata into the maximum, but the areal density of channels in strike section was unaffected by the higher subsidence rate. The explanation appears to lie in a dimensionless time-scale ratio that provides a way of estimating the effect of lateral tilting on channel dynamics (Kim et al., 2009). The basic idea is that lateral tilting influences channel pattern only if it can produce a significant lateral slope faster than migrating channels can redistribute sediment and smooth it out. The time-scale ratio involves a tectonic time scale T_t given by:

$$T_\mathrm{t} = \frac{S_x L_y}{\Delta\sigma} \tag{5}$$

This time scale measures the time required for a change in subsidence rate $\Delta\sigma$ developed over a lateral distance L_y to create a slope equal to the downstream surface slope S_x. We compare this with an estimate of the time T_ch needed for flow to occupy the dry width of the basin:

$$T_\mathrm{ch} = \frac{B_\mathrm{T} - \Sigma b}{v_\mathrm{ch}}. \tag{6}$$

Here B_T is the total basin width, Σb the summed width of all channels, and v_ch a characteristic rate of lateral channel migration. At this point the physical controls on v_ch, which includes both continuous migration and avulsion, are not known. A starting point for analysis is that it scales as q_s/h where h is the channel depth and q_s is the average volumetric unit sediment flux. The dimensionless time-scale ratio T_* is then given by $T_* = T_\mathrm{t}/T_\mathrm{ch}$ Tectonic effects prevail when the tectonic time scale is fast (short) relative to the channel time scale, i.e. $T_* \rightarrow 0$.

The XES 05 experiment (Kim and Paola, 2007) was designed to test the time-scale argument. It had a more complex tectonic forcing geometry aimed at capturing the essence of a relay-ramp style normal

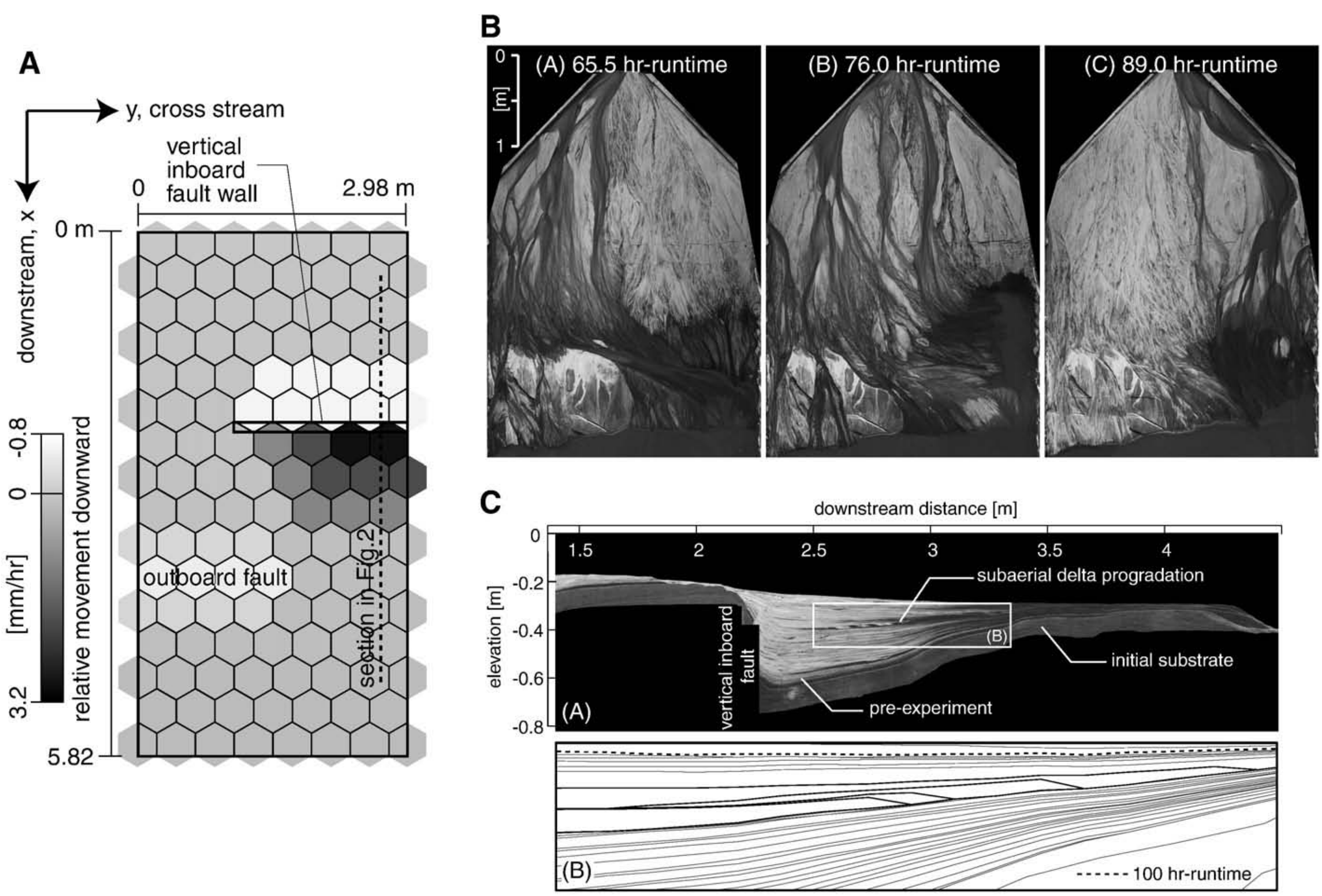

Fig. 17. (A) Pattern of relative subsidence (dark) and uplift (light) imposed in the XES 05 relay-ramp experiment. (B) Surface images of XES 05 taken during the run showing change in relative deflection associated with tectonic deformation: from left to right, channels go from being deflected around the uplift, creating an autogenic lake in the area of maximum subsidence, to crossing the uplift and filling the lake in. (C) Dip-section stratigraphy image and sketch across the vertical fault showing the autogenic lake deposits (dark interval in image) with fluvial strata above and below. From Kim and Paola (2007).

fault geometry. (Fig. 17). The geometry shown in Fig. 17 includes two regions of uplift. Since the XES system relies on substrate extraction to produce basement deformation, uplift is produced via continuous base-level fall so that cells that do not subside experience relative uplift. To produce a tectonic-dominated condition, the experimental design called for making T_* as small as possible. The tectonic time scale T_t is externally imposed, but T_{ch} can be influenced only indirectly — generally, the idea is to slow the channels down by keeping sediment feed rate low and depth high. Minimizing the sediment supply while maximizing the lateral tilt rate reduced T_* relative to the XES 99 experiment, with the results shown in Fig. 17. The imposed basement deformation clearly influenced the channel pattern, causing flow to deviate around the uplifts and be drawn into the subsidence maximum. The extent to which flow was deflected, however, varied autocyclically over the course of the experiment, producing alternation of lacustrine and fluvial deposition despite steady forcing. As discussed in Kim and Paola (2007), this effect extrapolated to field conditions could give rise to autogenic variation with a time scale of the order of 10^4 yr or more and produce stratal patterns that would likely be interpreted as allogenic.

3.2.6. Avulsion and architecture

We introduced this subject in the previous section, where we saw that experimental testing of proposed relations between tectonics and alluvial architecture has yielded new insights about how channel and tectonic time scales interact. Here we look at work specifically focused on avulsion (Slingerland and Smith, 2004) and depositional architecture. This line of research begins with the landmark paper of Leeder (1978), who first proposed a quantitative relation among avulsion, sedimentation, and channel stacking in the subsurface. The first experimental study of the influence of sedimentation rate on avulsion frequency was that of Bryant et al. (1995), who showed for an experimental alluvial fan that avulsion frequency could increase faster than linearly with sedimentation rate. All else equal, this would imply a relation between sedimentation rate and stacking density opposite to that originally proposed by Leeder (1978). A weaker but still positive relation between avulsion frequency and sedimentation rate was found by Ashworth et al. (2004), who used the University of Leeds rising-feed flume and a more conservative definition of avulsion. The important result arising from both studies is experimental confirmation, in two different systems, of the strong dependence of avulsion frequency on sedimentation rate. Further work on avulsion frequency and topographic evolution by Ashworth et al. (2007) showed that avulsion frequency did not change with downstream position and that spatial variation in flow-occupation frequency did not correspond to changes in net sedimentation rate. The latter result reinforced the findings of Sheets et al. (2002) that flow occupation is decoupled from net sedimentation rate; both studies point to a dominant role in deposition of short-lived depositional events. In braided river and fan experiments, these events are typically local in-channel flow expansions; in low-gradient field settings the equivalent events appear to be crevasse splays lateral to the main channel. In a third study using the Leeds facility, Moreton et al. (2002) studied the relation between surface dynamics and depositional architecture in an experiment Froude-scaled to a coarse-grained depositional system in South Island, New Zealand. The experiment produced highly structured deposits with roughly three orders of magnitude variability in expected hydraulic conductivity. The experiment reproduced the main facies and deposit geometries observed in the field (Fig. 18).

In the XES 99 experiment designed to compare architecture under different spatial subsidence patterns and rates, Hickson et al. (2005) and Strong et al. (2005) found that reduction in subsidence rate did not lead to a significant change in channel-stacking density, as the Leeder (1978) theory had suggested, once a quantitative correction was made for changes in the overall depositional mass balance as the subsidence rate declined. The mass-balance correction was the one introduced in Section 3.2.2. As pointed out in Bryant et al. (1995) and Heller and Paola (1996), the local rate of increase in avulsion frequency with sedimentation rate controls the relation between channel-stacking density and sedimentation rate. If the increase is linear then channel-stacking density is unaffected by changes in sedimentation rate. Taken as a whole, the experimental results

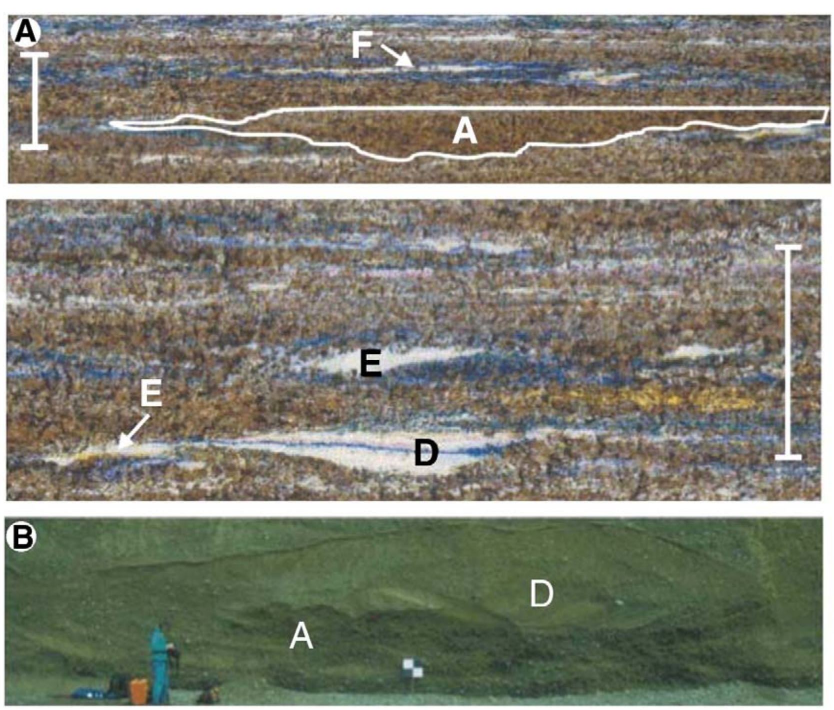

Fig. 18. (A) Strike panels from an experiment on alluvial architecture in a braided stream undergoing steady aggradation compared with (B) field example from the Canterbury Plains, South Island, New Zealand. Labelled facies types: A = primary channel fills, D = fine-grained channel fills, E = erosional remnants, F = floodplain fines. Scale indicators in (A) are 20 mm, in (B) 0.5 m. From Moreton et al. (2002).

support Leeder's (1978) original proposal that decreases in subsidence rate are associated with increased channel-stacking density, but through a different mechanism than the one he envisioned: decreasing subsidence rate, holding other controls constant, leads to facies progradation. Proximal fluvial facies, with higher ratios of transport to net deposition, generally have greater (denser) channel stacking than distal facies, so the progradation leads to increased stacking density. At present we do not know to what extent these experimental observations apply at field scales. The experimental systems do not have flood plains and are braided rather than single-thread rivers; it remains to be determined how these differences affect avulsion frequency and spatial pattern. At a minimum, the experimental work offers a new way of thinking about controls on channel-stacking density; now it seems to be time to return the spotlight to the field. The hypothesis is that the "Leeder effect" does occur, but is associated with downstream facies migration rather than increased local reworking at constant avulsion frequency.

3.2.7. Other autogenic processes

Avulsion is only one of the autogenic behaviors of channelized flow systems. Small-scale autogenic processes, associated with bar and channel dynamics, are primarily stochastic and can be thought of crudely as a kind of noise superimposed on the large-scale, allogenic stratal patterns (e.g. Fig. 16). On the other hand, experimentation has revealed the presence of long-period autogenic change whose stratigraphic effects that would most likely be interpreted as allogenic (Muto et al., 2007). For example, Kim and Paola (2007) observed the development of autogenic alternation between lake and fluvial deposition in the XES 05 experiment introduced above. The experiment involved relative uplift and subsidence in a simplified relay-ramp configuration, so one question was whether the channels would cross the uplift or be diverted around it — one of the transverse-channel problems investigated by Douglass and Schmeeckle (2007). In fact, the system did both, alternating between a lacustrine mode where flow was diverted around the uplift and an all-fluvial mode where it crossed the uplift. Using mean channel-scour depth h_{ch} as a length scale and h_{ch}/σ as a time scale allows extrapolation of the experimentally measured period for creation and filling of the lake to field scales. For example, a field channel depth of 5 m and a subsidence rate σ of 1 mm/yr give a scaled period of roughly 10^5 yr and a stratal thickness of 50–100 m for one cycle. These values are comparable to numerical estimates of the time scale for erosion cycles in relay-ramp systems (Allen and Densmore, 2000), and longer than the time span encompassed by most of the fluvial sedimentation cycles measured in the Loreto Basin by Dorsey et al. (1995).

Migration of alluvial channels is a classic form of autogenic behavior — indeed, the recognition of autogenic effects in stratigraphy started with Beerbower's (1964) explanation of fluvial fining upward sequences via what he termed "autocycles" of meander bend cutoff and growth. As discussed in the previous section, a general idea emerging from experimental studies is the existence of a characteristic channel time scale T_{ch}, which also represents a fundamental autogenic time scale. At present, we are just beginning to understand what controls it. It is clear that as the total bedload flux and/or the net rate of deposition increases, the time required to create topography that forces lateral shifting decreases, thus decreasing the channel time scale T_{ch} (Wickert, 2007).

The shoreline is a sensitive indicator and recorder of autogenic dynamics. Kim et al. (2006a) showed, using data from the XES 02 experiment discussed above, significant variation in rate of shoreline migration even after lateral averaging to remove the effect of channel switching. The variability is greatest during transgression, when the shoreline is migrating against the average transport direction, and least during regression when the shoreline is migrating with the transport direction. Kim et al. (2006a) explained the behavior using a model of sediment storage and release over the length of the fluvial system. Storage occurs during periods when the fluvial channel network is relatively dispersed and inefficient, and is accompanied by slight increases in fluvial slope. Sediment release occurs via autogenic incision and excavation, relieving the excess slope and pushing the shoreline basinward. The slope variability is not large – of the order of a few percent of the mean slope, well within the scatter of measured fluvial slope values – but acting over natural basin length scales, even small slope variations can store and release significant sediment volumes. For instance, Kim et al. (2006a) found that storage-release events extrapolated from their experimental results would create fluctuations in sediment volume delivered to the shoreline sufficient to produce parasequence-scale shoreline excursions.

Autogenic phenomena associated with eustatic rise and fall have also been studied by Tetsuji Muto and colleagues at Nagasaki University. Muto and Steel (2001) showed that even under simple conditions – steady supply of well sorted sand, and steady sea-level rise – depositional lobe switching led to creation of a distinctive stepped morphology that they termed "autostepping" (Fig. 19). Analogously, channel switching and migration of the locus of incision during eustatic fall creates flights of terraces associated downdip with depositional lobes (Fig. 14) (Muto and Steel, 2004). The terraces in particular could be mistakenly interpreted as evidence for climate changes or unsteadiness in the base-level fall. Completing the picture, Kim and Muto (2007) have found comparable autogenic fluctuations in the location of the fluvial–bedrock transition during steady base-level fall and rise. The fluctuations were strongest during base-level fall and weakest during rise. Interestingly, fluctuations in the updip boundary of the depositional fluvial system vary in amplitude in precisely the opposite sense of those at the downdip (shoreline) end of the system (Kim et al., 2006a; Kim and Muto, 2007).

An important general idea emerging from this line of research is the intimate connection between autogenic dynamics and sediment storage and release. This idea, which was nicely illustrated in Ashmore's (1982, 1991) work on braided rivers, connects autogenic variability with the strong nonlinearity of sediment transport. The nonlinearity takes the form of transport thresholds and/or strongly nonlinear exponents in transport laws. The combination of sensitivity of flow to subtle changes in topography and sensitivity of sediment transport to local flow strength leads to sediment moving in steps with a wide range of length and time scales.

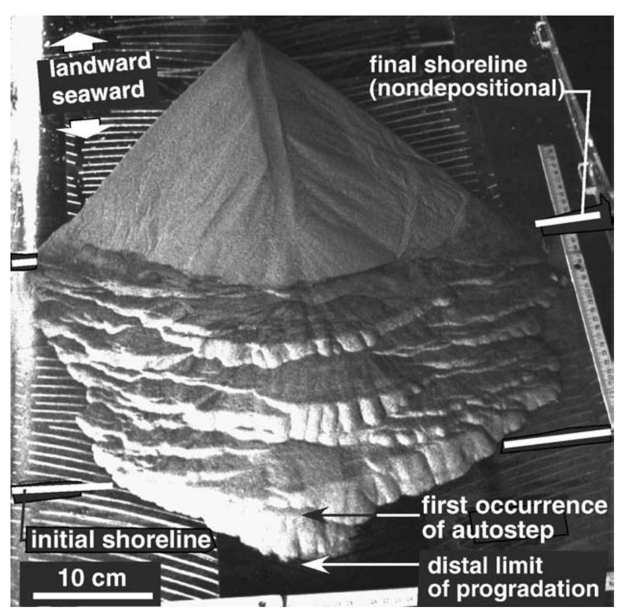

Fig. 19. Autostepping: autogenic delta lobes constructed during steady base-level rise in a relatively wide tank. From Muto and Steel (2001).

Finally, Sheets et al. (2002) studied the problem of how channel-controlled autogenic depositional processes add up to produce overall sedimentation patterns that compensate for subsidence. They found that the transition from channel control to subsidence control requires a time equivalent to the deposition of about six channel-depths worth of sediment. This defines, in effect, an integral scale for deposition. Lyons (2004) applied this idea to a field case of deposition from submarine channels, with the surprising result that the integral scale, defined in terms of deposit thickness to channel depth, was approximately the same as in the experimental case. The idea has been further extended by Straub et al. (2009), who used the decay rate of variability of observed rate of deposition with increasing stratal thickness as a measure of the extent of compensational stacking (the tendency for successive depositional bodies to avoid one another and thus fill space efficiently).

3.2.8. Summary and general findings from stratigraphic experiments

We begin by mirroring our previous comment about erosional systems: the single most important result of stratigraphic experiments is how well they work. The self-recording nature of depositional systems means that they create a record of their own temporal evolution. Thus the observation that experimental depositional systems self-organize to create not only surface configurations but also three-dimensional stratal patterns that resemble full-scale systems suggests that important aspects of time evolution as well as morphology are being captured. The main surfaces and facies tracts of sequence stratigraphy are readily reproduced even in extremely small and simplified experiments. The clear implication of this, as stressed by previous workers (Posamentier et al., 1992; Koss et al., 1994; Schlager, 2004), is that the sequence stratigraphic model and the main processes comprising it are fundamentally scale independent.

Another consistent theme emerging from the experimental work is the interplay of allogenic and autogenic dynamics in creating stratigraphy. The work of Muto and his colleagues illustrates particularly well the extent to which simple, steady eustatic change can produce complex stratigraphy. Most sequence-stratigraphic models have either underemphasized the role of autogenic dynamics or ignored it entirely. Using a metaphor introduced above, if stochastic autogenic variability is the stratigraphic equivalent of weather, then base-level cycles and other types of allogenic forcing can be thought of as “stratigraphic climate”. At reservoir and outcrop scales, the variability we see is mostly “weather”, i.e. it is autogenic. But the distinction between autogenic and allogenic effects on the basis of time scale may not be as clear as once thought. As discussed above, the entire structure of the depositional system is self-organized, and we now have experimental evidence for more than one case where this self-organization leads to long-period variability in response to steady forcing. As the temporal limit of autogenic processes is extended to longer time scales, the overlap regime between autogenic and allogenic processes becomes broader and the importance of understanding how they interact increases.

Subsiding-floor experiments make clear both the fundamental role of subsidence in controlling the form and preservation of sedimentary strata, and the extent to which it may produce little or no direct effect on the sediment surface. When an experiment begins, there is an initial period of rapid adjustment as the transport system sets itself up. Experiments with topographic scanning confirm that, as subsidence steadily extracts sediment from the transport system, the topography adjusts continually to the mass loss, smoothing out any direct imprint of the basement deformation. One prominent exception to this is provided by the XES 05 experiment described above, in which uplift and subsidence actively steered the fluvial channels; the key to understanding how subsidence directly influences channels at the surface seems to be the time-scale ratio discussed in Section 3.2.5.

The process by which an initially water-filled basin, supplied with sediment and water and subject to subsidence, develops a dynamic channel network, a migrating shoreline, and an offshore slope and apron, is entirely one of self-organization. The resemblance of the morphologic and stratal geometries to those of continental margins leaves little doubt that the continental margin, albeit influenced by the transition from continental to oceanic crust, is essentially a self-organized sedimentary landform.

Finally, we note that although model testing is in one of the best uses for stratigraphic experiments, it remains surprisingly uncommon. So far experiments have been mostly used to test qualitative predictions and as phenomenological case studies. One exception is the use of the results of the XES 96 experiment to test the Pitman (Pitman, 1978) shoreline-response model (Paola, 2000; Paola et al., 2001), and eventually to improve it (Swenson et al., 2000, 2005). Another model-testing application is the work of Kubo et al. (2005), in which XES data were used to evaluate the SEDFLUX model package developed by James Syvitski and his colleagues at University of Coloardo INSTAAR. Finally, Postma et al. (2008) have used experimental results to evaluate one of the oldest ideas in quantitative stratigraphic modeling, linear diffusion models for fluvial response. They found that the linear model does not perform well, and propose a new nonlinear version that gives better predictions of observed topographic profiles. These examples are a good start, but we hope that the next review of this topic will include much more use of experiments for quantitative model testing.

3.2.8.1. General implications for sequence stratigraphy. Much of the work summarized above has focused on testing and refining sequence stratigraphy as a predictive model. The clearest finding so far is that even in the simplified world of small-scale experiments, the response of stratigraphic systems to sea-level change is more complex than envisioned in textbook sequence-stratigraphic models. Major changes, for example in the direction of shoreline migration, can occur during steady rise or fall of base level (Muto, 2001; Muto and Steel, 2004; Swenson and Muto, 2007; Petter and Muto, 2008). Sequence boundary development is complicated by ongoing modification of the erosional valley and time lags associated with propagation of effects across the system (Van Heijst and Postma, 2001; Strong and Paola, 2006, 2008). The most far-reaching claim regarding conventional sequence stratigraphy is that made recently by Muto et al. (2007), who assert that sequence stratigraphy is valid only for time scales too short for the nonlinear system responses documented above to come into play. This would severely limit its applicability to natural systems.

While acknowledging the complexity of stratigraphic response to eustatic changes, we nonetheless take a less drastic view. Some of the discord between experimental findings and sequence stratigraphy arises from the natural tendency of researchers to focus on how their findings require changes in conventional thinking. So we begin by pointing out that experiments with cyclic changes in base level, which is what sequence stratigraphy was orginally intended for, produce results that are broadly consistent with the conceptual core of sequence stratigraphy. This point was first made by Posamentier et al. (1992), using a natural small-scale experiment in the unprepossessing form of a fan delta growing into a drainage ditch. In our experience teaching industrial short courses, we find that experiments are a powerful and effective way of seeing sequence stratigraphy in action, and connecting stratigraphic products to surface processes. Canonical sequence-stratigraphic elements — sequence boundaries, maximum flooding surfaces, transgressive surfaces, and low-stand, high-stand, and transgressive systems tracts — can easily be reproduced at small scales (<1 m) using a simple tank, water, and sand. (Plans for such a tank can be found at www.nced.umn.edu.) The point is that the sequence stratigraphic model in its basic form is generic and scale independent, as Posamentier et al. (1992) concluded. This observation is consistent with its widespread successful application in industry in a range of settings.

The question then is how to use the findings of past and future experiments to improve sequence stratigraphy. The fundamental point being made by Muto et al. (2007) and Paola et al. (2001) is that the potential complexity of stratigraphic response to simple forcing requires a conservative approach to interpretation. Experimental stratigraphy makes it extremely clear that *the relation between cause and effect in stratigraphy is generally not one to one*. The work summarized above provides a number of examples in which simple inputs produce complex outputs that invite unnecessary elaboration of causes. In these cases, the experiments provide new templates that field workers can use in searching for the simplest explanations for the complex structures they observe. The first question must always be: how much of what I see could be autogenic? What is the minimal external forcing required? We also have cases, such as the XES 02 experiment discussed above, in which basin-scale amalgamation causes multiple allogenic cycle to produce a single, composite stratigraphic signature that, at best, can be teased apart only using relatively subtle criteria.

A second point is that classical sequence stratigraphy is more “allogenic” than it should be. Even for the canonical case where sequences are produced by simple eustatic cycles, allogenic effects interact in interesting and sometimes surprising ways with the internal dynamics of the transport system. Some of the effects highlighted above include multiple phases of response, substantial time lags leading to out-of-phase system response to the same signal, and strong stochastic overprinting of the allogenic signal. In addition, sequence stratigraphic surfaces are dynamic, so that they evolve in time. The degree of time variability varies with the type of surface. The sequence boundary, often depicted as a valley that forms during falling stage and then fills passively, actually evolves over the whole base-level cycle. In the limit as cycle time scale becomes long compared to the rate at which rivers migrate laterally, the valley is replaced by a broad erosion surface.

A final point is that experimental results support the idea, advocated by many others on the basis of field observations and theory, that sequence stratigraphy has overemphasized eustatic cycles in explaining stratigraphic patterns. Even where deposit patterns produced by other forcing processes (e.g. water supply cycles) differ from those produced by eustatic variation, it remains important to keep in mind the potential for misinterpretation. Experiments, by allowing us to study the effects of specific combinations of imposed change, are an ideal method for developing better quantitative methods for teasing apart the effects of multiple superimposed allogenic effects on preserved stratigraphy.

We have focused in these first two sections of Section 3 on erosional and depositional systems as a whole. Now we turn to experimental studies of specific morphodynamic environments.

3.3. Alluvial fans

The steep slopes and coarse grain sizes of many alluvial fans make them a natural target of experimental study, exemplified by classical studies such as Hooke and Rohrer (1979) and the experiments discussed in Schumm et al. (1987). Other more recent work involving experimental alluvial fans has been aimed at measuring avulsion and alluvial architecture and thus was discussed earlier, in Section 3.2.6. Whipple et al. (1998), as part of an applied study of controls on depositional slope in a mine-tailings fan, evaluated the effect of changes in grain size and water and sediment discharge on fan slope under conditions of steady uniform aggradation. The work included successful tests of a quantitative fan theory developed by Parker et al. (1998). Bedload-dominated fans exhibited nearly straight profiles, consistent with many natural alluvial fans, in which water discharge exerted a much stronger control on morphology than grain size as long as shear stresses were well above critical.

Milana (1998) and Milana and Tietze (2002) have carried out experiments on the effect of changes in water supply on the morphology and stratigraphy of depositional alluvial fans. This work was mainly aimed at stratigraphy and was discussed above in Section 3.2.4. The morphologic results parallel those of Whipple et al. (1998), but one point to note is their observation of profile convexity produced by the net effect of decreases in water supply that led to increased sediment storage and slope in the upstream parts of the fan.

3.4. Deltas

Most of the depositional-system experiments discussed above end in standing water, so in a sense they can be considered to be deltas, or more exactly, fan-deltas. Here we look at experiments focusing on deltas *per se*, emphasizing the common case of fine-grained deltas. Overall it is surprising how little attention deltas have received from experimentalists given the number of compelling motivations for studying them: the presence of some hundreds of millions of people living on or near deltas that are increasingly vulnerable to sea-level rise; efforts to restore the Mississippi Delta following Hurricane Katrina; ongoing interest in deltas as hydrocarbon reservoirs; and deltas as the “type” clinoform, the fundamental building block of continental margins. Experimentation has a significant role to play in understanding the self-organized channel networks and clinoform shapes by which deltas grow and maintain themselves.

One thread of experimental research on deltas has targeted their overall longitudinal profile, emphasizing controls on foreset slope. In a series of papers combining experiments with theory and field data, Kostic et al. (2002) and Kostic and Parker (2003a,b) showed how overpassing hyperpycnal turbidity currents make deltaic foresets longer and gentler because, in effect, the basal shear stresses they apply act to augment gravity. This reduces the slope needed for sediment transport. Lai and Capart (2007, 2009b) build on this approach, showing how the modified foreset profile can be represented via a two-diffusion moving-boundary model that nicely matches a series of experimental profiles produced with varying rates of hyperpycnal flow. One especially important result is that, because the hyperpycnal flows cause ongoing sediment transport on the deltaic foresets, the upper foreset can undergo net erosion during transgression and relative sea-level rise. The authors propose this as a potential mechanism for initiating submarine canyons; it could also presumably lead to ravinement and net loss of shoreface deposits during transgression. The most recent in this series (Lai and Capart, 2009a), and the related paper by Lorenzo-Trueba et al. (2009) nicely illustrate both morphodynamic modeling using moving-boundary methods (with three boundaries) and how these can be effectively tested using simple but elegant experiments.

As is the case for streams, the channel network that develops on experimental sandy delta tops is braided. Most natural deltas, formed of cohesive sediment with a strong influence of vegetation, have distributary channel networks with a more or less lobate plan form and a correspondingly intricate shoreline, quite different from the produced by a sandy braided fan-delta. The experimental stratigraphy group at ExxonMobil Upstream Research laboratory has developed a new weakly cohesive-sediment mix that allows creation of deltas with well developed bird's foot style distributary patterns at laboratory scales (Hoyal and Sheets, 2009) (Fig. 20). This is a major advance that will increase the applicability of experimental delta research to the field, including the various societal issues listed above. The first experiments reported using the new mix (Hoyal and Sheets, 2009) demonstrate its capability to capture field-scale processes in fine-grained deltas in terms of both system evolution and stratigraphy. One of the most interesting ideas emerging from this work is the idea that avulsion in distributary systems is driven not only by upstream conditions (aggradation) but also from downstream, an effect that Hoyal and Sheets (2009) refer to as a “morphodynamic backwater”. The influence of downstream conditions on the eventual success of channel avulsion has

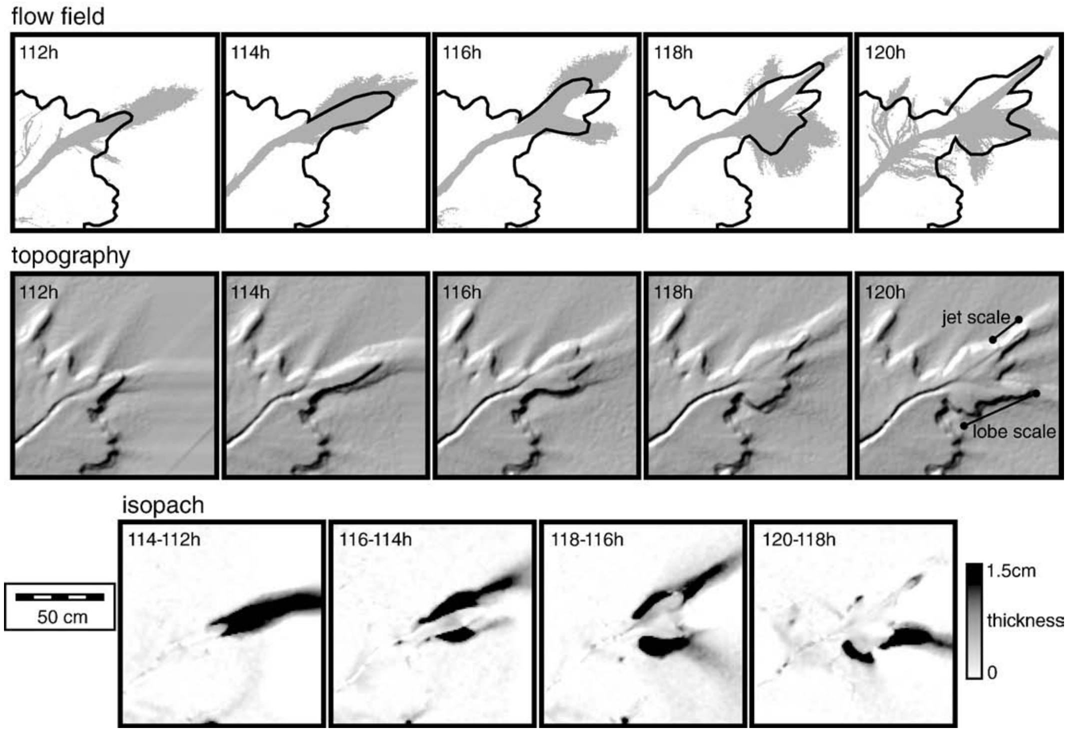

Fig. 20. Spatial distribution of flow, topography, and deposition in an experiment using a novel weakly cohesive-sediment mix developed at ExxonMobil Upstream Research laboratory. Highlights include a well developed distributary pattern, and episodes of channel extension (112–114 h), bifurcation (116–118 h), and overbank flow (120 h). Dark line in the flow field images is a contour 0.01 m above the base of the experiment. Characteristic length scales for jets and depositional lobes are shown in the topography images. From Hoyal and Sheets (2009).

previously been underappreciated but could be important in other types of channel systems as well.

3.5. Rivers

3.5.1. Bedrock and erosional channels

Recent years have seen a renewed interest in the dynamics of erosional channels, especially bedrock and mixed bedrock — alluvial channels. At the largest-scale end of this is an experimental study by Douglass and Schmeeckle (2007) on development of transverse drainages, i.e. erosional channels cutting across local topographic highs. A noteworthy feature of this work is their use of a clever inflatable bladder mechanism to induce local uplift; this is a technique that should see wider application in experimental geomorphology Douglass and Schmeeckle were able to investigate the full range of proposed mechanisms for creating transverse drainages under controlled conditions. The work illustrates, for example, the influence of downstream sediment transport on the potential for a stream to maintain course over a bedrock uplift by controlling the removal of debris generated by erosion of the rising uplift. The research also led to improved observational criteria, detailed in the paper, for differentiating among causal mechanisms of transverse drainage in the field.

The development and dynamics of erosional (bedrock) channels have been another area of major research effort. One important research thread is the influence of sediment cover on overall erosion rate; this work falls below the scale cutoff for this review, but we point the interested reader to the works of, for example, Sklar and Dietrich (2001, 2004) for more information. At a somewhat larger scale, Johnson and Whipple (2007) showed that, analogously to alluvial channels, the form of erosional channels is strongly coupled to bedload sediment transport. The forms developed include potholes, bends, slot-type inner channels, and undercuts, all comparable in form and relative size to field examples (Fig. 21).

3.5.2. Braided rivers

Braided rivers constitute a minority of natural rivers but the majority of experimental rivers. Laboratory-scale braided rivers have been made in materials ranging from silt to pea gravel (Figs. 22 and 23). The ease with which braiding can be created over a wide range of scales and other conditions led Murray and Paola (1994) to suggest that it is the fundamental physical instability of all but very narrow or laterally constrained non-cohesive channels. Experimental braided rivers are relatively amenable to classical scaling (Section 2.1) via one of the variants of Froude scaling (Peakall et al., 1996; Cazanacli et al., 2002). The problem of scaling the sand-size sediment present in most gravel-bed rivers still remains, as discussed in Section 2.1.3, meaning that the sediment cannot be exactly scaled unless the field prototype is exclusively gravel-sized or coarser.

Dimensionless morphologic indices from experimental braided rivers correspond well to field examples (Ashmore, 1982; Ashmore and Parker, 1983; Ashworth et al., 1999). These studies also show that the main morphodynamic processes in the experimental braided rivers are qualitatively similar to their field analogs. In addition to elements like bars and confluences, this also includes strong stochastic variation in the bedload flux resulting from internal processes of sediment storage and release (Ashmore, 1991; Ashmore and Gardner, 2008). A key morphologic relation – the ratio of scour depth relative to mean channel depth and confluence angle – has also been shown to be the same (~5) from experimental braid channels of the order of 0.01 m deep to the largest braided river on Earth, the Ganges–Brahmaputra (Ashmore and Parker, 1983; Best and Ashworth, 1997). Leddy et al. (1993) have shown that processes of anabranch avulsion in experimental braided rivers are comparable to those of natural-scale rivers as well. Braided river research has also provided the first concrete example of a method by which kinematic (as opposed to purely geometric) similarity might be quantified. The method proposed by Sapozhnikov and Foufoula-Georgiou (1997, 1999) is based on

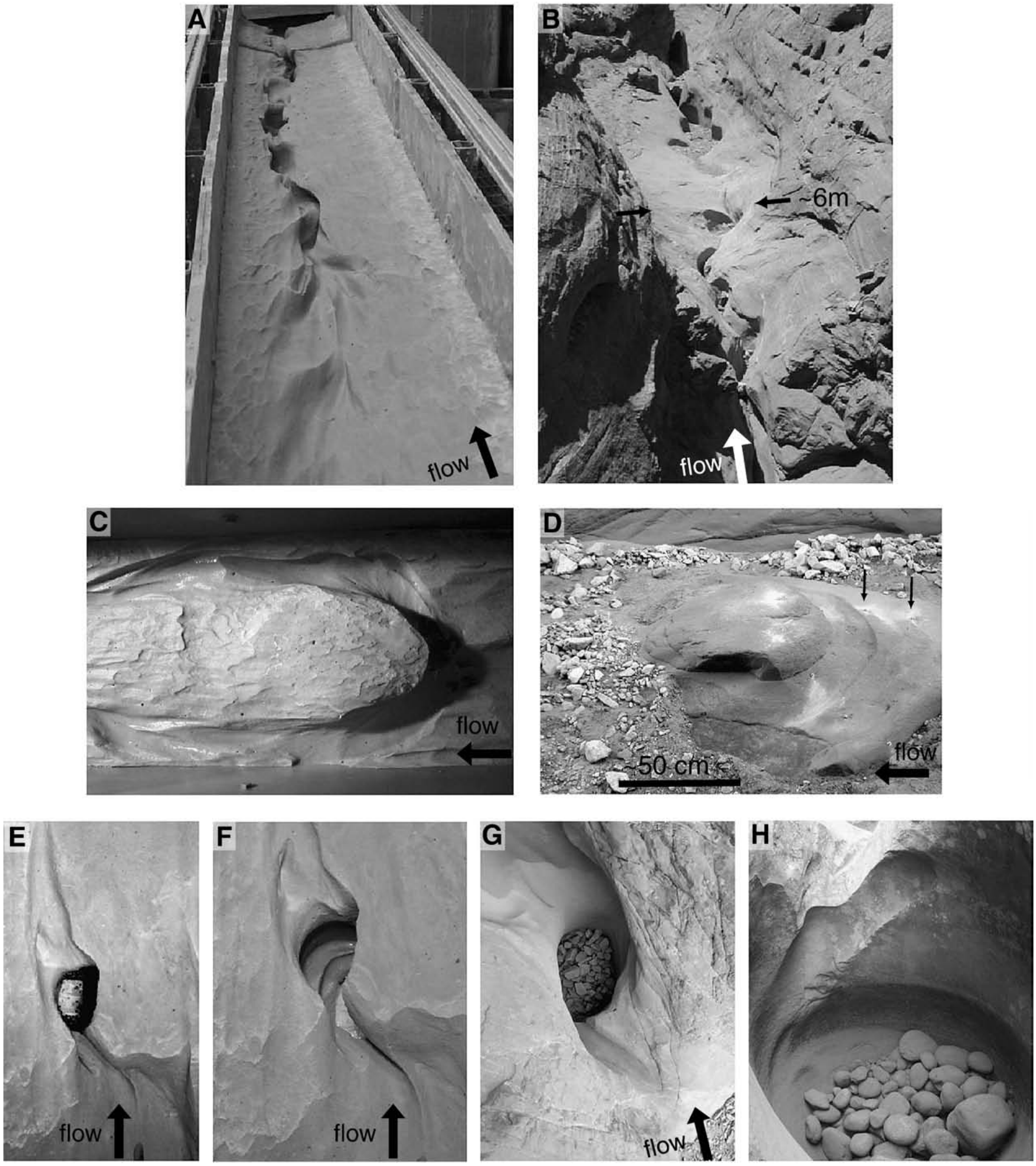

Fig. 21. Erosional channel morphology from an experiment and the field. Field photographs are from channels and canyons in the Henry Mountains, Utah cut into the Navajo Sandstone. (A) Oblique view looking downstream, flume experiment, flume width 0.4 m. (B) Natural inner channel, south fork of Maidenwater Creek. (C) Overhead view of erosion around a broad protrusion molded into the bed, flume experiment. Flume width 40 cm. (D) Similar geometry in a Henry Mountains channel, initial condition unknown. White patches are pulverized rock due to drilling. (E), (F) Views of an experimental pothole. The bed was molded to have vertical steps, and the pothole developed where the bed was initially horizontal and planar. Pothole diameter ~0.055 m. (G) Field pothole (diameter ~0.6 m), partially filled with sediment clasts. (H) Field pothole, diameter ~2 m. From Johnson and Whipple (2007).

measuring the probability distribution function for planform changes by size (area of change) as a function of the time interval over which the changes are measured. The result is, in the case of braiding, a power–law relation between the length and time scales of change. So far the method has been applied so far only to experimental data, but it is not intrinsically scale dependent and so should apply equally in the field. Collecting the required data in the field presents challenges not only of much greater length and time scales but also of filtering out the effect of discharge variation. We hope that having a well defined theory to test will inspire somone to make the effort.

The multiplicity of channels is one of the key features that distinguish braided rivers from other planform types. Researchers at the universities of Trento and Western Ontario have used experiments to study braided rivers under a range of constant and variable water discharges. The results confirm the dynamism of steady-state braiding and the tendency to widen and add channels as discharge increases (Bertoldi et al., 2009b; Egozi and Ashmore, 2008) A major result of this work is to demonstrate quantitatively that only a fraction – a typical value seems to be about 40% – of the braid channels have active sediment transport (Bertoldi et al., 2009b; Egozi and Ashmore, 2009).

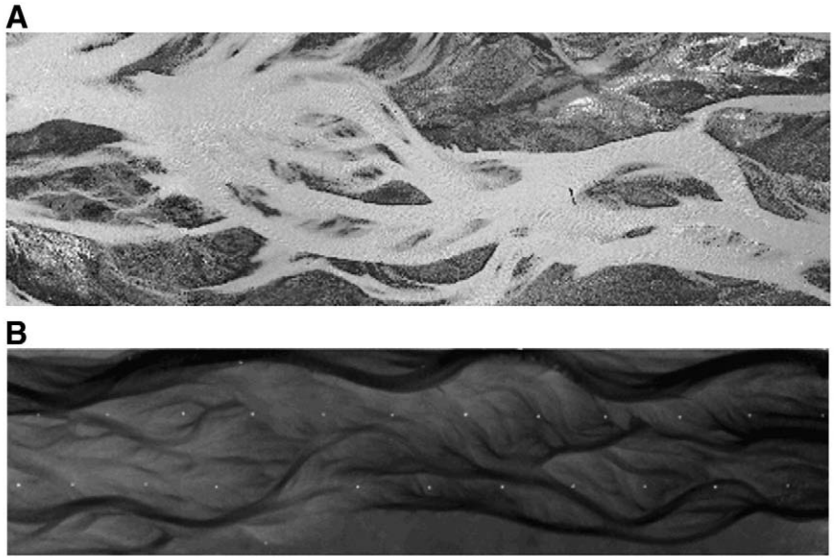

Fig. 22. (A) Overhead orthophoto of the Sunwapta River, Alberta, compared with (B) DEM of a scaled experimental (approximately 1:30) braided river in a laboratory flume, area approximately 12 m × 3 m. Darker shades indicate lower elevation. Confluences are marked with ovals and bend scours with rectangles in the experiment image. Flow from left to right. Images courtesy of Peter Ashmore, University of Western Ontario (Canada).

The distinction of inactive and active parts of the flow is crucial for correctly estimating the total bedload flux (Bertoldi et al., 2009a). For low-intensity braiding, there is often only one main active channel, which dominates the morphodynamics (Egozi and Ashmore, 2009). As Egozi and Ashmore (2009) point out, this observation offers an avenue both for simplifying analysis of braided-river behavior and for strengthening the conceptual connections between braided and single-thread rivers.

3.5.3. Single-thread rivers, including meandering

Braided rivers are easy to produce experimentally, but globally they are relatively uncommon. Single-thread rivers, and in particular meandering rivers, are common in nature but have proved surprisingly hard to produce experimentally. In fact, to our knowledge no one has yet produced a fully self-formed and self-sustaining meandering river experimentally. Most experimental reports on the onset of meandering are referring to the presence of alternate bars and weakly

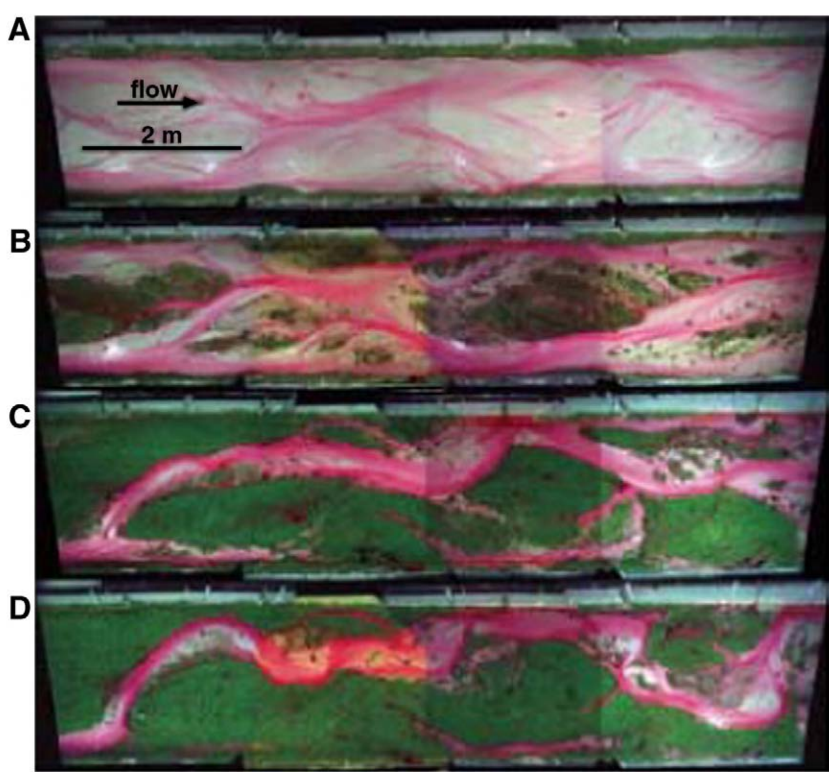

Fig. 23. Overhead images of an experiment in which a dynamic single-thread channel developed through interaction of vegetation and an initial sand-bed braided channel. The irregularly sinuous, single-thread channel develops well defined banks and a floodplain through "corralling" of the flow by vegetation seeded and allowed to grow during repeated cycles of reduced water and sediment discharge followed by shorter periods of high discharge. The 8 m study reach of the flume is shown (A) at initial unvegetated steady state, (B) after 6 flood cycles (24 d), (C) after 18 flood cycles (72 d), and (D) after 23 flood cycles (92 d). The channel reached a dynamic steady state in which vegetated area was destroyed along cut banks at the same rate that new area was created by bar deposition and colonized. From Tal and Paola (2007).

sinuous thalwegs; left to evolve, these begin cutting new channels across the alternate bars and end up producing low-index braiding. Some inspiring early attempts at creating experimental meanders are described in Schumm et al. (1987). They grew meanders to high amplitude by beginning with stratified sediment: sand topped by a thin layer of clay. The clay added enough strength to the sediment to discourage cutting of new opportunistic channels across the point bar, and the meanders grew to high amplitude. But the stratified sediment structure was imposed by the experimenters, so the meanders were not self-maintaining. Smith (1998) used small discharges and a mix of fine-grained, cohesive-sediment types to produce entirely self-formed meanders of high amplitude. The meander morphology is angular relative to field examples, but the bends appeared capable of generating cutoffs and then re-growing to high amplitude. Gran and Paola (2001) showed that adding growing vegetation (alfalfa, *Medicago sativa*) to an experimental braided stream led to creation of a single-thread channel whose banks could withstand relatively high shear stresses without eroding. Tal and Paola (2007) have shown that repeated cycles of high and low discharge coupled with repeated alfalfa seeding caused reorganization of the braided network to a dynamic single-thread channel that showed a number of characteristics of natural meandering: point bar growth, high bend amplitudes, bend cutoff and regrowth, and channel avulsion (Fig. 23). In parallel with this work, Peakall et al. (2007b) produced sinuous channels using cohesive sediment, without vegetation. It appears that the experimental community has finally solved the problem of producing dynamic, self-sustaining meandering experimentally, opening the door to experimental evaluation of management and restoration techniques for meandering streams.

An interesting problem that deserves more attention is how individual river channels interact with faults. Important early work on this was reported by Schumm et al. (1987). Relatively little has been done since then, with the exception of a clever experiment by Ouchi (2004). Ouchi inserted a laterally movable segment into a single-thread experimental river channel formed in a sand–clay mixture. Displacement of the segment effectively lengthened the channel and locally increased its sinuosity, creating a kind of artificial meander. The channel responded by attempting to recover a consistent long profile through upstream aggradation and downstream incision, analogous though precisely opposite in sense to the well known case of channel response to channel straightening by meander cutoff.

3.5.4. Autogenic river processes

Most of the autogenic processes seen in experimental river studies have been discussed in the previous sections on erosional and depositional systems. These include autogenic terrace formation (Hasbargen and Paola, 2000; Muto and Steel, 2004; Strong and Paola, 2006), stream piracy and capture (Hasbargen and Paola, 2000; Douglass and Schmeeckle, 2007), and avulsion (Bryant et al., 1995; Muto and Steel, 2001; Cazanacli et al., 2002; Ashworth et al., 2007). One common observation in weakly channelized transport systems is autogenic variability in the degree of channelization (Whipple et al., 1998; Kim and Paola, 2007; Van Dijk et al., 2009). Episodes of greater channelization are associated with incision, reduction in overall slope, and reduced sediment storage in the transport system. This is a somewhat larger-scale version of the fundamental process of sediment storage and release in multi-thread river systems originally identified by Ashmore (1982, 1991), in which zones of sediment storage are associated with production of channel width, and sediment flushing with total-width reduction.

As shown by Tal and Paola (2007), the addition of a stabilizing agent such as vegetation not only leads to reduction in the total number of active channels but also changes the mode of autogenic channel mobility from stochastic channel switches on a range of length and time scales to a combination of coherent lateral migration and well defined but less infrequent avulsions.

3.6. Deep-water processes

Experimental research takes on a new importance in the deep-marine realm, where water depth and infrequent occurrence of flow events hamper direct observations of turbidity currents and debris flows (Hay, 1987; Khripounoff et al., 2003; Xu et al., 2004; Yu et al., 2006). Dynamic processes are much harder to observe directly than in terrestrial landscapes. Laboratory experiments have been used to test specific hypothesis but have also acted as tools to improve our general intuition of deep-marine processes. This experimentally derived intuition is especially critical in the deep-marine environment because strong coupling of flows to their ambient fluid, seawater, in the deep-marine makes direct transfer of terrestrial morphodynamic laws into the deep-marine difficult. In contrast to the scarcity of data on the fluid mechanics of deep-water processes, there is a wealth of geometric data describing the static morphology and stratigraphy of these systems (Kenyon et al., 1995; Twichell et al., 1995; Deptuck et al., 2003, 2007). These data are the product of recent advances in geophysical imaging of continental margins. Recent experiments have focused on characterizing the flow properties of turbidity currents and debris flows that produce depositional features similar to what is being observed with these new geophysical tools (Mohrig et al., 1999; Metivier et al., 2005; Yu et al., 2006; Peakall et al., 2007a; Straub et al., 2008).

Turbidity currents and debris flows have been studied at laboratory scales for over forty years. Most early experiments were performed in flumes designed to minimize lateral variation in the flow field, and their focus was the fluid mechanics of deep-water flows (Middleton, 1966; Hampton, 1972; Pantin, 1979; Simpson and Britter, 1979; Parker et al., 1987; Dade et al., 1994). These experiments refined our understanding of the streamwise evolution of 2D confined flows but have proved difficult to check at field scale due to the lack of available flow data from natural systems. Several flume experiments have focused on the depositional and erosional trends of turbidity currents (Parker et al., 1987; Garcia, 1994). These experiments improved our theoretical understanding of the depositional mechanics of turbidity currents and aided the development of 2D numerical models of turbidity currents. Mapping of continental margins and associated depositional features in seismic volumes reveals highly sinuous channels with extensive overbank deposits and large unchannelized depositional lobes (Kenyon et al., 1995; Pirmez and Flood, 1995; Posamentier, 2003). These features point to the highly three-dimensional nature of the sediment transport field in deep-marine flows, which cannot be captured in two-dimensional laboratory studies. Recent experiments have focused on capturing the importance of the third dimension. Broadly, these experiments have been focused in two different areas: construction of submarine fans and interaction of turbidity currents with channels. In addition to these areas of study, several recent experiments have focused on the interaction of turbidity currents with mini-basins a subject of particular importance to the petroleum industry at present. These studies have included both two- and three-dimensional experimental setups.

Unlike many recent experimental studies of terrestrial processes, which focused on autogenic processes, deep-water process experiments have almost exclusively focused on determining how allogenic forcing influences morphodynamics. This is likely the result of a difference in degree of theoretical understanding in the two environments. In comparison to the terrestrial, many first order controls of sediment transport are unknown in the marine. It is likely that as greater understanding of deep-water system responses to allogenic forcing occurs, more experiments focused on autogenic processes will take place.

To date, the great majority of deep-water experiments have addressed issues of scaling through classical engineering methods (Section 2.1). As far as possible, this is done through comparison of three components: 1) geometric scaling of the static topography; 2) dynamic scaling of flow properties for estimating equivalence between model and natural flows and; 3) dynamic scaling of the sediment transport to compare model and natural sediment-transport regimes. Similar to classical scaling of terrestrial systems, the greatest problems in comparing laboratory- and field-scale deep-marine flows come from the reduced *Re* of laboratory flows and difficulty in matching the *Gr* of field-scale flows.

In the following sections we summarize recent advances in deep-water processes resulting from laboratory experiments, focusing on 1) submarine fans, 2) interaction of turbidity currents with channels, 3) interaction of turbidity currents with mini-basins, and 4) debris flows. We limit our discussion to experiments with evolving beds and sediment transporting flows. These conditions allow comparison of experimental data with the greatest amount of information available for natural submarine systems, seismic images of topography, and associated deposits.

3.6.1. Submarine fans

The largest deposits of sediment on Earth are found in submarine fans. In addition to occupying a critical place as the final sink in source-sink transport systems, submarine fans host many of the largest producing petroleum reservoirs (Weimer and Link, 1991).

Given the obvious difficulty of studying submarine fans in the field, it is not surprising that they have attracted the attention of experimenters. The initial series of experiments, beginning with the work of Luthi (1981), focused on bulk fan morphology and dynamics. Luthi's experiments showed decreases in flow speed, deposit thickness and grain size with downstream distance for sustained turbidity currents in an unconfined basin. These overall patterns matched those observed in the field, including development of a downstream succession of bedforms corresponding to the classic Bouma sequence for turbidites. That Luthi's observed downstream succession of grain size and bedform type compares well with the field is reassuring, and provides a nice application of Walther's Law across scales. The observed downstream fining is especially noteworthy, because downstream fining has been difficult to reproduce in fluvial experiments (Paola et al., 1992b). There, unless special care is taken, experiments typically show no change or downstream coarsening (Solari and Parker, 2000). Evidently the bedload-dominated sorting typical of fluvial experiments is more susceptible to scale effects (via the bed slope, according to Solari and Parker) than the suspended-load dominated sediment mechanics of turbidity currents.

Following Luthi's work, several authors examined the interaction of unchannelized flows with obstacles meant to represent sea-floor topographic features such as tilted fault blocks (Alexander and Morris, 1994; Kneller, 1995). The small obstacles induced stationary mixing vortices near the obstacles and enhanced deposition downstream of the obstacles. These experiments demonstrated the influence of suspended-sediment concentration, grain size, and obstacle shape on the location of stationary vortices and enhanced deposition, but have not yet led to a relationship to predict these sites given an arbitrary combination of the controlling parameters. Parsons et al. (2002) studied a case of flow interaction with self-generated topography by releasing sequences of unconfined flows. Autogenic flow focusing

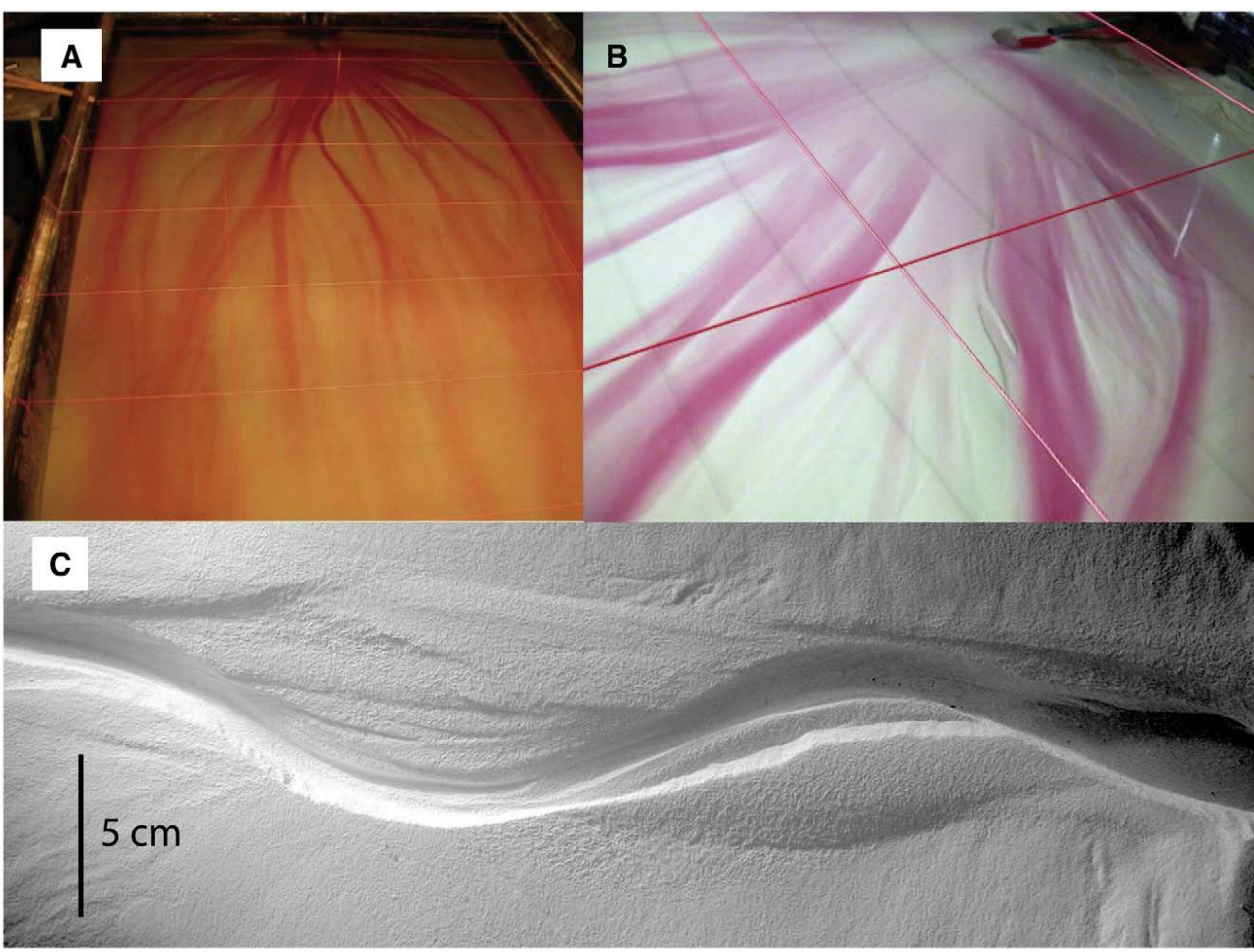

Fig. 24. (A) View of intricate topography associated with channelization of the experimental fan reported by Yu et al. (2006). Flow of sediment-laden turbidity currents was from top to bottom. The width of the basin is 2.15 m. (B) Close-up view of the channels just below the basin entrance point. (C) Erosional submarine channel created by a saline current over low-density sediment (Metivier et al., 2005).

developed after a significant number of individual flow events and led to creation of distinct depositional lobes. As these lobes grew, the ratio of lateral to downstream deposit slope increased until the transverse gravitational potential energy gradient was high enough to induce lobe avulsion.

The experiments described above have increased our general understanding of submarine fan development. But most submarine fans are intensively channelized (Pirmez and Flood, 1995; Schwenk et al., 2003), while the experimental submarine fans discussed above were devoid of channels. In a recent breakthrough, three different experimental configurations have produced self-channelized subaqueous fans by density currents (Metivier et al., 2005; Yu et al., 2006; Hoyal et al., 2008). Metivier et al. (2005) reported the first spontaneous formation of submarine channels longer than a few centimeters. They accomplished this by running brine density currents over an erodible bed composed of low-density ($\rho_s = 1080$ kg/m^3) plastic sediment. The key was the low-density sediment, which reduced the near bed shear stress required to mobilize the bed. Channels in this experiment displayed a range of sinuosities up to about 1.1 and included channel bends that migrated downstream. Based on their experiments Metivier et al. (2005) suggested two critical conditions for channel formation via incision: 1) currents must be long lived, and 2) near-bed shear stress must be sufficient to erode the bed.

Yu et al. (2006) produced self-channelized subaqueous fans with an experimental setup that substantially differed from the initial experiments reported by Metivier et al. (2005). Yu et al. produced channelized fans through the continuous feed of a sediment-laden flow into an experimental basin (Fig. 24). One important factor in their success was the addition of a significant (>20%) fraction of kaolinite clay in the turbidity-current sediment mixture. The kaolinite added cohesion to the fan surface after deposition, which was critical to initiating and maintaining channels. The second imporant condition for channel formation was that the input turbidity current was too small to cover the entire area of the fan at any one time. This resulted in flow instabilities that created a mix of erosional and depositional channels on the experimental submarine fans. The strength of this experimental setup is that channel formation occurred from suspension dominated turbidity currents.

Finally, Hoyal et al. (2008) have reported preliminary findings on self-channelized submarine fans generated by the continuous feed of brine density flows over a mixture of plastic sediment, silica sediment and a polymer that makes the sediment mixture weakly cohesive. The resulting submarine channels display distributary network characteristics similar to many natural fans.

The initial experiment series described above show that overall fan morphology, grain size and bedform patterns, and autogenic lobe switching can be produced experimentally. Analogously to the development of laboratory techniques for creating dynamic single-thread channels, the "channelization breakthrough" of the last three years opens a new door by showing the way to experimental study of the channels and channel networks that create the mesoscale stratigraphy of submarine fan deposits.

3.6.2. Interaction of turbidity current with channels

Although creation of self-channelized submarine fans opens several lines of experimental investigation, the scale of the channels in these experiments is too small to allow detailed measurement of the interactions between flow and local channel topography. Examination of these interactions has so far occurred in pre-constructed non-erodible channels that are more than in order of magnitude larger in width and depth than the self-generated channels described above. While experiments with fixed beds do not allow study of erosional processes, topographic influences on deposition can be examined in detail. Recent laboratory experiments focused on turbidity-current interactions with channels have focused on two issues: 1) interaction of turbidity currents with channel bends and 2) construction of channel-margin levees.

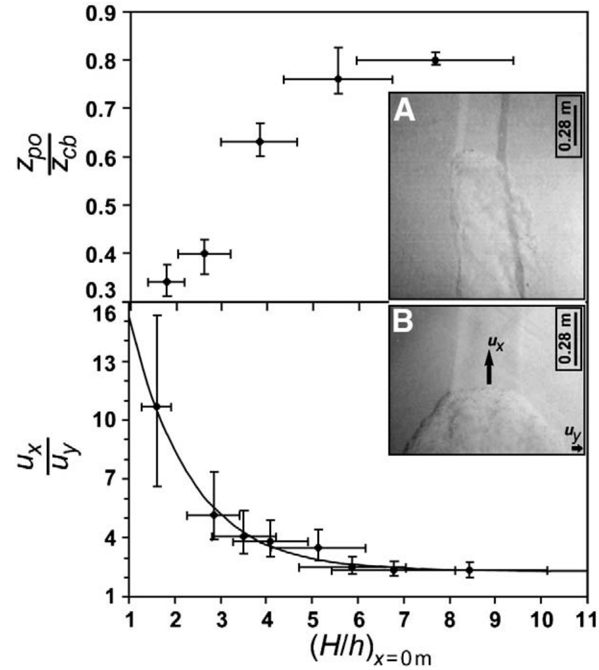

Fig. 25. Change in sedimentation pattern and flow spreading as a function of turbidity current thickness to channel depth, H/h, reported in Mohrig and Buttles (2007). (A) Ratio of proximal overbank sedimentation, z_{po}, to channel bottom sedimentation, z_{cb}, in the experimental channel system. Error bars bracket total variability in measured values of z_{po}/z_{cb} and total measurement error associated with H/h. Inset photo captures the head of the current when $H/h = 1.6$. (B) Ratio of longitudinal to lateral components of velocity u_x/u_y for eight turbidity currents. Error bars bracket uncertainty associated with each value of u_x/u_y. Inset photo captures the head of the current when $H/h = 8.4$.

Most submarine channels in excess of 100 km long are moderately to highly sinuous, including the Amazon (Pirmez and Flood, 1995), Indus (Kenyon et al., 1995), and Bengal (Schwenk et al., 2003) channels. While the planform statistics of submarine channels are similar to those of rivers (Pirmez and Imran, 2003) there are important differences in the cross-sectional geometry of channels in the two environments. Seismic cross-sections indicate that aggradation commonly occurs at higher rates than horizontal migration of channel bends (Pirmez and Flood, 1995). This situation rarely occurs in rivers and seems to be partially the result of high rates of overbank deposition relative to in-channel deposition for submarine channels. These observations have motivated several experimental studies aimed at determining the in-channel depositional mechanics and processes responsible for transporting sediment overbank. Mohrig and Buttles (2007) found that one cause of high overbank deposition in submarine channels is that shallow submarine channels can be constructed from relatively thick turbidity currents. They monitored the lateral and downstream velocity of ten turbidity currents interacting with a low sinuosity channel to determine the conditions under which currents are steered by channels. With each successive turbidity-current event, the ratio of current thickness to channel depth decreased through preferential in-channel deposition. Mohrig and Buttles found that currents are effectively channelized if their thicknesses are less than 1.3 times the local channel depth, while currents do not transition to a fully unconfined state until current thickness is 5 times local channel depth (Fig. 25). The key insight is that a thick current is effectively channelized as long as its velocity maximum, typically located low in the flow, is contained within confining channel levees. The flow superstructure above the velocity maximum is free to go overbank, which helps explain the high rates of overbank deposition for many sinuous deep-marine channels.

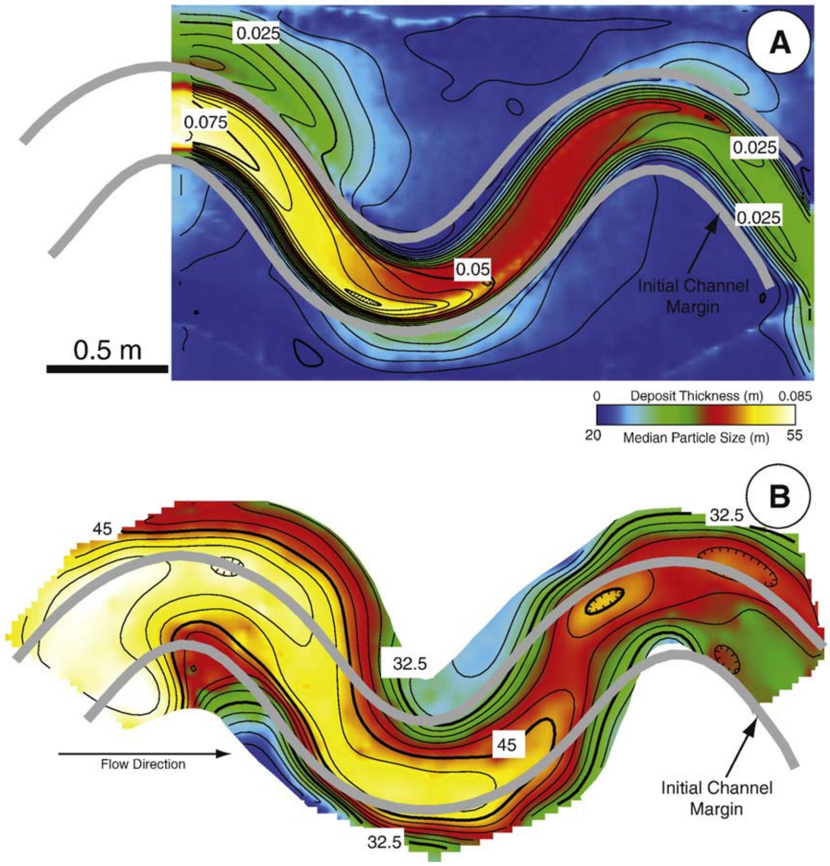

Fig. 26. Maps of deposit thickness (A) and median grain size (B) of deposit resulting from deposition of 24 sediment-laden turbidity currents interacting with a sinuous channel reported in Straub et al. (2008). Channel flow was from left to right in each map. Bold gray lines represent location of channel margin prior to deposition by turbidity currents. Strong cross-channel asymmetry of deposit thickness and deposit median grain size at channel bends resulted from inertial runup of turbidity currents onto outer banks of channel bends. Contour intervals are 5 mm and 2.5 mm for deposit thickness and deposit median grain size respectively.

Straub et al. (2008) recently reported results from a laboratory experiment on the evolution of a sinuous channel via sedimentation from turbidity currents. A major focus of the experiment was testing the consequences of general flow-splitting models that describe processes responsible for transport of sediment to overbank environments at channel bends. In this experimental series, 24 depositional turbidity currents with constant initial conditions were released into a channel with three bends and a sinuosity of 1.32. Data collected during the experiment included flow velocity, thickness, and channel topography and deposit particle size. The sedimentation pattern, as indicated by deposit thickness and particle size, was skewed towards the outer banks of bends (Fig. 26). The asymmetry in deposit thickness was large enough that levee crest deposits on the outer banks of bends were as coarse as sediment deposited in the center of the channel. The authors hypothesized that the high asymmetry in deposit properties is the result of inertial runup of currents onto the outer banks of channel bends. Current superelevations measured at the experimental channel bends greatly exceed those predicted by standard superelevation equations, which balance centrifugal accelerations with associated restoring pressure gradient forces. The excess runup allows for more effective transport of coarse sediment into overbank environments than would be expected from standard flow-splitting models (Piper and Normark, 1983; Peakall et al., 2000).

Further investigation of turbidity-current interaction with channel bends reported in Straub (2007) shows the importance of bends in vertically mixing the suspended sediment within the interiors of turbidity currents. Results from two experiments compare current velocities, deposition rates, and deposit composition (grain size) in a straight versus moderately sinuous (sinuosity = 1.32) channel. He released 10 turbidity currents into both channels, each with constant initial height, fluid discharge, and excess density. Vertical sediment-concentration profiles collected at the centerline of each channel at the same distance from the current source show that currents moving through the straight channel become more stably stratified than currents moving through the sinuous channel. This resulted in higher near bed concentrations and therefore higher deposition rates for currents in the straight channel at equivalent distances from the current source. Associated vertical particle-size profiles also reveal that coarser sediment is suspended higher up in currents moving through the sinuous channel. Evidently, channel bends enhance large-scale vertical mixing in turbidity currents. This mixing helps maintain relatively high suspended-sediment concentrations in the current interiors, and consequently the excess-density structure necessary to drive the currents down slope. Straub (2007) hypothesized that wholesale vertical mixing of currents induced by channel bends aids increased runout of turbidity currents in sinuous submarine channels.

Construction of channel-margin levees has been a research focus in the submarine community in recent years (Skene et al., 2002; Dennielou et al., 2006; Straub and Mohrig, 2008). In net aggradational settings levees are the primary elements of self-formed submarine channels, yet little is known about their morphodynamics. A sequence of experiments reported by Straub and Mohrig (2008) documented the evolution of levees built by the continuous overspill of flow from channelized turbidity currents in a straight channel. By coupling flow data to the time evolution of levee morphology they demonstrated that the most important parameters controlling levee morphodynamics are the degree of channel confinement and the vertical structure of suspended-sediment concentration profiles. These observations, coupled to observations of levee morphology from offshore Brunei Darussalam provided the basis for constructing a levee growth model that couples a simple advection settling model for currents with a vertical sediment concentration profile defined by the Rouse equation. The model reproduces the evolution of levee slope with channel relief for their experimental data set. In addition, under a reasonable combination of current thickness, velocity and suspended grain diameters, the model reproduces the evolution of levee morphology for the Brueni Darussalam field site. The experiment by Straub et al. (2008) described above also was used to examine the growth of levees in sinuous channels. Levee growth was greatest at the outer banks of channel bends, helping to preserve the integrity of the channel form to act as a conduit for future flows.

3.6.3. Turbidity currents in intraslope minibasins

Intraslope basins (minibasins) associated with salt diapirism are a dominant morphological feature on many continental slopes. These features have received attention in the deep-water experimental community in the last decade due to their importance as containers of sand-rich deposits that constitute prime targets for oil exploration. Minibasins form from diapirism driven by buoyant instability of a mobile substrate (e.g. a salt body) overlain by a load of denser sediment. The basins typically have several hundred metres of relief and span areas on the order of 10^1–$10^2\,km^2$. Many, but not all, minibasins are connected to one another by submarine channels that form a drainage network that eventually discharges onto the abyssal plain. In many cases, turbidity currents are believed to have filled minibasins

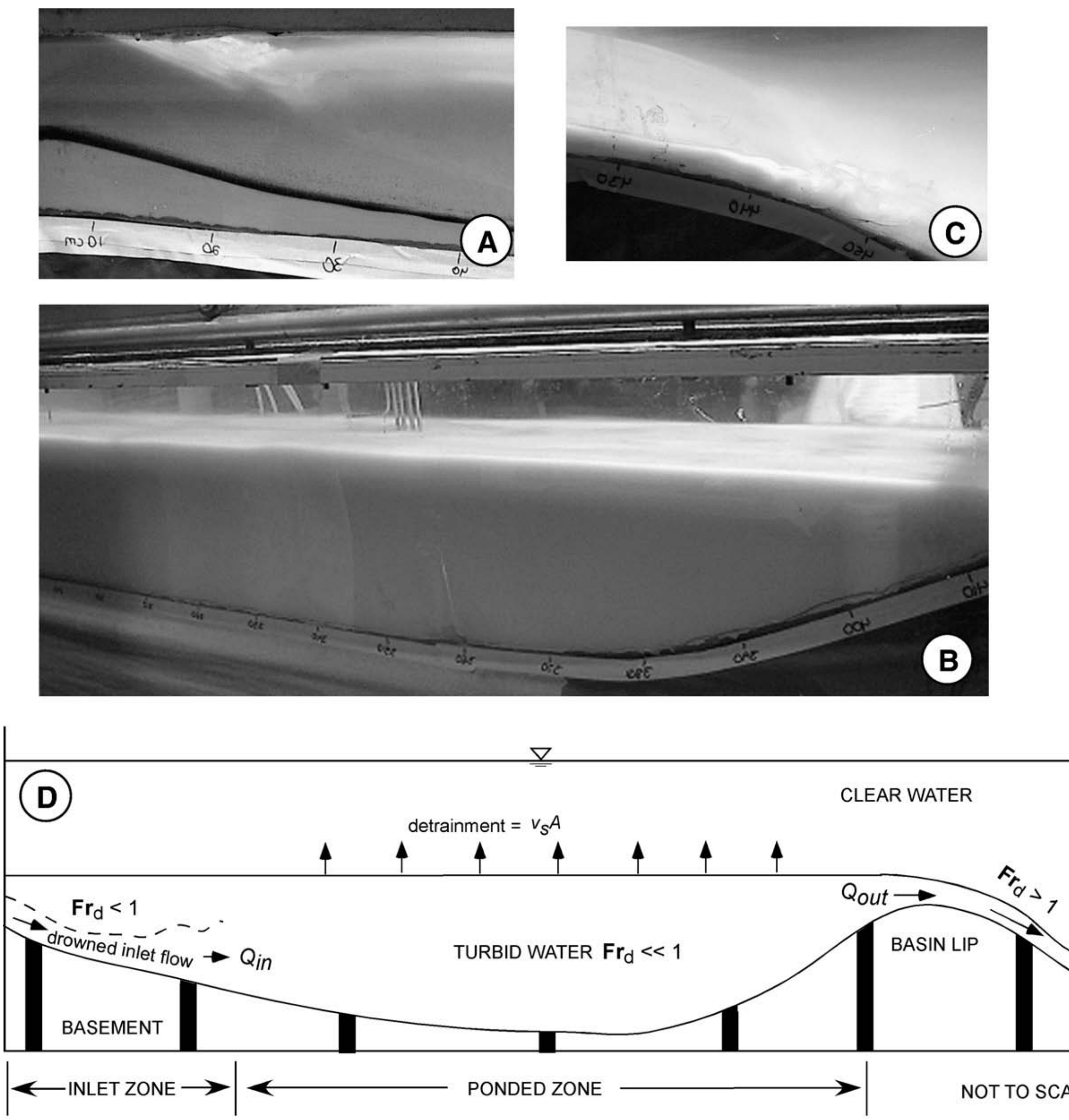

Fig. 27. Photographs and schematic illustration from experiment reported in Lamb et al. (2006) after development of quasi-steady ponded flow in a minibasin. Flow direction is from left to right and the scale along the flume bed is in cm. (A) Photograph of the inlet zone from approximately 0.05 to 0.45 m from the head gate showing mixing between the bottom inlet current and the turbid water above. (B) Photograph of the ponded zone from approximately 3 to 4 m from the head gate showing the glassy settling interface. Note that the fluid above the interface was sediment-free water, not air. Also visible is a rake of siphons used to sample the flow. (C) Photograph of turbid flow going over the basin lip from approximately 4.3 to 4.6 m from the head gate. (D) Schematic diagram of an idealized sustained turbidity current after set-up of ponded flow. Q_{in} is the discharge of turbid water entering the ponded zone and Q_{out} is the discharge of turbid water out of the ponded zone. The ponded turbidity current drowns the head gate, creating a drowned underflow ($Fr_d<1$) in the inlet zone. In the ponded zone, the turbidity current is highly subcritical ($Fr_d<<1$) and detrains water across a setting interface at the rate w_sA, where w_s is the settling velocity of the sediment and A is the surface area of the ponded zone. The turbidity current is critical ($Fr_d=1$) at the lip of the basin and becomes supercritical as it accelerates out of the basin.

and eroded channels into neighboring ridges through a process known as "fill and spill". In the "fill and spill" model the degree to which past turbidity currents have deposited sediment and filled one slope basin influences the flow characteristics of turbidity currents in basins immediately downslope.

Given this tendency of minibasins to form linked transport systems, it is not surprising that experimental work has emphasized the input–output characteristics of minibasins. In the first detailed experimental study of turbidity-current interactions with minibasins, Brunt et al. (2004) focused on flow and deposit characteristics in proximal and distal basins separated by a sill. As the ratio of flow thickness to sill height decreased, mean deposit grain sizes increased in both the proximal and distal basins, but for different reasons. Increasing flow thickness relative to sill height in effect reduces the degree of confinement of the flow. Progressively greater proportions of coarser sediment bypassed downstream as the degree of confinement was reduced. The mean grain size retained in the upstream basin also increased as more fine-grained material bypassed to the downstream basin as confinement was reduced.

A novel approach to the minibasin input–output problem was developed at St Anthony Falls Laboratory (SAFL) by Gary Parker and his research group. The idea was that through settling effects alone a minibasin could trap all or part of the input sediment flux, reducing the mean size and total flux of the outflow. In the limit of high settling velocity, low input flux, and large basin area, the input current could be trapped completely. The theory is scale independent and thus well suited to experimental testing (Lamb et al., 2006).

At the scales of individual flows, the topography created by variable subsidence makes the problem of flow–topography interaction discussed in the last section especially relevant to minibasins. Due to their reduced relative density, it is easier for turbidity currents to reach supercritical conditions than it is for subaerial channel flows. The change from supercritical to subcritical conditions creates hydraulic jumps, which dramatically increase mixing and could influence stratal geometry where they form. Lamb et al. (2004) and Toniolo et al. (2006a) found, through a series of experiments on the relationship between flow dynamics of turbidity currents and the stratal architecture of minibasin deposits (Fig. 27), that deposit shape was influenced by internal hydraulic jumps formed when turbidity currents thin cross inter-basin sills. These hydraulic jumps can initiate upstream migrating bores that stabilize at an upstream location determined by entrance flow conditions (Toniolo et al., 2006a,b). By thickening flows, thus reducing vertically averaged flow velocities, the jumps also enhance the trapping effect described above (Lamb et al., 2006). In addition Lamb et al. (2004) found that surge-type flow events with durations less than the amount of time needed to stabilize the longitudinal location of a hydraulic jump resulted in more ponding of deposits than continuous flows. When scaled to field conditions, the critical flow duration was estimated at 1 hour.

The only minibasin experiment to date in which turbidite sedimentation was accompanied by active subsidence is reported by Violet et al. (2005). The experiment, done in the XES facility described in Section 3.2.1, comprised three stages. In the first, stage, the basin subsided steadily in a bowl-shaped pattern with steady supply of a mixture of sand and silt; in the second stage the supply remained the same but subsidence was stopped; and in the third, the subsidence remained off and the sand in the supply was eliminated. Contrary to expectations, the subsidence in the first stage did not appear to influence sedimentation patterns relative to the second stage. Rather, the primary control on deposit geometry was the sand content of the input flow: reducing this in the third stage caused migration of the depocenter toward the low point in the topography. The sand-poor input currents bypassed the updip portions of the deposit and transferred sediment downdip. The limited control of subsidence pattern and/or rate on deposit architecture likely resulted from the overall highly depositional nature of the flow events in this experiment. Future experiments using lower density sediments might reveal stronger couplings between subsidence and depositional geometry for near-bypass flows. Kubo et al. (2005) compared results of the numerical SedFlux flow/transport model developed by James Syvitski and colleagues at the University of Colorado to the results of this experiment. Once the time step in the model was shortened to reflect experimental time scales, the model output compared well with that observed in the experiment. This is noteworthy for being a rare case in which experimental findings have been compared to an independent model, not developed by or in collaboration with the experimental group.

3.6.4. Submarine debris flows

Debris flows are the second major class of submarine mass flows. They are distinguished from turbidity currents in that debris flows comprise subequal parts debris and water, whereas the debris fraction in turbidity currents is $\ll 1$. Roughly speaking, flows with mean volumetric sediment concentrations $C_s < 10\%$ are turbidity currents, and flows with $C_s > 40\%$ are debris flows. Transitional conditions are possible; relatively dilute debris flows verging on turbidity currents are referred to as "weak", a condition that overlaps with so-called "high density" turbidity currents.

Since our theme in this review is experiments focusing on relatively large-scale morphodynamics we must pass over the extensive literature on debris-flow rheology, though it includes a number of interesting experiments. There is, however, one major development that bears on the role that debris flows play in constructing large-scale submarine topography, including submarine fans. On the face of it, debris flows, especially relatively strong ones, would be expected to be highly friction-dominated and thus to be incapable of moving on low slopes. Thus they could play a role in constructing submarine topography in, at most, only the relatively steep upper regions of submarine fans. But in a series of experiments comparing the motion of subaerial and submarine debris flows, Mohrig et al. (1998, 1999) showed that an entirely new phenomenon can occur in the submarine case that allows debris-flow movement on low slopes. The phenomenon is hydroplaning, in which dynamic fluid pressure at the front of the flow becomes high enough to lift the flow off its bed, effectively eliminating friction and allowing the flow to glide for long distances as long as the dynamic pressure is maintained. The condition for hydroplaning is a balance between the immersed weight of the debris $(\rho_s - \rho_f)$, where h is the height of the debris flow, and the dynamic pressure, which scales as $\rho_f U^2$. Combining these we see that the onset of hydroplaning is controlled by the densimetric Froude number (Fr) defined in Section 2.1.1. It is noteworthy that although long runout of debris flows onto low submarine slopes had been known for some time from the field, the hydroplaning mechanism for explaining it was observed first experimentally and explained theoretically afterwards. Once again, the capacity of experiments to surprise us is a fundamental motivation for doing them.

3.6.5. Summary and next steps: deep-marine systems

The deep-marine studies described above provide additional examples of the "unreasonable effectiveness" of experiments. As with experiments aimed at terrestrial geomorphology, the experiments are not exact dynamic scale models of submarine environments. The obstacles to full dynamical scaling are similar to those for rivers, but two problems are especially important in the submarine realm: 1) entrainment of ambient fluid into the body of turbidity currents, an effect absent in the terrestrial case, is controlled by turbulence in density flows and is thus difficult to scale; and 2) muds, which are hard to scale for reasons explained earlier, make up much of the seafloor and the sediment transported by turbidity currents. In spite of these limitations, the surface morphologies and deposit architectures reported in the experiments described above bear remarkable resemblance to natural systems (Metivier et al., 2005; Violet et al., 2005; Yu et al., 2006; Straub and Mohrig, 2008).

The work we report in this section represents mainly initial experiments — the renaissance of experiments on submarine morphodynamics is evidently still in its early stages. For instance, all three experimental studies on the formation of channelized fans report only “preliminary” findings (Metivier et al., 2005; Yu et al., 2006; Hoyal et al., 2008). Each group reported that their primary finding was the ability to generate channelized fans. Now that we know we can produce reasonable self-formed channel systems, the next step is to turn to fundamental issues such as the interplay of autogenic and allogenic processes. It will be interesting to see if the different experimental approaches used by each group (bedload dominated (e.g. Metivier et al., 2005; Hoyal et al., 2008) versus suspended-load dominated (Yu et al., 2006)) lead to similar or dissimilar seafloor morphodynamics.

Experimental research has also opened up a host of new questions in turbidity-current–channel interactions. Among the most pressing of these is how to produce turbidity currents that can navigate hundreds of channel bends without succumbing to the twin perils of deposition induced by coarse sediment load or entrainment of ambient fluid. Straub (2007) suggested a possible mechanism that could enhance runout distance through channel bend induced mixing of the suspended-sediment profile but this theory has yet to be fully tested and likely is not the only mechanism responsible for long runout of turbidity currents.

To date the standard experimental procedure for submarine mass flows (debris flows as well as turbidity currents) involves injecting some form of prepared mixture into a flume or tank. While this approach is undoubtedly an efficient way of repeatably creating mass flows large enough to measure and study, it does not shed any light on how the mass flows originate, or what sets the initial conditions that experimenters currently control directly. Exploration of the origins of mass flows is a new and potentially informative frontier for experimental research on submarine transport processes.

Finally, we note that submarine transport includes other processes unrelated to mass flow. The best known of these is transport by geostrophic currents, which are maintained by a balance of Coriolis and pressure forces. Because geostrophic currents generally flow parallel to pressure or topographic contours, they are often referred to as “contour currents” and their deposits as “contourites”. Although it is possible to produce geostrophic currents experimentally using rotating tanks, these facilities are expensive to build. To the best of our knowledge they have not been systematically used to study contourite sedimentation.

3.7. Summary

The main points emerging from the broad areas of erosional landscapes and stratigraphy have been summarized in their respective sections. Looking over the whole body of experimental landscape research, several themes emerge. The first and most important point has been made repeatedly above: experimental landscapes organize themselves in ways that are remarkably similar to what is observed in the field. Overall, we have done the best with erosional and coarse-grained systems. Small-scale experiments reproduce major planform and behavioral aspects of stream braiding and alluvial fans. Experiments also capture the planform structure of dendritic erosional networks; the extent to which they also capture kinematics is unknown because it is difficult to study landscape kinematics in the field. One interesting difference between experimental and field-scale erosional systems is the persistently low observed values of slope–area exponent in the experiments, leading to topographies deficient in longitudinal curvature. These are comparable to debris-flow dominated landscapes, but the effect does not seem to be caused by debris-flow dominance at experimental scales. Why is profile curvature scale dependent in a way that, apparently, planform structure is not? Another point is that the degree of autogenic landscape instability observed in the experimental systems varies significantly. The explanation is not clear – variable rainfall, or overall system geometry – but the more important question is what the level of variability is at field scales. This is unknown at present, but is a tractable if difficult problem for field researchers to tackle.

Depositional-system experiments allow study of the entire sedimentary recording process from surface dynamics to preserved strata. They reproduce all major elements of sequence stratigraphy, and have expanded and clarified our concepts of how sequences develop and what they mean. A major advance has been development of deformable-floor facilities, which provide full control of the last element of the “stratigraphic trinity” of sediment supply, eustasy, and subsidence. These systems have allowed evaluation of qualitative and quantitative theories for the interplay of subsidence and sedimentation, including channel stacking, shoreline dynamics, and the influence of subsidence on sequence stratigraphy. Deformable-floor systems also open the way to studying the interaction of tectonics and sedimentation, for example via the time-scale arguments outlined above. These first steps toward prediction also provide an example of new ideas arising from experiments that can be tested in the field. Work has begun on experiments that couple complex structural deformation to evolving surface morphology, and that couple erosional and depositional realms. We expect rapid development in these areas in the coming years.

There are many aspects of even relatively simple, abiotic experiments with only sand-size and coarser particles that we are still a long ways from being able to predict. Nonetheless, as we have seen, progress is being made to overcome the twin obstacles of cohesive-sediment dynamics and vegetation. Research on both fronts has led us to the verge of resolving one of the persistent failures of experimental landscape research: the inability to create self-sustaining high amplitude river meandering. Another major frontier, now accessible with experiments that combine vegetation and suspended sediment, is the mechanisms of floodplain creation and maintenance.

A final broad theme emerging from the experiments reviewed here is the diversity and importance of autogenic dynamics and other forms of nonlinearity in morphodynamic systems. Specific internally generated events, such as connection events, autobreak and autogenic detachment, stream capture, avulsion, and changes in channel morphology, cause significant and often rapid changes in system behavior and sediment delivery. Autogenic dynamics creates persistent variability that is part of the stratigraphic and landscape record. To what extent is this variability stochastic? Can it be formally shown to be deterministic chaos, and if so can we quantify its degree of predictability? So far, the power of experiments to provide quantitative, well constrained data sets has led to a good start in describing, and some progress in quantifying, many of these autogenic phenomena. But we still have a ways to go before we have a full set of predictive tools for stochastic geomorphic and stratigraphic processes. Better definition of the length, time, and amplitude scales of autogenic processes will underpin the statistical mechanics of landscapes, and also help differentiate autogenic from externally caused effects. In addition, we have only begun to explore the interesting and highly nonlinear mechanisms by which autogenic and allogenic processes interact. This is clearly an attractive area for future research.

4. Small worlds and large worlds: scaling revisited

The principal finding of our review is that stratigraphic and geomorphic experiments work surprisingly well. By this we mean that morphodynamic self-organization in experiments creates spatial patterns and kinematics that resemble those observed in the field. Nonetheless, as nearly all experimenters are at pains to point out, all of the experiments fall well short of satisfying the requirements of dynamic scaling discussed in Section 2.

Similarity in landforms and processes in the absence of strict dynamic scaling is what we mean by "unreasonable effectiveness". It would hardly be surprising if perfectly scaled experiments reproduced the form and behavior of natural systems. So the obvious question, which will be the focus of this section, is: what does the unreasonable effectiveness of laboratory experiments on landscape dynamics tell us? Can we move beyond the rigorous but limiting confines of classical dynamical scaling to develop a better basis for designing and using laboratory experiments? Can we develop a quantitative way of analyzing to what extent an experiment that looks like a field system really is like the field system?

Our starting point is a comment written by Roger Hooke more than forty years ago (Hooke, 1968). He proposed the idea of "process similarity", which amounts to saying that the same processes seem to be at work in laboratory experiments as in the field. In recent years this idea has been expressed using terms such as "analog model", again implying some level of process analogy with field systems that falls short of full, rigorous scaling. Here we offer an extension and elaboration on Hooke's basic idea, using ideas that had not been developed when his paper was written, as a step towards a formal basis for process similarity.

The aggregate implication of the success of unscaled geomorphic and stratigraphic experiments is that important aspects of morphodynamics are scale independent, over a wide range of scales. Hence our goal is to refocus the discussion of geomorphic experimentation away from formal scaling and toward the causes, manifestations, and limits of scale independence. As we will see, scale independence in nature has many dimensions. One of the most important of these is that many natural systems advertise the presence of scale independence via self-similarity.

Barenblatt (2003) provides a rigorous analysis of the relation between dynamic scaling in the classical engineering sense, and scaling in the sense of self-similarity and scale independence. The former traces its roots to 19th century fluid mechanics, while the latter bloomed in the late 20th century following the introduction of fractal geometry by Mandelbrot (1982). In our case, the idea is to use concepts of scaling and similarity to help understand the relation between small-scale laboratory experiments and the field. But before going any further, it will be helpful to have more precise terminology for defining and analyzing similarity, which can take a variety of forms.

4.1. Kinds of similarity

The terminology we propose extends existing terminology, which is included for completeness. In addition to the analysis of Barenblatt (2003), we have used the work of Sapozhnikov and Foufoula-Georgiou (1995, 1996b), who proposed the idea of internal and external fractal-scaling exponents.

4.1.1. Similarity and affinity

Similarity means that two systems have similar properties under scale transformation. Affinity is a generalization of similarity that allows for distortion by a constant multiplicative factor. Two squares of different sizes are similar; a small square and a large rectangle are affine. Unless otherwise noted, affinity may be substituted for similarity in any of the definitions below, implying that the scale transformation is accompanied by stretching by a constant factor. The vertically distorted Froude models mentioned in Section 2.1.1 are examples of experiments that are affine to their field counterparts.

4.1.2. Geometric, kinematic, and dynamic similarity

Geometric refers here to form, kinematic to motion, and dynamic to forces. An architectural model is geometrically similar to the building it represents, and an electric model steam engine could be said to be kinematically similar to a full-scale steam engine. Neither is dynamically similar to its prototype. In each case, the level of similarity restricts the extent to which an observed value of some variable in the model can be converted to the equivalent value of that parameter in the prototype (or vice versa). For example, dynamic similarity implies that any force measured in the model could be converted algebraically to an equivalent value in the prototype. The three categories are hierarchical in the order given above, i.e. dynamic similarity implies kinematic and geometric similarity, and kinematic similarity implies geometric similarity. Informally, we might say that systems showing geometric similarity look like one another, those showing kinematic similarity act like one another, and those showing complete dynamic similarity are mechanistically like one another.

4.1.3. Exact and statistical similarity

Mathematically constructed shapes, including fractals, can show exact similarity. The usual case in nature is statistical similarity, in which the appropriate scaling transformation produces two systems that are statistically indistinguishable, but not identical.

4.1.4. Internal and external similarity

This is the most important similarity idea for this paper. It is an extension of the distinction between internal and external fractal scaling proposed by Sapozhnikov and Foufoula-Georgiou (1995, 1996b). If a small part of a system is similar to the whole system, we say that the system shows internal similarity, which is usually referred to as self-similarity. If a small version of a large system is similar to the large system (in any of the senses discussed above), the two systems show external similarity.

4.1.5. Natural and imposed similarity

Natural similarity occurs spontaneously, while imposed similarity is imposed by design.

These definitions allow us to describe more precisely the various kinds of similarity we find in human constructions and in nature. An electric model of a steam locomotive shows imposed, external, kinematic similarity to a real steam locomotive. Fluid-dynamics experiments fully scaled using the classical methods described in Section 2 are examples of imposed, external, dynamical similarity. Mathematical fractals such as the Koch 'snowflake' are examples of exact, imposed, internal, geometric similarity. The mathematical forms took on new importance when fractals were found to be common in nature; these are examples of natural, internal, geometric similarity. The kinematic similarity discovered in braided rivers by Sapozhnikov and Foufoula-Georgiou (1997) is an example of natural, internal, kinematic similarity. All of the natural examples cited above show statistical rather than exact similarity; unless otherwise specified we will likewise assume statistical similarity throughout the discussion to follow.

The goal of stratigraphic and geomorphic experiments is *to achieve the maximum degree of external similarity between the experiment and the field*. The highest level of similarity is complete dynamic similarity. One sure way to achieve this is through imposed dynamic similarity, using the techniques summarized in Section 2. However, as we saw in Section 2.1.3, a variety of fundamental limitations make complete dynamic scaling practically impossible. Thus the rest of this section is focused on exploring examples of natural similarity and scale independence in nature, and on the possibility of using these as a basis for experimental design.

There is at present no "theory of scale independence", but Barenblatt's (2003) work gives us some ideas to start with. He points out a fundamental connection between the process of creating dimensionless variables for a given system and scaling as an emergent property of certain systems. Dimensionless variables are multiplicative power–law functions of basic dimensional variables. The requirement of dimensional homogeneity constrains the exponents in these functions. Once the dimensionless numbers are constructed, however, any mathematical relation among them is possible, since they are already nondimensional. In certain cases, the relation of the dimensionless variables to one another is itself a power law, i.e. it mimics the relations within the dimensionless variables themselves. Barenblatt terms this case *incomplete similarity*; it is a property of some systems and processes. Because of the scale invariance of power–law functions, natural fractals fall into the category of incomplete similarity. However, not all systems with power–law properties are fractal.

4.2. Scale independence

By "scale independence" we mean that the important dynamics of a system are independent of scale over a significant scale range. Scale independent systems are amenable to experimental study at reduced scale without recourse to classical dynamical scaling. Scale independence and similarity are not interchangeable, though they are obviously related; similarity could be said to be the outward manifestation of scale independent dynamics.

Where scale dependence arises from a single physical effect that can be embodied in a dimensionless number, one might expect scale independence in the limit as the number tends to infinity or zero. For example, the Weber number introduced in Section 2.1.1 measures the magnitude of surface tension forces relative to fluid inertial forces, and when the former are small relative to the latter, they can be ignored. Comparing two cases of differing scale, as long as the value of *We* is high enough in both, the surface tension is irrelevant and it does not matter if the *We* values match.

Applying the same argument to the Reynolds number *Re* leads to the classical inviscid-flow solutions treated in fluid dynamics texts. Inviscid theories work well for phenomena like surface gravity waves for which viscosity really is unimportant: gravity waves involve a balance between pressure and gravity forces and fluid accelerations, neither of which involves viscosity. Surface gravity waves are scale independent over a very wide scale range, and as such have been studied successfully in laboratory experiments for many years. (This makes it all the more surprising that wave processes how little recent experimental morphodynamic work has been done involving waves.)

In applications where there are interactions between the flow and solid boundaries, viscous effects can never be eliminated, no matter how large *Re* becomes. But many high-*Re* flows nevertheless show some form of *Re*-independence, which we introduced in Section 2.1.3. Two examples are the relatively constant values of drag coefficients over a wide range of *Re* values, and the ubiquity of the logarithmic vertical velocity profile in ordinary turbulent boundary layers, independent of *Re*. Since *Re* is one of the important parameters directly influenced by scale changes, *Re*-independence is a form of natural scale independence. It is intimately related to another observed property of turbulence, which is that it comprises a hierarchical, fractal structure in which energy is passed from large scales via a cascade of self-similar structures (eddies, loosely speaking) to fine scales where it is dissipated to heat. Thus turbulence shows both external and internal similarity, both of which are linked to *Re*-independence.

Re independence is an implicit element in the design of landscape experiments, but it can be fully justified only for coarse-grained systems and modest reductions in scale. More importantly, *Re*-independence is a particularly well documented example of "unreasonable effectiveness" in scaled-down experiments, i.e. experiments that produce useful results despite failing to satisfy the requirements of classical scaling. What can we learn from it?

The physical basis for *Re* independence is fairly well understood. Broadly speaking it arises because the cascade of energy from large scales to small scales of turbulence results in a separation of the dynamics of the large scales from that of the small scales. For two otherwise similar flows, increasing *Re* decreases the characteristic size of the energy-dissipating scales without changing the behavior of the large scales, and it is the latter that are primarily responsible for the overall dynamics. Thus, a key characteristic of turbulence is the cascade that separates the large-scale and small-scale dynamics. The large-scale aspects of turbulent flows are thus intrinsically insensitive to scale. Note that the self-similarity of turbulence across scales is not the result of some sort of self-organized consistency in *Re* across scales; moving down the hierarchy from large to small scales the *Re* value continuously decreases towards 1, its value at the finest (Kolmogorov) scale.

Various authors have proposed analogies between turbulence and landscapes (Paola, 1996; Passalacqua et al., 2006; Haff, 2007). Here we use the analogy only to point to scale separation as one explanation for the "unreasonable effectiveness" of scaled-down landscape experiments. In turbulence the scale separation is effected via a self-similar cascade, in which the main dynamical element (eddies) reproduce themselves across a range of scales — a case of natural internal similarity (i.e. a fractal). So we will turn next to examples of geomorphic systems that also show natural internal similarity, as an indicator of natural scale independence.

4.3. Natural internal scaling in landscapes

Given that the type example of a fractal is a geomorphic feature (the coastline of Britain), it is not surprising that there are many examples of natural internal scaling in geomorphology. The best known cases of fractal geometry and associated power laws derive from erosional landscapes. The measures that have been developed to analyze fractal geometry and the many ways in which they apply to natural erosional landscapes are thoroughly covered in Rodriguez-Iturbe and Rinaldo (1997). Overall, the case for internal geometric similarity in erosional landscapes is very strong. Current methods for measuring erosion rates do not yet allow us to measure spatial variability in erosion rate for evidence of kinematic scaling, and as yet the available experimental data have not been analyzed for this either.

Evidence for fractal geometry (internal similarity) is present but less well developed in other systems. The pattern of channel division and confluence that defines braided rivers appears to repeat itself over a range of channel sizes. Sapozhnikov and Foufoula-Georgiou (1996b) quantified this geometric scaling in braided rivers. The scaling appears to extend over at most two orders of magnitude in scale range — much less than the range for erosional landscapes. Fractal scaling in nature is always bounded by physical limits, but what sets these in morphodynamic systems is still under investigation. The upper limit of the scaling range is presumably related to observed bar-confluence length scales (Ashmore and Gardner, 2008).

As mentioned above, Sapozhnikov and Foufoula-Georgiou (1997) added a new dimension to scaling analysis by showing that the steady-state autogenic dynamics of braided rivers exhibits power–law scaling in space and time — to our knowledge, the first-ever study of kinematic scaling. The basic idea is that if one has movies of a braided river taken over different size regions, represented by a length-scale ratio L_1/L_2, they could be rendered indistinguishable by a change in time scale (i.e. changing the projection speed of one of the movies) represented as T_1/T_2. The power–law hypothesis is that $T_1/T_2 = (L_1/L_2)^a$ where a is a scaling exponent, empirically found to be equal to about 0.5.

The case for self-similarity in meandering rivers is not as clear. Stolum (1996) proposed that meander growth and cutoff is a self-organized critical process as defined by Bak et al. (1987). That would imply similarity at least up to the kinematic level, but Stolum's claim has not received widespread support. There seems to be little independent evidence for fractal geometry in meandering rivers, and the idea is fundamentally at odds with observational evidence that meandering has a well defined characteristic length scale of 7–10 times the channel width (e.g. Blondeaux and Seminara, 1985).

The successive bifurcation of distributary channels in deltas would lead one to expect some level of scaling and internal similarity in deltas as well. Surprisingly, this topic has not been extensively developed. At this point we can only speculate that the extent to which deltas show internal similarity must depend on the main sediment-distribution processes. Alluvial fans also have distributary networks in some cases, but to our knowledge it is not known if these channel networks are fractal or not. Finally, one case of a channelized depositional system that has been thoroughly analyzed and is apparently not fractal is tidal marshes (Rinaldo et al., 1999). Summing up, self-similarity seems to be well developed in erosional channel networks, present over a limited range in braided rivers, and possibly present, to an unknown extent, in other systems such as alluvial fans and deltas.

4.4. Does internal similarity imply external similarity?

For our purposes, internal similarity is important as an indicator of external similarity, the condition we desire for landscape experimentation. What is the connection between internal and external similarity? It is striking how often researchers have naturally gravitated to doing unscaled or partially scaled experiments on systems whose field-scale counterparts show well developed internal similarity. This observation leads us to ask how these two forms of similarity are related. As Sapozhnikov and Foufoula-Georgiou (1995) point out, researchers have assumed that external similarity implies internal similarity by using, for example, using fractal measures from forms of different sizes (external measures) to estimate the internal fractal properties of the form (internal measures). Sapozhnikov and Foufoula-Georgiou go on to show by example that estimating internal measures from external ones is not generally correct. For our purposes we do not need strict equivalence between external and internal exponents. We simply propose that, since internal similarity and external similarity are both manifestations of scale independence, *internal similarity implies external similarity*. To the extent that a small part of a large system behaves like the large system, then a small copy of the large system should behave like the large system. Considering all the natural systems reviewed in the previous section, we find no exceptions to the observation that internally similar systems also occur naturally over a wide range of whole-system scales.

External similarity is the fundamental condition for relating unscaled experiments to field-scale systems. If internal similarity implies external similarity, and both are manifestations of scale independence in the governing processes, then internally similar (fractal) natural systems are natural targets for experimentation. Formal dynamical scaling in these cases is no longer crucial: natural systems that display internal similarity are, in effect, advertising scale independence through their very structure.

At present, we have much to learn about what similarity in nature really means. Despite research into the physics of fractals, the field remains focused on phenomenology and geometry. For this reason we cannot provide a rigorous proof of our conjecture that internal similarity implies external similarity. Observationally the two appear to be linked, but like any scientific hypothesis this can never be proved by observation — one well documented counterexample is enough to disprove it. In this case a sufficient counterexample would be a system that is internally similar (fractal) over a range of length scales that is substantially greater than the range of overall (system) length scales over which it shows external similarity.

4.5. External similarity does not require internal similarity

Internal similarity appears to imply external similarity, within defined scale ranges. However, the converse is clearly not true, i.e. external similarity need not imply internal similarity. For instance, within broad limits, the theory of ordinary surface gravity waves shows them to be scale independent, and hence they show external similarity over a broad scale range. Yet although surface gravity wave fields are sometimes fractal, they often are not — for example, ordinary ocean swell is not fractal.

Likewise, external similarity in morphodynamic systems is not limited to systems showing fractal geometry. A good example has been provided by the clastics research group at ExxonMobil, which proposed similarity in the shapes of depositional bodies across a wide range of scales and depositional environments (Hoyal et al., 2003). The underlying physical idea is that depositional-body geometry is set by the form of expansional jets, whose well known external similarity is a good example of *Re* independence. Because the idea has been published only as an extended abstract it is difficult to evaluate it in detail, but we stress that the examples provided show only external similarity. Schlager (2004) has likewise emphasized the scale invariance of stratigraphic sequences, and attempted to link this external similarity to observations of internal similarity in stratigraphy. However, the only well documented example of internal similarity in stratigraphy to date is the distribution of time gaps in vertical sections (Jerolmack and Sadler, 2007; Plotnick, 1986; Sadler, 1999). At this point it is not clear to what extent depositional processes create internally similar (fractal) morphology, but it is evidently less common than is the case for erosional processes, and internal similarity is clearly less common in depositional systems than external similarity is. Why this should be is an open question that awaits a better understanding of the physical basis for similarity.

A major advance in understanding to scale independence in morphodynamics is the work of Lajeunesse et al. (in press) who show that a wide range of morphodynamic processes work similarly in laminar as compared with turbulent flows. Lajeunesse et al. termed this “convergence of physics”. Though it is not scale independence *per se*, the laminar-turbulent transition is the most dramatic change one sees as a function of flow scale. The small-scale (laminar) flows are clearly and qualitatively different from large-scale turbulent flows, but the dynamics that matters – the relation between shear stress and topography, and that between shear stress and bedload flux – is similar enough that the morphodynamics is surprisingly consistent across this major, scale-dependent transition. Similarity across the laminar-turbulent transition is a dramatic example of external similarity. Some of the systems (e.g. braided streams) considered by Lajeunesse et al. (in press) have been shown elsewhere to have fractal properties but others have not; in general the results of this study illustrate the principle that internal similarity is not a necessary condition for external similarity.

Going up in scale, consider the response of the experimental XES system to a change in base level discussed in Section 3.2.3. The river channel networks are braided, but the general dynamics of valley formation – incisional convergence of the flows, downcutting accompanied by lateral valley erosion – are not dependent on the details of the river pattern. The experimental geomorphic evolution and stratal patterns are fundamentally similar to those seen at field scales even where the river pattern is not braided. These examples illustrate an important contributing mechanism to external similarity: insensitivity of the dynamics at the scales of interest to the details of the behavior at smaller scales (Werner, 1999). It is as if the small-scale behavior were communicated up the scale hierarchy through a screen, so that only its general form could be discerned. The fluid and sediment processes that are the focus of classical dynamical scaling occur at the fine-scale end of most landscape experiments, so insensitivity to fine-scale dynamics translates to scale independence and insensitivity to the strictures of classical dynamical scaling.

Both of these examples illustrate an important general point: morphodynamics, though clearly linked to fluid dynamics, is not a branch of fluid dynamics. Fluid flow, which often appears to be the main focus of morphodynamics researchers, is only part of the picture. Based on the examples above, it may be a relatively small part. The addition of granular material, which has its own highly complex mechanics, to flowing fluid produces a system that is capable of a remarkable range of self-organized behavior. This behavior arises from the fluid–sediment interaction, not from the fluid flow *per se*. Although fluid flow is clearly necessary to get things going, it may well be that once the bed is activated, morphodynamic evolution takes on a life of its own that is sensitive only to general properties of the flow. This again helps explain why getting the fluid-flow regime correct may not be critical for successful landscape experiments. (In this context it is fortunate that the classical dimensionless parameter most closely tied to sediment dynamics is τ_*; as discussed in Section 2.1.2, τ_* values typical of the field are relatively easy to reproduce experimentally.)

Summing up, the "unreasonable effectiveness" that this and previous reviews demonstrate is a manifestation of scale independence that, in various forms, is well developed in morphodynamic systems. Scale independence is connected to the widespread occurrence of fractals and other kinds of similarity in natural landscapes, in ways that we do not fully understand. Beyond providing a basis for landscape experiments, scale independence is an important research topic in its own right; after all, natural landscapes vary over a greater scale range than the difference between typical experiments and small natural systems. We stress that information about scaling and scale dependence can be obtained by comparing systems of different sizes only to the extent that the small system is *not* an exact dynamic scale model of the large one. This provides an opening to turn the "scaling problem" into an advantage — a source of insight about scale dependence and independence. We cannot create true scale models or exact analogs. But this is necessary only if we intend to apply the experimental results to the field wholesale, through some form of direct algebraic transformation. Abandoning this naïve idea, and freeing experimental landscape research from its death-grip on the control wheels of dynamical scaling, opens the way to using experiments to study scale dependence and independence, in addition to other possibilities that we will turn to in the next section.

5. Synthesis, strategies, and future of landscape experiments

5.1. Synthesis

Overall, several key observations arise from our review of experimental stratigraphy and geomorphology over the past two decades. The first is simply how diverse the experiments are, in terms of facilities and materials used, locations and groups involved, and questions asked. There is far more experimental landscape research going on around the world now than there was twenty years ago. The second observation explains the first: it is the capacity of even very simple experiments to create complex forms and behaviors spontaneously, and thus to surprise and inform us. We invite the reader to flip through the images in Section 3, and then consider the fact that every one of the complex, familiar, and yet puzzling patterns shown began from a simple shape: a flat surface; a block of sediment; a step. In some cases, the forcing is steady, and in others variable. Some of the experiments are influenced by tectonic effects, many are not. But in no case is the complexity of the input conditions commensurate with the complexity of the experimental results. Taken as a whole, the landscape experiments summarized here are a testimonial to the capacity for self organization of systems of fluids and particles. It is not surprising that no numerical model is yet capable of reproducing these results in detail. But it is surprising that so little has been done to use experimental results to develop and refine numerical landscape models. We are encouraged by the increasingly quantitative bent taken in experimental landscape research, but there is a long way to go. In many cases, the interpretation of the results remains descriptive, and is done in an analog rather than an analytical framework.

We are fascinated by the mix of structure and randomness shown by so many of the experiments reviewed in Section 3. The stochastic component of experimental landscapes is internally generated (autogenic) — to the best of our knowledge no one has yet really looked at stochastically forced landscapes experimentally, though in our view this would be well worth doing. To all appearances what has been observed so far constitutes deterministic chaos (in the original broad sense of stochastic behavior from a deterministic system), but little has been done to analyze this quantitatively, and use it to investigate the limits to predictability in landscapes. The origin, nature, and limits of stochastic landscape behavior bear fundamentally on how past history is recorded in landscapes (Slingerland, 1990). For depositional systems, the key questions revolve around how external processes are recorded, how they interact with autogenic processes, and how best to disentangle them. The same questions apply to erosional systems, with the addition of the basic question of how long it takes net erosion to erase the memory of past events entirely. These questions are ideally suited to experimental study; indeed, the general question of how landscape systems process and record information certainly transcends scale.

We are also struck by the parallels between erosional and depositional systems that experiments reveal. One is the presence of a steady state in which, on average, the divergence of surface flux balances the net tectonic mass flow; another is the superposition of autogenic and allogenic processes. Both system types have large-scale mechanisms for abrupt lateral channel shifts: stream capture in the case of erosional systems, avulsion in the case of depositional ones. These and smaller-scale autogenic processes in both types of system are associated with the ubiquitous tendency of sediment to move through the landscape in steps via processes of local storage and release. This "stick-slip" behavior is associated with the threshold nature of sediment detachment and movement. In other types of systems (e.g. groundwater) storage-release processes give rise to effects such as "thick tailed" probability distributions, in which extreme events are more common than one would expect from, for example, a Gaussian distribution. These and other potential effects of sediment storage-release processes are well worth detailed exploration in which experiments should play a major role.

The potential of experiments to contribute to understanding scale independence is most apparent in studies aimed at the largest natural scales, where experiments with dimensions down to a few decimeters are being used to study processes on the scale of whole mountain ranges or sedimentary basins. Apart from measuring whole-system response to external changes in tectonics, sea level, and climate, another major theme at these scales is large-scale connections – teleconnections – among the parts of the system. The opening and closing of such connections can make prediction challenging (for example, teleconnections are a major problem in ENSO forecasting). In landscape dynamics, we know almost nothing about such potential large-scale connections, but their potential is illustrated by, for example, experimental findings discussed in Section 3 on the "connection time" required for evolving channels to link one part of a system with another. Having a whole system "in a box" is a powerful way to investigate what kinds of connections are possible and how they might work. At these large scales, the division between geomorphologists focused on erosional systems and stratigraphers focused on depositional systems also becomes increasingly artificial and problematic. In recent years researchers have ignored this division, devising experiments in which source and sink interact, as they do in nature.

At the scales of individual reaches and channels, experimental studies are shedding new light on topics ranging from flow and sediment dynamics in submarine channels to the origin of river meandering. It is striking that turbidity-current research has led the injection of quantitative and experimental methods into industrial research, and by extension, into academic stratigraphy. The extra degrees of freedom brought about by a small density contrast and mixing at the free surface appear to make turbidity-current dynamics subtle enough that qualitative methods fail even for qualitative questions. Meandering also links submarine and river research, and once again we are struck at the potential of submarine systems to inform river research, and vice versa. In an era of increasing specialization it is easy to overlook the expansion of insight that can come from changing a basic parameter like relative density. For instance, the reduced relative density of submarine flows exaggerates vertical scales and acts as a magnifier for key overbank processes like levee construction. Vegetation, which appears to be effective in encouraging meandering of rivers, is obviously absent in the deep sea — yet the channel patterns are very similar.

5.2. Strategies

At the end of Section 4, we argued for natural scale independence as a basis for designing and interpreting landscape experiments. This idea arises from the convergence of material properties that make complete dynamical scaling impossible with evidence for scale independence in morphodynamic systems that makes it unnecessary. But at present we do not have a complete physical basis for understanding scale independence in morphodynamics because we do not have the mechanistic governing equations that would be used to provide it. So we are left in the somewhat awkward position of proposing scaling and scale independence both as a basis for designing experiments and as a prime target for experimental research. We see no way around this. We will have to learn by doing: comparing field and experimental systems, using experimental results to develop new questions for the field, and using similarities and differences to understand how landscape processes depend on scale and material properties. At present, the limits to our understanding prevent us from formulating a full, rigorous natural-similarity basis for experimental design analogous to classical scaling methods. Instead we offer the following outline of a new framework for landscape experiments:

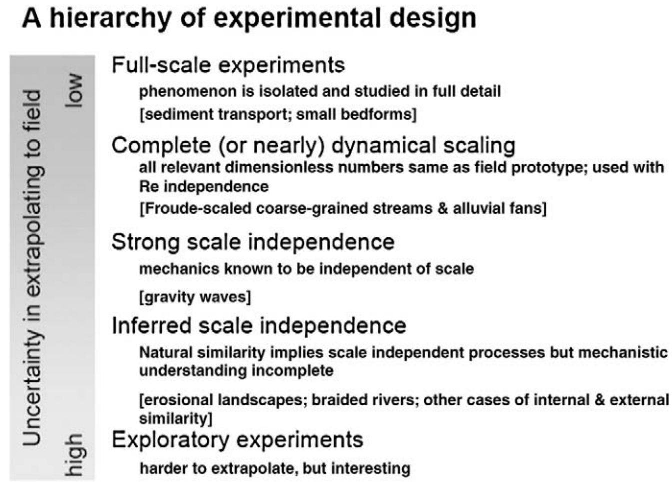

Fig. 28. Summary of design frameworks for morphodynamic experiments, arranged roughly in order of uncertainty in extrapolating the experimental results directly to the field.

(1) *Natural similarity provides a far more flexible and expansive framework for experimental design and interpretation than does imposed similarity via classical dynamical scaling.* This is the main point of the preceding section, where we presented a number of examples of natural external similarity, and proposed that internal similarity is a sufficient though not necessary condition for it. We remind our readers that even traditional engineering model studies often rely on a well known case of natural similarity: Reynolds-number independence. At present it is not known how widespread natural scale independence is in morphodynamics, but the evidence to date suggests it is common. In any event, we will not know until we have investigated further, and comparison of morphodynamic systems across scale is an obvious way to tackle this. This brings us to our next point:

(2) *Focus attention on understanding the origins and limits of scale dependence and independence in landscapes.* The work done since the 1960s on fractals and power laws in nature has taught us to think of scale as an independent variable in its own right. Understanding how processes change as a function of scale is fundamental to understanding how systems work, and also to developing a mechanistic basis for the various forms of similarity we have seen in this review. Freed from the rigid and unrealistic goal of providing exact miniature analogs of field systems, experiments are an obvious source of information on how dynamics changes with scale, both as small-scale end members and as a way of investigating scale changes systematically. (On a practical note, there is a scale gap between large laboratory experiments (few m) and small field examples (several km or more) where we have almost no data.)

(3) *Use large-scale dimensionless variables to place experiments in parameter space and put apparently different systems into a common framework.* Theoretical advances are providing a suite of new dimensionless variables that add to the classical ones aimed at local flow and sediment characteristics. The new dimensionless variables, examples of which were presented in Section 2.2, incorporate emergent system properties such as autogenic length scales and response times, as well as system-scale properties such as length and time scales, and overall mass balance. Dimensionless variables provide the common language for comparing superficially different systems, including ones that differ in scale. These variables must be chosen according to the problem at hand, but the papers reviewed here provide a number of examples that can serve as starting points.

(4) *Strengthen the connection between experimentation and theory.* We will not realize the full quantitative value of landscape experiments as applied to field systems until experiments and field studies are embedded in a web of quantitative models and analytical methods, with rapid cycling among all three elements of the triad: experiments, field work, and theory. Experiments are ideal for model testing because they offer full control of independent variables and the opportunity to measure exactly what is needed; likewise, a good theory, once tested, provides a means for evaluating the effects of changing scales or parameter values. It is surprising how little systematic comparison there has been between landscape models and theory, and we consider this an attractive research avenue. The four points above imply a final one:

(5) *Do not exaggerate the importance of scaling limitations*. Nearly all of the papers cited in this review, especially those aimed at the landscape and basin scales, include some form of disclaimer about scaling, which go as far as asserting that the experimental results are only of qualitative value. Although it is understandable that the limitations of imposed scaling and the lack of a full understanding of natural similarity lead experimentalists to be cautious, it is also no wonder that skeptics find it easy to dismiss the results altogether. The question of how to relate experiments to the field has been largely couched in terms of a framework developed more than a hundred years ago for the specific purpose of building analog scale models to predict the performance of engineered structures. This is not, or at least should not be, the goal of landscape experiments. We are instead investigating dynamic, self-organized patterns under controlled conditions, and testing theoretical models that engineers of a hundred years ago could scarcely have dreamed of. For these purposes, the classical engineering methods are simply one element in a spectrum of approaches that includes

natural similarity and newly emerging system-level dimensionless variables. These new ways of thinking about the effects of scale are being developed now, and they should be recognized as the exciting research opportunities that they are.

We return then to Hooke's (1968) suggestion that we treat experiments not as models or miniatures, but simply as small systems in their own right. We urge abandoning terms such as "analog model" and "physical model" as descriptions of morphodynamic experiments. A model is an idealization or theory about how nature works. An experiment is part of nature, however simplified or reduced in scale it may be.

Recognizing that researchers vary in their level of tolerance for uncertainty in the connection between experiments and the field, we offer in Fig. 28 a summary of experimental approaches in morphodynamics, with a rough guide to uncertainty in translating the experimental results to the field. Most of the experiments summarized in this review fall into the middle levels of this hierarchy.

5.3. Next steps

We discussed a number of potential growth areas for experimental landscape research in the previous sections. Here we add a few general areas not covered yet:

5.3.1. Replication and reproducibility

So far landscape experiments have been done by individual groups working within their own laboratories. Many of the experiments have been exploratory, and in a sense it is clear that even 20 years after the publication of *Experimental Fluvial Geomorphology*, (Schumm et al., 1987) the field is still finding out what it can and cannot do. Most experiments are run only once. Replication and reproducibility have not been systematically studied. It seems to us that the field has matured to the point where we should begin taking these issues seriously. Landscape experiments are time-consuming, so replication might be best begun through interlaboratory comparisons, which are common in fields like geochemistry. An important dimension of work on reproducibility of experiments is the inversely related question of "equifinality": can the same result, in terms of an erosional landscape or of stratigraphy, be obtained by two substantially different input scenarios? Most researchers seem to believe the answer is generally yes, but this has not been tested under controlled conditions.

5.3.2. Cohesion and life

Two major players that have so far been missing from most landscape experiments are biota and cohesive sediment. Both effects introduce new forms of scale dependence: for example, the ratio of surface area to volume goes up inversely with grain size so that cohesive effects inevitably become more important as the absolute grain size diminishes. Few would question that cohesive-sediment effects, from their influence on mass-flow rheology to their role in floodplain development and preventing re-entrainment of deposited sediment, are important in surface morphology. But the specifics of particle cohesion in the environment, influenced by geochemistry, compaction, flocculation, and various microbial effects, have been difficult to quantify. This has made it difficult to know how to include them in landscape experiments in a realistic way. On the other hand, recent work by Schieber et al. (2007) suggests that fine particles agglomerate and behave as larger, effectively noncohesive particles even without help from solutes or strong biotic effects. The gap between the dynamics of noncohesive and cohesive sediment may not be as great as we thought, and we thus may have overestimated the complexities of dealing with cohesive forces at reduced scale. Capitalizing on these new possibilities should be a high priority for experimental research in the years to come.

So far, no unmistakable signature of biota on morphology has been detected at scales greater than a few m (Dietrich and Perron, 2006). The effects of plants on river channel planform discussed in Section 3.5.3, although strong, are not uniquely biological. But even if biotic processes ultimately do not exert a strong influence on landscape morphology, they are critical for environmental prediction and are certainly influenced by surface morphology and morphodynamics. Organisms and ecosystems are intrinsically more scale dependent than purely physical systems, so large-scale experiments, especially outdoor facilities, really come into their own once we bring biota into the picture (for example, the Outdoor Stream Lab at St Anthony Falls Laboratory: http://www.safl.umn.edu/facilities/OSL.html) On the whole, we see the coupling of biology and morphodynamics as a major growth area for experimentation.

5.3.3. Submarine landscapes

Submarine experiments so far have focused on processes in and around channels of fixed, imposed geometry. As this is being written early studies of entirely self-formed submarine channels are underway and producing encouraging results. It seems plausible that the next decade will see experiments in which these become routine and even scaled up to submarine channel networks. It is also striking that submarine flows – turbidity currents and debris flows – are still created by direct injection. There is as yet nothing in the submarine experimental world analogous to the natural way in which experimental fluvial landscapes self-organize to create sediment transporting flows from distributed rainfall. Development of the necessary techniques would also lead naturally to experimental linkage of submarine systems to the rest of the continental margin.

5.3.4. Statistical dynamics

One of the chief advantages of landscape experiments is that they allow study of kinematics and evolution that occur over extremely long time scales in nature. In many cases this evolution involves a major stochastic component. Quantifying stochastic processes can be as rigorous as quantifying fully deterministic processes, but the data requirements are much greater because one is estimating probability distributions rather than single variables. We see great potential in experimental landscape research to provide large data sets, including kinematics and surface evolution, to fuel development of new analytical tools for the statistical mechanics of landscapes and stratigraphy.

5.3.5. Coupling geodynamics to surface processes

With the exception of the recent work of Graveleau and Dominguez (2008), the experiments we report here use simplified representations of tectonic motion that are imposed without feedback onto the transport system. This is a reasonable starting point, and the fact that the results are still surprising and complex indicates that there is plenty of work left to be done with arrangements like this. Nonetheless, we know that there is strong coupling between tectonic and surface processes, starting with the effects of isostasy but including less well understood but potentially important feedbacks between, for example, surface transport and folding and faulting. The next generation of experimental facilities should be designed to allow for this, for example by placing the experiment on a dense, viscous substrate that can deform on time and length scales commensurate with the surface evolution. In this regard, the extensively developed field of experimental salt tectonics (Guglielmo et al., 2000; Hudec and Jackson, 2007) provides a good place to turn for inspiration and ideas for how to begin.

5.3.6. Extraterrestrial landscapes

The availability of high-resolution imagery and topography from space probes has opened the door to process geomorphology of other bodies in the solar system. So far most of the work has been directed at Mars but this will certainly expand in the near future. Extraterrestrial landscapes are especially inviting targets for experimental research for many reasons, including lack of direct access and opportunities for simulating the effects of novel surface conditions (e.g. atmospheric density, gravitational acceleration) by experimenting with non-standard fluids and particles. Experimental studies of extraterrestrial systems are in an early stage (e.g. Kraal et al., 2008), but this is a tremendous opportunity – not only to learn about extraterrestrial landscapes, but, by stretching our ideas about what is possible, about Earth's as well.

5.3.7. Model testing

The advent of theoretical models, most implemented numerically, for predicting the evolution of landscapes is part of a larger transformation in landscape research from a qualitative, analog-based framework to a new framework that is quantitative and analytic. In an analytic world, experiments come into their own as a means of testing and refining models under controlled conditions; once tested, theoretical models provide insight as to how experimental observations can be applied at field scales. The power of experiments in this context arises not from their status as miniature analogs, which can never be guaranteed, but from their ability to capture under controlled conditions important elements of the natural complexity we are trying to model. This change in context is in an early stage; experimentation, like theoretical modeling, includes elements of exploration in which its practitioners are simply trying to find out what they can do. Thus both experimenters and theorists have often been content to produce results that resemble field cases. But this phase should be nearing its end, and we expect to see increasing use of experiments to test and develop theoretical models. This is especially true where the emphasis is on self-organization and autogenic dynamics, for which modeling often relies heavily on phenomenological input. In erosional systems, the experiments of Hasbargen and Paola (2000) were directly inspired by numerical predictions of static steady state, and subsequent advances such as that of Pelletier (2004) represent healthy back-and-forth between experiments and theory. Good examples of tight coupling between experiments and theory are provided in the work of the Rennes group cited above comparing experimental landscape response to numerical models. More recent theoretical studies yielding predictions that are amenable to experimental testing include predictions of erosional-landscape response to changes in uplift rate and climate (e.g. Densmore et al. (2007)), and predictions that erosion rates and topography are strongly influenced by erosion thresholds and rainfall variability (e.g. Molnar, 2001; Snyder et al., 2003; Tucker, 2004). On the depositional side, most of the XES experiments discussed in Section 3.2 were designed directly around testing of qualitative or quantitative theoretical predictions. As several of the studies discussed above in Section 3.6 make clear, coupling between theory and experiments is especially close in the realm of turbidity currents and submarine processes, where the complexity of the fluid dynamics and flow interaction with topography present especially significant theoretical challenges. The recently established Community Surface Dynamics Modeling System (CSDMS) (Syvitski et al., 2003) aims to promote community based numerical-modeling efforts, from individual modules to fully linked, comprehensive models, for prediction across all environments in the surface-Earth system. CSDMS should lead to a step increase in the quantity and quality of model predictions available for testing, through experiments as well as in the field. Critical to this is the easily overlooked step of framing the models as far as possible in non-dimensional terms.

One might think that, in the era of increasingly powerful computer models, the need for experiments would be fading. It seems to us that what is happening is precisely the opposite. We know of no time that has seen more construction worldwide of new laboratory facilities for the study of landscapes and sediment dynamics than the present. It is not hard to see why. Particle transport by flowing fluids has resisted first-principles mechanistic modeling since the 19th century, and quantitative methods used to predict sediment flux remain largely empirical. Direct numerical simulation of systems of particles and sediment, even at small scales, requires extravagant levels of computing power. The full scale range (*Gr*) for an experiment 1 m square using fine silt is of the order of 10^5, and as we have seen even such a small experiment has a remarkable capacity to organize itself into complex structures and behaviors. The configurational possibilities for 10^{10} independent particles interacting with flowing fluid represent a decent step towards infinity. And this is without other major pieces in the surface-dynamics puzzle, such as biota.

Even assuming that Moore's law continues to hold, the time needed for an increase in resolution of one order of magnitude for a four dimensional (three space dimensions plus time) computer model is roughly $13.3 * 1.5\ \mathrm{yr} = 20\ \mathrm{yr}$ ((Voller and Porte-Agel, 2002); Clarence L. Lehman, pers. comm.). We do not see computer models replacing landscape experiments anytime soon. The power of experiments to reveal self-organization and system behavior under controlled conditions, to suggest effects to search for in the field, to investigate scale dependence, and to provide rigorous tests of theory, is perfectly matched to the growth of modeling and prediction. We are entering an era in which experimentation will assume fully its natural role alongside theory and field study in the triad of fundamental approaches in stratigraphy and geomorphology.

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