Sedimentology (2000) **47** (Suppl. 1), 121–178

Quantitative models of sedimentary basin filling

CHRIS PAOLA
Department of Geology & Geophysics and St. Anthony Falls Laboratory, University of Minnesota, Minneapolis, MN 55455, USA (E-mail: cpaola@tc.umn.edu)

ABSTRACT

Quantitative modelling of the filling of sedimentary basins was begun in earnest in the 1960s. Dozens of themes and variations have been proposed since then, and have yielded an abundance of idealized stratigraphic patterns as functions of both imposed changes and basin properties. Post-plate-tectonic modelling began with 'rigid-lid' models, which show the stratigraphic signature of subsidence variation. This work introduced the connection between stratigraphy and the rheology of the lithosphere. Rigid-lid models are the simplest type of geometric model, in which the sediment surface is assigned prescribed geometries, usually corresponding to different depositional environments. These can reproduce many aspects of overall stratal geometry but are formally restricted to relatively long timescales, for which quasi-steady surface topography can be assumed. So-called dynamic models attempt to represent the morphodynamics of the sediment surface by abstracting and averaging short-term transport processes. Most of the dynamic models proposed to date can be seen as special cases of a single general morphodynamic equation.

The most important result of the first wave of quantitative basin-filling models is that even relatively simple models can produce reasonable stratal patterns. We now have a wide array of tools for exploring scenarios, searching for general behaviours and effects, and making initial quantitative predictions. We have also learned that basin response to external forcing as recorded in stratigraphy can be as sensitive to the characteristics of the basin as to the forcing.

The main brake on the development of basin modelling is not computing power but lack of methods and data for testing the models we have already developed. Physical experiments, which are only just beginning, are one means of doing this. Experimental stratigraphy is a bridge to quantitative field tests, which will require collaboration among academic researchers from a wide range of areas, and between academia and industry, on projects of greater scale and degree of integration than we have seen so far. The advancement of quantitative sedimentary geology will also require significant changes in the way the subject is taught, at all levels.

Keywords Basin, sedimentation, stratigraphy, quantitative model.

'The sediments are a sort of epic poem of the Earth.' – Rachel Carson

INTRODUCTION

Goals and scope of the paper

The 'epic poem of the Earth' is written in a language that we still do not entirely understand. So far, interpretation of the language of sediments has been done with scant recourse to mathematics, the common language of science. However, this is changing. As our science matures, we find that application of quantitative techniques lets us ask sharper questions, makes our hypotheses less ambiguous and hence easier to test, rules out some apparently plausible explanations and suggests new ones that would never have occurred to our unaided intuition. In the last 20 years or so this analytical approach has led to a bloom of quantitative models of the filling of sedimentary basins.

Much has been written about the supposed goals and motivations for modelling in the Earth sciences, but in the end the goal of modelling is the same as that of all other aspects of science: to generate insight. Models are simply expressions of our ideas about how things work. All of us use models, no matter how inductive or empirical we think we are. Models of one form or another have been used since geology began, to guide our observations and to help us make sense of what we see and measure. What has changed is that to our collection of descriptive models we have added more analytical, quantitative models based on physical, chemical and biological principles. This paper is a survey of what has been done in theoretical and physical modelling of sedimentary systems at large scales, what we have learned from it and what some of the main challenges are now.

It is worth clarifying some terms before going any further. I will follow common usage if not etymology and use 'stratigraphy' for the study of sedimentary sequences at large scales (roughly formations and greater), and 'sedimentology' for the study of sediments and sedimentary rocks at smaller scales. 'Sedimentary geology' includes both. 'Physical modelling' means experimental studies of sedimentary processes, at either full or reduced scale. 'Theoretical modelling' means any form of modelling based on quantitative analysis. Theoretical models become 'numerical models' when they involve equations that must be solved numerically. In some cases numerical models may use equations that are supposed to embody real processes but are not directly derived from differential equations; this subcategory of numerical models is often called 'rule-based'. In the extreme case, rule-based models may be semiquantitative or even qualitative even though they are implemented numerically.

This paper is a review of quantitative stratigraphic models. For the most part, this means that its scope is restricted to models of sedimentation on scales for which subsidence is important. However, the focus is on basin filling, not subsidence itself. And, although deep-marine sedimentation is of great importance in modelling sedimentary systems, aspects of this are covered in the accompanying review by Kneller & Buckee (2000) and so will not be discussed here.

The development of quantitative basin analysis

Space does not permit a lengthy historical review of the evolution of the quantitative approach to understanding sedimentary systems. However, as this paper was being written the sedimentary–geology community learned with sorrow of the death of Francis Pettijohn. Pettijohn was one of the founders of modern basin analysis, and his work has greatly affected the research reviewed here. His approach was most comprehensively expressed in his well-known books (Krumbein & Pettijohn, 1938; Pettijohn, 1957; Pettijohn & Potter, 1964; Pettijohn *et al.*, 1973; Potter & Pettijohn, 1977), of which the latter two are the most relevant to basin analysis.

Francis Pettijohn was an outspoken advocate of a pragmatic, field-orientated approach to basin study. He doubtless would not have approved of many of the more esoteric aspects of the work described in this paper. Nonetheless, his approach was also consistently analytical. The Pettijohn school placed great importance on reconstructing spatial transport patterns via palaeocurrents, a topic that is currently not in vogue but that will return to play a major role in modelling as we move from two to three dimensions. An early example of quantitative analysis developed by Pettijohn and some of his students involved the exponential model proposed by Sternberg (1875) for downstream fining in fluvial systems. The idea was that by fitting this model to grain-size data from ancient systems, one could extrapolate upstream to estimate how far away the source area was (Schlee, 1957; Pelletier, 1958; Yeakel, 1962; Meckel, 1967). The estimate has considerable uncertainty but is nonetheless better than having no estimate at all, both still typical features of applications of quantitative models to ancient systems. The Sternberg fining model was one of the first quantitative models developed in fluvial geomorphology. Although essentially empirical, it was supported by a simple physical theory that added to its appeal. Its application by Pettijohn's group nicely illustrates the close connection between geomorphic and stratigraphic modelling. The basic logical structure – combining a quantitative model with careful measurements from rocks and general geological constraints to derive new information about the depositional system – is also still recognizable in the application of models to the stratigraphic record today.

The first modern stratigraphic model was proposed by Sloss (1962). It was conceptual, rather than quantitative, but it was formulated in a careful and disciplined way that made it as well suited as a basis for quantitative modelling as for conceptual refinement. Sloss argued that the shape of a body of sedimentary rock is controlled by four variables:

Q: the quantity of sediment supplied per unit time, equivalent in current models to the supply sediment flux $Q_s(\mathbf{x}_0, \mathbf{x}_1, \ldots)$ where $\mathbf{x}_0, \mathbf{x}_1, \ldots$, are the locations of sediment sources at the edges of the basin.

R: the 'receptor value', the rate of subsidence relative to base level. This is now called 'accommodation' and is equivalent in current models to $\sigma + \partial H/\partial t$, where σ is subsidence rate, H base level and t time.

D: the dispersal, a measure of the total effect of all the sediment transport processes, expressed as a rate. This is represented in various forms in the work discussed below but is closest to the sediment flux distribution within the depositional system, $q_s(x, y)$.

M: a measure of the composition and texture of the sediment supply. This could be represented in various ways depending on the detail required, but fundamentally it is the probability density $f_{D0}(D)$ of grain size D in the sediment supply.

On the basis of this model, Sloss worked out a series of scenarios for transgression and regression (Fig. 1). His generalized formulation implied a broad view of the causes of sedimentary signals. This view was to some extent lost in the early stages of sequence stratigraphy, with its exclusive emphasis on sea-level (a restriction to which Sloss himself later objected; Sloss, 1991). Modern basin-filling models, however, very much reflect this spirit of searching for a complete, fundamental set of independent controlling variables for stratal geometry.

The conceptual branch that grew from Sloss's ideas blossomed in the 1970s in the form of sequence stratigraphy. The quantitative branch took longer to bear fruit; the results are the subject of this paper. An early attempt to formalize and quantify Sloss's conceptual framework was presented by Schwarzacher (1966), who proposed a one-dimensional sedimentation model (a first-order, linear, ordinary differential equation) based on the hypothesis that the sedimentation rate under water increases with depth. This is how some aspects of carbonate sedimentation are modelled today. The most important and prescient result of Schwarzacher's analysis was that his model evolved toward a steady-state solution in which sedimentation rate balances accommodation (subsidence plus eustatic sea-level rise).

The first modern-style dynamic model of basin filling is, interestingly enough, a model of evaporite sedimentation proposed by Briggs & Pollack (1967). All the elements are there: analysis from first principles to derive a model; a target field area (the Michigan Basin) to provide appropriate palaeogeographical constraints; numerical solution of the equations; and (naturally) state-of-the-art computer graphics (Fig. 2a). It is interesting that after this early work, evaporite sedimentation has all but disappeared from quantitative basin-filling models (though for an exception, see Waltham (1992)).

Both Sloss's paper and the Briggs–Pollack evaporite model figure as inspirations for the now classic book by Harbaugh & Bonham-Carter (1970). In a single stroke, this book showed the way to build computational stratigraphic models. In addition to general solution methods and algorithms, it presents a remarkable range of models (Fig. 2b): diffusion in one and two dimensions, potential flow, delta evolution, evaporite precipitation and carbonate deposition (not to mention palaeoecology and hydrothermal ore formation). Stochastic effects are included, with an emphasis on Markov chains. A number of the sedimentation algorithms had been developed earlier by Harbaugh and were published beginning in 1966 (Harbaugh, 1966) by the Kansas Geological Survey in its Computer Contribution series, one of the most farsighted publication series ever produced by a regional geological survey. It would be difficult to exaggerate the contribution that Harbaugh and Bonham-Carter's work made to the development of quantitative stratigraphy. If it can be criticized, it would only be for being slightly ahead of its time: the missing ingredient was plate tectonics.

The development of plate tectonics obviously had an enormous influence on sedimentary geology as on most aspects of Earth science – it is no accident that one of the crucial papers that began the modern tectonic era (Dietz, 1963) was a reinterpretation of the classical geosyncline. Thus the connection between plate tectonics and stratigraphy was established immediately. The advent of plate-tectonic theory also had a more subtle, long-term effect on thinking in the Earth sciences: it led to a great flowering of geomechanics. The realization that

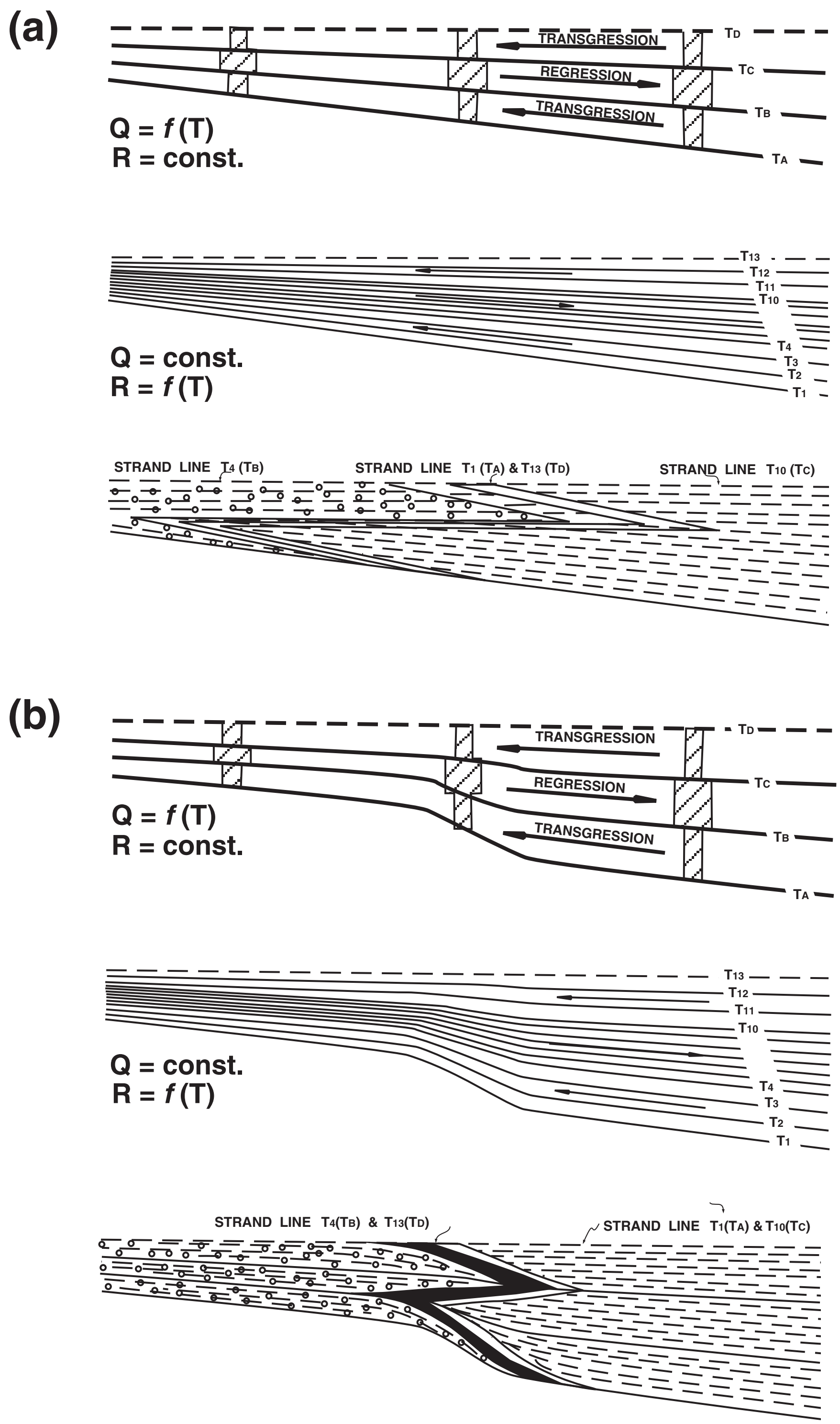


Fig. 1. Stratigraphic cycles as modelled by Sloss (1962). In each scenario, the transgression–regression pattern (bottom) can be produced either by changing sediment supply (top) or relative sea-level (middle). The cross-hatched areas in the top panels represent sediment supply, and the lines in the middle panels represent equal time intervals. (a) Linear increase in subsidence; (b) nonlinear increase in subsidence.

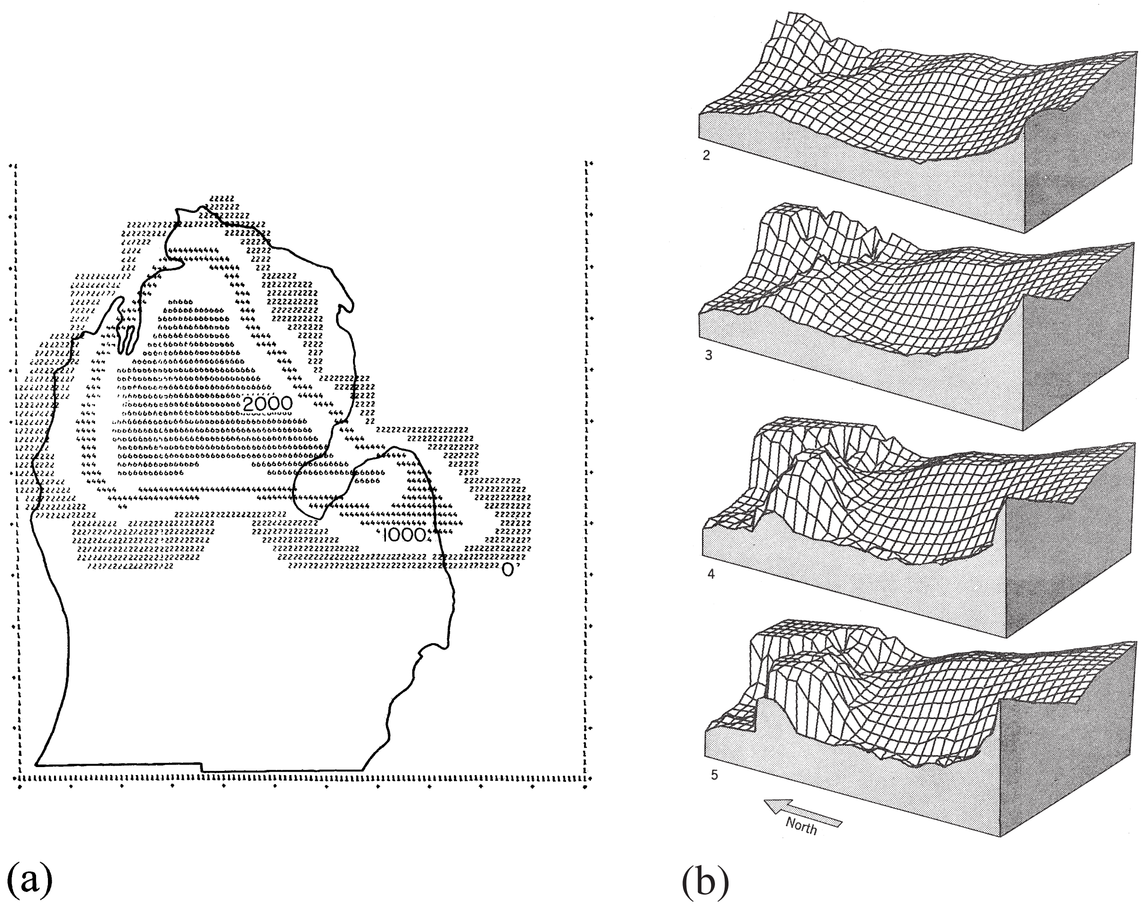


Fig. 2. The state of modelling of basin sedimentation in the late 1960s. (a) Evaporite thickness in the Michigan basin, reprinted with permission from Briggs & Pollack (1967; © 1967, American Association for the Advancement of Science, contour interval is in feet). (b) Three-dimensional image of carbonate bank migration from Harbaugh & Bonham-Carter (1970; © 1970, reprinted by permission of John Wiley & Sons, Inc.). The block is 64 km long.

so many apparently unrelated geological phenomena could be unified in the light of a bold, simple mechanical model inspired new efforts to understand other aspects of geodynamics in terms of simple mechanical analogues. In sedimentary geology, the most dramatic effect of this growth of 'mechanism' was the advent of quantitative models for basin subsidence. These, in turn, allowed development of quantitative models for the geometry of the strata filling the basin.

The crucial paper that launched the revolution in stratigraphic modelling was published in 1978. As it happens 1978 saw the publication of several papers that together make it a turning point in the analysis of sedimentary systems. Here we depart briefly from following scientific themes to focus on that year.

1978 – the true beginning of the third millennium

It is a commonplace that decades as defined by a particular zeitgeist often do not begin and end on exact calendar boundaries. Compared with a few years' leeway for decades, it does not seem unreasonable to define the start of a new millennium for sedimentary geology 22 years early, in 1978. Consider the following developments.

Physically based subsidence models

These were developed in the geophysics community. In a very brief paper, McKenzie (1978) provided the first quantitative model for thermal subsidence following extension. This model followed from Sleep's (1971) observation that the time history of basement elevation in a passive margin is equivalent to the spatial distribution of crustal elevation about an ocean spreading centre. McKenzie then proposed that passive-margin subsidence can be modelled in terms of an instantaneous extension followed by conductive cooling and shrinking of the stretched lithospheric column. A series of papers elaborating on this model soon followed, but McKenzie's basic idea remains at the centre of analysis of extensional subsidence. Flexural subsidence was first modelled quantitatively by Haxby *et al.* (1976). Beaumont (1978) built on this by proposing a physical model for basin subsidence on viscoelastic lithosphere, i.e. lithosphere whose rheology is elastic-dominated on short timescales and viscous-dominated on long ones. This work introduced the idea that stratigraphic patterns may bear the signature of the rheology of the underlying lithosphere. A few years later, Jordan (1981) and Beaumont (1981) proposed elastic and viscoelastic (respectively) first-order models for foreland basins. Later we will look at how all of these subsidence models were coupled to surface-evolution models to build stratigraphy.

Shoreline migration

It had been assumed implicitly since the problem was first considered that palaeoshoreline as observed in the rock record is a direct recorder of relative sea-level variation. Pitman (1978) provided the first formal analysis of the relation of shoreline to sea-level in the light of tectonic subsidence, and came up with the surprising result that shoreline need not be directly coupled to relative sea-level. Pitman's result is shown in Fig. 3. The key assumptions are: steady rigid-beam subsidence; constant coastal-plain slope (in space and time); sea-level variation around a steady long-term fall (the target interval was mid-Cretaceous through Miocene time); and, most important, zero sedimentation rate at the shoreline. With these assumptions, Pitman was able to derive a first-order linear differential equation for the lateral motion of the shoreline. In the long-term limit, the model approaches a steady state in which the shoreline is at the point on the coastal plain where the subsidence rate is equal to the rate of eustatic fall. The timescale over which this happens is a 'tectonic time constant' that depends on basin length, subsidence rate and the slope of the coastal plain. For eustatic variations with periods that are long compared with this time constant, the shoreline responds in a quasi-equilibrium manner. Shoreline and sea-level are out of phase by 90°, with maximum transgression corresponding to maximum rate of sea-level rise, and vice versa. For typical basin geometries, the time constant is of the order of 10^4–10^6 years, well within the range of known eustatic forcing mechanisms.

Angevine (1989) extended Pitman's analysis to both short-term and long-term cycles (relative to the same tectonic time constant) and showed that the model implies that shoreline response to long-term sea-level change is not only phase shifted but also strongly attenuated (Fig. 3). Angevine also showed that in the high-frequency limit, the shoreline behaves in the expected manner, varying in-phase with sea-level with an amplitude set by the amplitude of sea-level variation and the slope of the coastal plain. Thus sedimentary basins could be thought of as high-pass filters for sea-level changes.

Pitman's counter-intuitive result sparked considerable research, including both further theoretical analyses and attempts to verify the phase-shift behaviour from field data. Beyond its obvious importance to interpreting sea-level

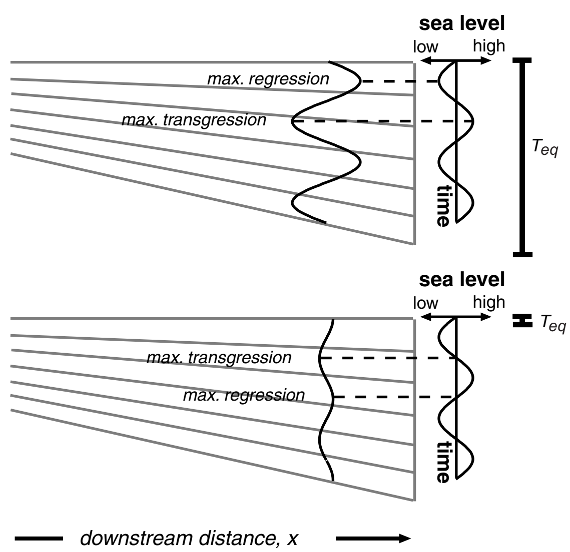


Fig. 3. Sketch of high- and low-frequency shoreline behaviour as predicted by Pitman (1978) and Angevine (1989). T_{eq} is the basin time constant. Transport is from left to right, and the position of shoreline is superimposed on equally spaced time lines for the high-frequency (top panel) and low-frequency (bottom panel) cases. Note the change in both phase and amplitude of the shoreline response for the two cases.

history from the stratigraphic record, Pitman's model was a breakthrough in bringing a measure of mathematical sophistication to modelling a stratigraphic signal (the marine–nonmarine facies boundary). It involved an oversimplification of the mechanics of coastal-plain sedimentation, but its real importance lies in having opened up a new set of questions that previously had not even been considered.

Alluvial architecture

Building on earlier qualitative work (Allen, 1965; Allen, 1974), Leeder (1978) proposed the first physically based quantitative model for alluvial architecture. Leeder envisaged a fluvial channel belt of fixed cross-sectional area avulsing at a constant mean frequency to random locations within a fluvial basin. Each avulsion removes any previously deposited floodplain fines, which fill all space not occupied by channel-belt deposits. The result most frequently cited from Leeder's initial work is that, all else being held constant, periods of rapid subsidence produce cross-sectional panels with lower density and connectivity of channel-belt deposits than periods of slow subsidence.

Leeder's elegant and intuitively appealing approach provoked numerous refinements and variations, as well as field tests, which are discussed below. One of the key assumptions, that the avulsion rate is independent of sedimentation rate, was later shown experimentally to be incorrect (Bryant *et al.*, 1995), and current architecture models allow avulsions to be determined by sedimentation or imposed externally (Mackey & Bridge, 1995). None of this diminishes the fundamental contribution of Leeder's model in establishing the quantitative connection between avulsion dynamics and the mesoscale architecture of fluvial deposits.

River-channel width

The problem of what sets the width of rivers has long been one of the most fundamental questions in fluvial geomorphology. In engineering studies of modern river systems, the width is usually specified, but in modelling ancient systems over long timescales, it must be determined from basic principles. Parker (1978a,b) proposed the first width models for channels with active sediment transport. For sand-bed rivers (Parker, 1978a), the width is set by a condition in which bank erosion is balanced by a lateral turbulent flux of suspended sediment away from the channel centre. For bedload-dominated (gravel) channels (Parker, 1978b), there is no bank erosion. Lateral drag slows the near-bank flow, allowing for weak bedload transport in the channel centre. These papers provided an independent, physically based means of coupling channel width to other variables (e.g. sediment flux, slope), thus providing a closed system of equations for dynamic river evolution.

It is worth noting that this group of landmark papers includes contributions from geophysics, civil engineering and stratigraphy. The growth of analytical sedimentary geology has, above all, been sparked by a confluence of ideas from different fields.

The growth of quantitative, physically based stratigraphic models

The publication of the initial subsidence models provoked a flurry of work examining the implications of various subsidence scenarios for stratal geometry. Again, much of the initial work was done in the geophysics community. The novelty and enormous potential of this approach meant that work on large-scale stratigraphic modelling soon overshadowed analytical study of smaller-scale aspects of sedimentary systems. The initial emphasis was on rheology and lithosphere dynamics, and on predicting large-scale stratal relationships that could be identified in seismic sections. In this context, the sediment fill was generally represented by a horizontal surface; variable stratal geometry was produced by the transport and rotation of palaeosurfaces by the deforming basement. Following fluid-mechanics usage, we will term models with a fixed horizontal sediment surface 'rigid-lid' approximations.

An improvement on the stratigraphic side was made when these geodynamic models were coupled to more realistic representations of the sediment surface. The next step beyond a horizontal surface is to assign prespecified geometries (usually planes with fixed slopes) to surfaces associated with various environments (e.g. fluvial, or submarine). Models that do this are called 'geometric models'. The most recent step in the evolution of basin models has been to represent the dynamics of the sediment surface explicitly. Various approaches have been taken to this, which we will review in the next section.

This line of work has led us to the present state of large-scale modelling of sedimentary systems. Major collections that give a sense of the state of basin modelling are Cross (1990), Franseen *et al.* (1991), Slingerland *et al.* (1994) and Harbaugh *et al.* (1999). The second and fourth of these were produced under the auspices of the Kansas Geological Survey, continuing the role it played in launching quantitative stratigraphic modelling via its Computer Contribution series. An excellent synthesis of modern basin analysis, including extensive discussion of quantitative models, is presented in Allen & Allen (1990). The end of the twentieth century must be one of the most exciting times in history to be in sedimentary geology: joining the ever-expanding armoury of field techniques and conceptual models we now have a wide array of quantitative models. In the next sections, a review is presented of some of the main classes of these, and some of their principal results.

A FIELD GUIDE TO THEORETICAL MODELS

Because basin modelling is in a period of vigorous growth, people are still experimenting with a wide variety of approaches. Models in current use can be classified along the following nonexclusive axes.

Geometric vs. dynamic models

A *geometric model* is one in which the sediment surface is represented by one or more surfaces with predetermined geometry. The simplest kind of geometric model is a rigid-lid model, in which the surface is a fixed horizontal plane. In this case all of the variation in stratal geometry is associated with subsidence. Additional realism is gained by assigning slopes to the sediment surface, which may vary to reflect, for instance, fluvial vs. marine-shelf environments. A geometric model can be adjusted to conserve sediment mass and to allow for variation in sediment supply, a major improvement over rigid-lid models. Geometric models can produce many major features of stratal geometry, including unconformities and plausible time–space sections, and they are computationally simple. They offer a high degree of flexibility but incorporate few process-based constraints on what real transport systems can and cannot do. They also miss stratal features in which the dynamics of the sediment surface plays a role, for example unconformities that develop when fluvial slopes change in response to a change in water supply (Fig. 4). This shortcoming is remedied in *dynamic models*, in which the surface evolves according to some representa-

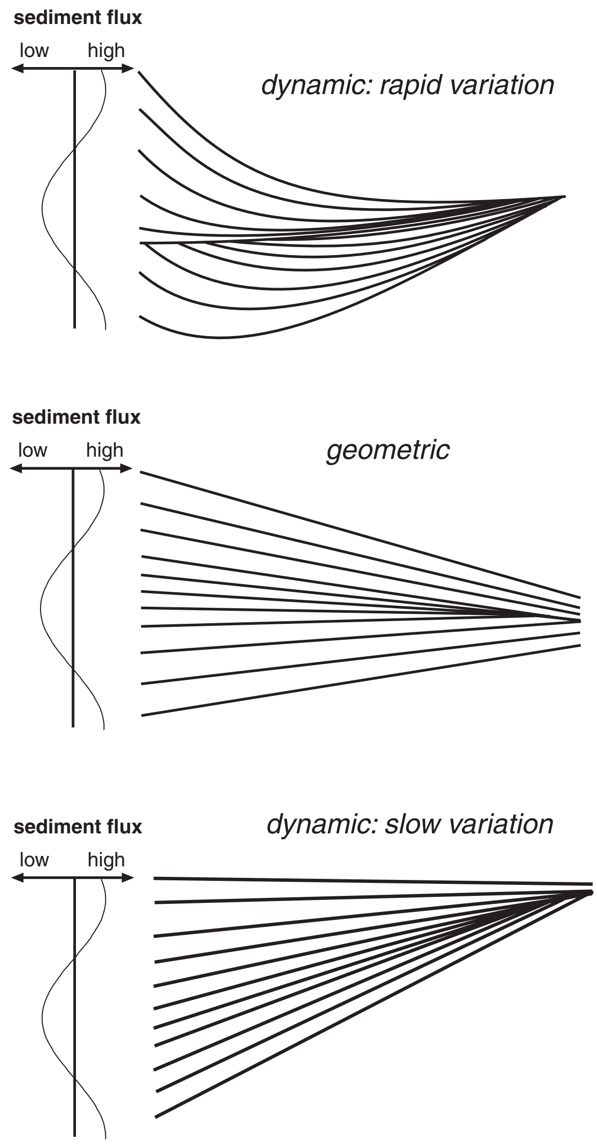


Fig. 4. Sketch showing how geometric and dynamic fluvial models respond differently to a change in sediment supply. In all cases, subsidence is a rigid beam with hinge to the right. A single cycle of variation in sediment supply is imposed. The only role that timescale plays in the geometric case is to determine the total amount of subsidence (rotation) during the cycle. In the dynamic case, rapid changes strongly affect the sediment surface. A proximal unconformity is produced by changes in the initial surface slope. However, slow changes produce a quasi-equilibrium response in which the sediment surface plays a minor role. Hence the stratigraphy produced by the geometric model more closely resembles that for the slow-variation dynamic case.

tion of how sediment and transport and deposition interact with subsidence to produce surface topography. Most 'dynamic' models of sediment surfaces are technically kinematic (i.e. involve only length and time) rather than dynamic (involving length, time and mass). However, if they have been derived carefully, force balances play a role in their origin if not their final form. Dynamic basin models are also sometimes called 'process-based' models.

One-dimensional vs. two-dimensional vs. three-dimensional models

The first dimension in stratigraphy is the vertical, and one-dimensional models are a natural starting point for thinking about vertical sequences. Two-dimensional models add the horizontal dimension parallel to the mean transport direction. The development of well-founded one- and two-dimensional models is an essential first step in basin modelling, even though many natural systems are three-dimensional. Two-dimensional models that attempt to account partially for three-dimensional effects (e.g. out-of-plane mass transfer) are sometimes called 'two-and-a-half' dimensional. Beyond the fact that lower-dimensional models are much easier to constrain and test, systems with significant three-dimensional variation can sometimes be understood as a succession of two-dimensional slices, if lateral mass fluxes are small relative to fluxes in the direction of the slices. Nonetheless, there are many cases, such as basins with multiple interacting source points, or combined lateral and transverse flow, or strongly three-dimensional subsidence, in which a fully three-dimensional model becomes necessary.

Rule-based vs. deductive models

Traditionally, quantitative models are developed by starting with fundamental governing equations, defining boundary conditions, making suitable simplifications and then solving the equations. The skill of the modeller is mostly in applying the right boundary conditions, making insightful approximations and finding the best solution method. As systems become complex, it becomes increasingly difficult to break them down into elementary objects and continua to which Newton's laws can be applied exactly. Typically the first recourse when things get too complicated is to substitute semi-empirical relations for fundamental equations. Trivial and, by now, standard examples in sedimentary-systems modelling include using drag coefficients and empirical sediment-flux laws, the latter in effect substituting for a momentum equation for sediment. A more extreme way of dealing with dynamical complexity is to construct a model from 'rules' that may range from attempts to get the gist of fundamental laws in simplified form, through attempts to abstract dynamics that are below the resolution of the model, to purely intuitive guesses. Models based on this approach are called *rule-based*. We will use *deductive* for models using the more traditional approach.

The rule-based approach to modelling is well developed in ecology, where a deductive approach starting from first principles is generally not practical (Starfield *et al.*, 1989, 1993; Starfield, 1990). In the physical sciences, the distinction between rule-based and deductive models is related to a more fundamental debate now going on between the so-called 'reductionist' viewpoint and an alternative that one could term the 'synthesist' viewpoint. The difference becomes important in the context of complex systems. Reductionism refers to taking the system apart into its fundamental components and constructing a model from the bottom up. The synthesist viewpoint (Tipper, 1992; Goldenfeld & Kadanoff, 1999; Werner, 1999) is that the dynamics of complex systems may occur on many levels (time or space scales), and that the dynamics at higher levels may be more or less independent of that at lower levels. Werner (1999) has emphasized the likelihood that higher-level dynamics is controlled by only a few crucial processes at the level below, with the implication that the modeller must focus on only those crucial processes. In this viewpoint, attempting to extract the dynamics at high levels from comprehensive modelling of everything going on at lower levels is at best clumsy and inefficient, and at worst entirely misleading – like analysing the creation of *La Bohéme* as a neurochemistry problem.

The synthesist viewpoint can be used as a basis for a rule-based approach to modelling complex sedimentary systems. The 'rules' are then the modeller's synthesis of the important lower-level dynamics. Models of sedimentary systems that are synthesist in spirit include Chase (1992), Werner & Hallet (1993), Werner & Fink (1993), Murray & Paola (1994) and Werner & Kocurek (1997). In current practice, however, rule-based models are also used simply because they are convenient, or for biological processes that cannot readily be modelled from first principles (e.g. carbonate precipitation). We will return to this subject later.

Coupled vs. uncoupled models

In general, *coupling* refers to connecting different components of a model system so that they can interact, via shared terms in governing equations, special 'coupling conditions' and/or shared boundaries ('internal boundaries'). The key issue in multicomponent models is the strength and direction of the feedbacks allowed. Modellers speak of *decoupling* components if they are either independent of one another or one is driven passively by the others. For instance, a fully coupled tectonics–sedimentation model accounts both for tectonic controls on sedimentation and for the effects of sediment transfer on tectonics (e.g. loading-induced subsidence, uplift forced by erosion). A decoupled (and very naïve) model would be one in which sedimentation was passively driven by tectonics, or they had nothing to do with one another. Since our focus here is on models of sedimentation, coupling will usually refer to joining single-environment models together to obtain a model of a whole sedimentary system. For instance, a complete model of the continental-margin system would require coupling models for the fluvial (erosional and depositional), continental shelf, continental slope and deep-marine environments. There is obviously some trade-off between the number of coupled components and the level of detail in each component. It is also easy to overlook the fact that the dynamics of a coupled model can be as sensitive to the conditions and solution techniques applied at the junctions between the component parts ('internal boundaries') as to the behaviour of the parts themselves (Swenson *et al.*, in press). Even a model composed of linear parts becomes nonlinear if the coupling between any of the parts is nonlinear. This can lead to surprisingly complex behaviour even from relatively simple models.

Analytical vs. simulation models

An *analytical model* means one designed primarily for analysing general aspects or behaviours of a system. It is typically intended to be an abstract, minimalist representation, retaining only those aspects of the dynamics that the modeller thinks are essential to the problem at hand. Analytical models are typically not intended to be used to reproduce particular cases. Rather they are aimed at illustrating fundamental principles or revealing general behaviours. Analytical models involve a minimal number of free parameters and are thus relatively testable, but they often involve high levels of simplification. *Simulation models* are intended to reproduce more of the details of specific cases. By design they are more 'realistic' than analytical models. Simulation models are generally much larger computationally than analytical models, and their flexibility usually comes at the cost of including more parameters that have to be set by the user, and so can be adjusted to fit particular cases. This makes simulation models harder to test. Their large size and number of parameters also makes it awkward to explore their parameter space comprehensively. The distinction between analysis as a search for general effects and simulation as an attempt to reproduce particular cases can as well be applied to the use of models as to models themselves. In this sense, simulation can be viewed as the modelling expression of what Sloss (1962) called the 'Twenhofelian' viewpoint, in which the emphasis is on understanding the unique character of each different basin. In contrast, analytical modelling would represent the search for 'order and system' that Sloss advocated as the key to moving stratigraphy out of its descriptive (adolescent) phase. A broader and more accurate view is that basins have both unique aspects that give them their 'personality', and general features and processes that they share more or less widely. A modeller might see a basin's 'personality' as the net effect of all the parameters and boundary and initial conditions that must be specified in a general process model in order to reproduce a particular case.

Probabilistic vs. deterministic models

A *probabilistic model* is one that includes random variables, i.e. variables that are represented by probability–density functions. A *deterministic model* does not. In most probabilistic models, the random variables are handled by using random-number generators, which means the model must be computational. (It is also possible to model probability–density functions via the calculus of random variables, but this has not been done in sedimentary-systems modelling.) Nearly all current models for basin filling are deterministic, at least at the largest scales. Random processes have been used extensively at the bed scale (which we will not cover here) and, in fluvial systems, at the mesoscale, as we will discuss below. Deterministic models can produce apparently stochastic behaviour if the system is sufficiently complex, a phenomenon often referred to as 'chaos'. Chaos is implicitly present in stochastic models, since random-number generators produce their output using deterministic computations.

THEORETICAL MODELS OF BASIN FILLING: A GUIDED TOUR

Geometric and carbonate models

In geometric models, the sediment surface has a prescribed geometry. The simplest geometry is a horizontal plane at sea-level; we term these 'rigid-lid' models. Since surface dynamics has been removed from the problem, rigid-lid models capture only the effects of subsidence in governing stratal geometry. The basic equations governing subsidence are by now a standard part of geodynamics (Turcotte & Schubert, 1982; Allen & Allen, 1990; Angevine *et al.*, 1990; Slingerland *et al.*, 1994). There is an excellent summary of the stratigraphic results of the initial rigid-lid models in Allen & Allen (1990); we will mention only a few highlights here. The McKenzie (1978) thermal subsidence model was one-dimensional, so the only stratigraphic variable was sediment thickness vs. time. Early modifications to the McKenzie model focused on including the related effects of differential extension (Royden & Keen, 1980), lateral heat flux (Steckler, 1985; Sawyer *et al.*, 1987) and finite rifting time (Jarvis & McKenzie, 1980; Cochran, 1983), and were aimed mainly at improving estimation of the subsidence history, particularly the partitioning of subsidence between the rift and thermal stages, and at modelling uplift adjacent to the rift. All of these early thermal models used a local isostatic balance. Local isostasy throughout deposition of the sediment column allowed for the development of relatively straightforward computational procedures for accounting for the role of sediment loading and compaction in amplifying subsidence. Removal of these effects along with water-depth variation allows one to estimate tectonic subsidence relative to eustatic sea-level through time, a procedure known as 'backstripping' (Angevine *et al.*, 1990; Gildner, 1990). More recent backstripping techniques also allow for flexural effects (Steckler *et al.*, 1993). Backstripping in two dimensions allows one to reconstruct approximate palaeotopography from present-day depositional geometry (Steckler *et al.*, 1999).

Two-dimensional rigid-lid models

More interesting stratal relations can be investigated by using two-dimensional subsidence models, particularly when flexural effects are included. One of the most important stratigraphic results of including flexural effects centres on the temporal evolution of flexural rigidity. The issue is important to stratigraphic modelling because the effects of variable rigidity could easily be misinterpreted as being caused by sea-level changes. Watts *et al.* (1982) pointed out that oceanic lithosphere tends to stiffen (increase in flexural rigidity) as it cools. For passive margins, this increase in rigidity causes the flexure at the margin to widen with time, producing stratigraphic onlap (Fig. 5a; Watts *et al.*, 1982). The opposite effect – stratal offlapping of the basin margin – can be produced by viscoelastic behaviour of the lithosphere (Fig. 5b; Beaumont, 1978). Viscoelasticity is a manifestation of the

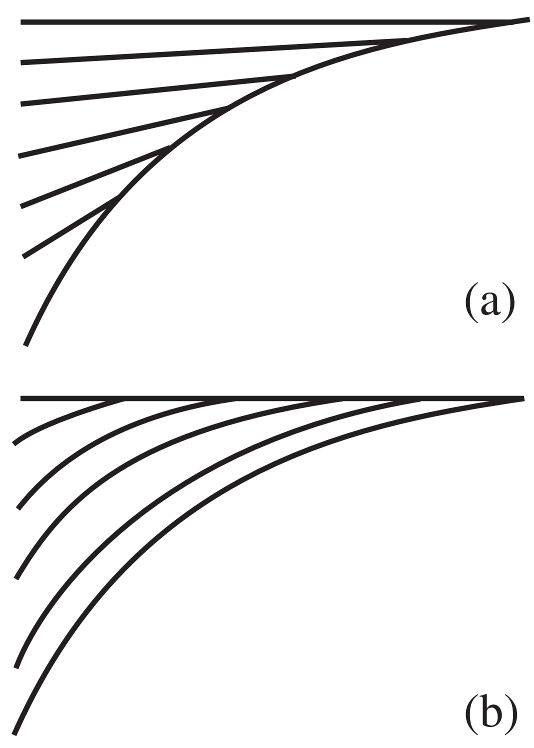


Fig. 5. Sketch showing the stratigraphic effects of two scenarios for time change in flexural rigidity at a basin margin. (a) Onlap associated with increasing rigidity due to cooling; (b) offlap associated with decreasing rigidity due to long-term viscoelastic relaxation.

familiar idea that rocks are elastic on short timescales and viscous on long timescales. Elasticity on seismic timescales and flow of the asthenosphere on timescales of the order of 10^4 years are well known aspects of viscoelasticity; what is in dispute is the upper limit of timescale on which viscous relaxation can be important. Karner *et al.* (1983), McNutt *et al.* (1988) and Kruse & Royden (1994) have presented evidence for long-term maintenance of flexural rigidity in basins and mountain ranges. Kusznir & Karner (1985) studied the behaviour of a multilayer model of the lithosphere based on measured material properties and concluded that most viscous relaxation takes place within 200 000 years of loading. Relaxation is slowest for thermally old (thick) lithosphere, but for that case also produces the smallest relative change in rigidity. There is now direct evidence for this multilayer model in the form of two scales of folding beneath Tibet (Jin *et al.*, 1994). On the whole there seems to be little evidence for significant viscoelastic behaviour in sedimentary basins on timescales of longer than a few hundred thousand years.

From a stratigraphic viewpoint, rigid-lid modelling is most appropriate for looking at stratal patterns generated by differential subsidence and rotation. It is still used for this (Hardy & Poblet, 1994; Mountney & Westbrook, 1997; ter Voorde *et al.*, 1997). Nonetheless, to anyone who has spent a career thinking about sedimentary processes, rigid-lid models seem extremely crude. Their utility in stratigraphy reflects the fact that the vertical scales associated with subsidence are many times greater than those of variation in the sediment surface generated by depositional processes. Apart from the continental slope and restricted subenvironments like alluvial fans and the edges of carbonate platforms, sedimentary slopes typically do not exceed a few parts per thousand. This means that over a basin length of the order of a hundred kilometres, depositional slopes produce elevation changes of the order of a few hundred metres. This is comparable with the range of eustatic sea-level variation, to the range of vertical flexing associated with intraplate stress variation (Cloetingh *et al.*, 1985; Cloetingh, 1986), and to the thickness of many individual formations. For models that attempt to resolve these features, it is inconsistent not to include surface variation, at least geometrically. Nonetheless, scales of a few hundred metres are small compared with the total subsidence of major sedimentary basins. The success of rigid-lid models highlights the dominant role of subsidence in governing stratigraphic patterns over long timescales. The rigid-lid approach also introduced the idea that these patterns might be used to measure the rheology of the lithosphere.

The next step is to model the sedimentary surface by imposing fixed geometric constraints on it, such as slopes or deposit lengths. This approach produces what are called 'geometric models'. The slope typically depends on the environment, with submarine slopes steeper than subaerial ones. A classical geometric model is the Pitman (1978) analysis of shoreline response to eustatic sea-level variation, which was discussed above. In this case, both subaerial and submarine parts of the coastal plain were assigned a single slope, and a sedimentation rate of zero was assumed for the shoreline. This forced the fluvial section to be always erosional. These simplifications allowed Pitman to derive a first-order linear differential equation for the movement of the shoreline. As discussed earlier, Pitman focused on the behaviour of this equation for slow eustatic cycles, i.e. those whose period is long compared with a tectonic time constant that appears as part of the solution to the governing equation.

Pitman's (1978) theory is also a good example of an analytical basin model. It was not intended to provide a detailed simulation of a particular stratigraphic section, but rather used judicious approximation to find a simplified equation whose behaviour could be readily analysed and used to obtain general results. The main basic prediction, that shoreline should track rate of sea-level rise rather than sea-level over long timescales, is straightforward and testable.

A wide range of geometric models have been developed subsequently. Two broad themes stand out. One is that geometric models are the main arena in which carbonate stratigraphy has been coupled with clastic sedimentation. The other is that geometric models are the ones most closely tied to sequence stratigraphy (Van Wagoner *et al.*, 1990; Weimer & Posamentier, 1994; Emery & Myers, 1996). This is not surprising: the dissection of stratigraphic sections into unconformity-bounded sequences is essentially a geometric process, and the conceptual models behind sequence stratigraphy are geometric, following on from Sloss (1962). In many cases, only quantitative implementation separates conceptual models from geometric computer models. The connection with sequence stratigraphy has led to the widespread use of geometric models to look at the effects of eustatic variation on the stratigraphic record. Ross (1991) gives a nice summary of the linked development of sequence stratigraphy and quantitative models.

Carbonate shelf models

Carbonate production is modelled using a rule-based approach, and at present it is difficult to imagine how a process so dominated by biological effects could be tackled any other way. The common feature of current models is that carbonate production rate depends primarily on depth, and very strongly (Fig. 6). The ways in which this is implemented illustrate the range of styles in rule-based modelling. The most flexible but least analytical is to specify a general form for the accumulation curve but to let the user set all the internal break points in the curve (Demicco & Spencer, 1989; Demicco, 1998). Since light attenuation is the basic cause of the depth dependence of carbonate production rate, more analytical forms can be obtained by relating carbonate sedimentation to known functions for light attenuation in the ocean (Bice, 1991; Bosscher & Schlager 1992, 1993; Bosscher & Southam, 1992; Galewsky, 1998) as well as other factors like turbidity and temperature (Lerche *et al.*, 1987) and rate of sea-level rise (Watney *et al.*, 1991). The user must still supply parameters, but their physical meaning is clearer and their values less arbitrary.

The depth control of carbonate sedimentation makes it possible to model carbonate sedimentation in one dimension; the fact that many shallow-water platform sequences are laterally continuous but show rapid vertical variation makes it natural to do so. Vertical shallowing-upward sequences modelled numerically vary in thickness from a few metres (e.g. Goldhammer *et al.*, 1987; Goldhammer & Harris, 1989) to a few tens of metres (Watney *et al.*, 1991). One-dimensional carbonate modelling has sought to explain these sequences in terms of superposition of cyclical eustatic forcing and depth-dependent sedimentation (Goldhammer *et al.*, 1987, 1993; Spencer & Demicco, 1989; Watney *et al.*, 1991). The dependence of model results on the form of production function assumed is illustrated by the work of Cisne *et al.* (1984), who assumed that in shallow water production rate increases with water depth. This allows for a steady-state solution in which water depth is proportional to the rate of relative sea-level rise, a result reminiscent of Schwarzacher's (1966) steady-state solution and Pitman's (1978) model of shoreline

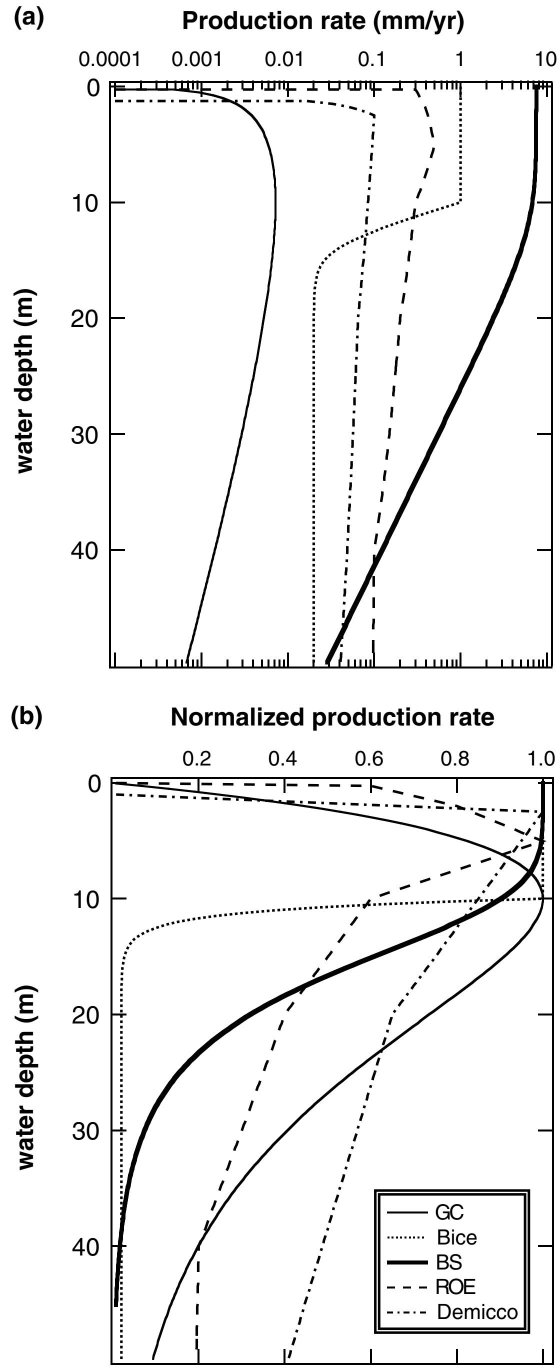


Fig. 6. Carbonate production functions used by various authors. (a) Sedimention rate in dimensional units, to show range of rates used, and (b) rate normalized to the maximum rate to show variation in shape. GC – Gildner & Cisne (1990), Bice – Bice (1991), BS – Bosscher & Schlager (1992), ROE – Read *et al.* (1991), Demicco – Demicco (1998). The low rates represent what Gildner & Cisne (1990) termed 'epeiric sea' conditions, the high rates 'reef' conditions.

position. A regime of positive depth dependence is included by Demicco & Spencer (1989), Bosence & Waltham (1990), Read *et al.* (1991), Goldhammer *et al.* (1993) and Demicco (1998). These and the more general model of Gildner & Cisne (1990) have a strongly decreasing production function below a depth of maximum production (Fig. 6). Depths of maximum production suggested by the above authors are typically in the range of a few metres. The increasing-production regime is absent from the semi-empirical reef model of Bosscher & Schlager (1992), which was adopted by Bice (1988) and Galewsky (1998).

Carbonate platforms also show significant lateral variation, particularly near their margins. Two-dimensional models have accounted for such features by including geometric treatment of sediment redistribution by physical transport. The strong depth dependence of carbonate production tends to produce steep-edged platforms (Fig. 7; Aigner *et al.*, 1989; Strobel *et al.*, 1989; Read *et al.*, 1991; Bosscher & Southam, 1992). Weakening the depth dependence of carbonate production, and/or strong offshore transport by physical processes, reduces this 'sharpening' tendency, leading to ramp rather than platform geometries (Aurell *et al.*, 1995, 1998). Progradation of carbonate platforms by physical transport produces clinoforms whose shape is related to the interplay of eustasy, transport and subsidence (Bice, 1988, 1991; Spencer & Demicco, 1989; Lawrence *et al.*, 1990; Bosence *et al.*, 1994; Galewsky, 1998). It seems that simple models with plausible parameter values can reproduce observed stratigraphic sections (Demicco *et al.*, 1991; Harris, 1991; Bosence *et al.*, 1994) and seismic patterns (Aigner *et al.*, 1989; Shuster & Aigner, 1994). It is also clear that the solutions are nonunique (Demicco *et al.*, 1991; Bosence *et al.*, 1994; Aurell *et al.*, 1995, 1998). A systematic approach along the lines of that described by Heller *et al.* (1993) for clastic systems would be a way of constraining this nonuniqueness. A characteristic feature of carbonate platforms related to carbonate production dynamics is the possibility of migration autocycles (Ginsburg, 1971) in which migration into deepening water leads to drowning (Schlager, 1981) and initiation of a new clinoform. This can be imposed via a 'drowning depth' (Demicco, 1998) but an analytical production function that decreases strongly with depth allows the autocycles to occur spontaneously (Galewsky, 1998).

Clastic and mixed systems

Geometric modelling of clastic systems started off in two dimensions (Pitman, 1978), following the original framework set out by Sloss (1962). Cant (1989) proposed a one-dimensional geometric model for clastic sequence stratigraphy, but incorrectly concluded from it that the causes of sequence boundaries cannot be distinguished using the stratigraphic record. Collier *et al.* (1990) have shown that a geometric shoreline model only slightly more complicated than Pitman's can reproduce sequences comparable to well-dated Quaternary sequences where the forcing is known. Comprehensive, two-dimensional geometric models for clastic sequence stratigraphy have been described by Jervey (1988) and Ross *et al.* (1995). The Ross *et al.* model focused on partitioning mud and sand among fluvial, shelf and turbidite systems and stands out as the first to separate the shoreline from the shelf–slope break. Both papers show how a geometric model based on sea-level, sediment supply and subsidence can be used to reproduce patterns in both lithostratigraphy (Fig. 8) and time–space diagrams. Jervey's approach was adopted in the SEDPAK model (Kendall *et al.*, 1991a, b), whose workings are explained in Strobel *et al.* (1989). SEDPAK is noteworthy for including both clastic and carbonate sedimentation. Like Jervey's and many other sequence-stratigraphic models, it is biased towards sediments rather than subsidence, although SEDPAK's treatment of subsidence goes as far as including basic flexural effects. In the sediment column it includes compaction along with deposition in fluvial, clastic-shelf, carbonate-shelf, continental-slope and deep-marine environments. The carbonate section in particular is fairly elaborate. Because of its many geometric parameters, SEDPAK requires a good deal of input from the user, and is highly tunable. Models like this are very hard to test, but in a comparison with seismic stratigraphy of the Great Bahama Bank (Eberli *et al.*, 1994), SEDPAK reproduces major features of the observed section using the Haq *et al.* (1987) sea-level curve and reasonable parameter values (Fig. 9a). SEDPAK is a simulation rather than an analytical model. It is, in effect, a very flexible tool for converting ideas about depositional geometry and external forcing into stratigraphy. In this regard, it has been useful as a way of systemizing and refining conceptual sequence-stratigraphic models (Helland-Hansen *et al.*, 1988).

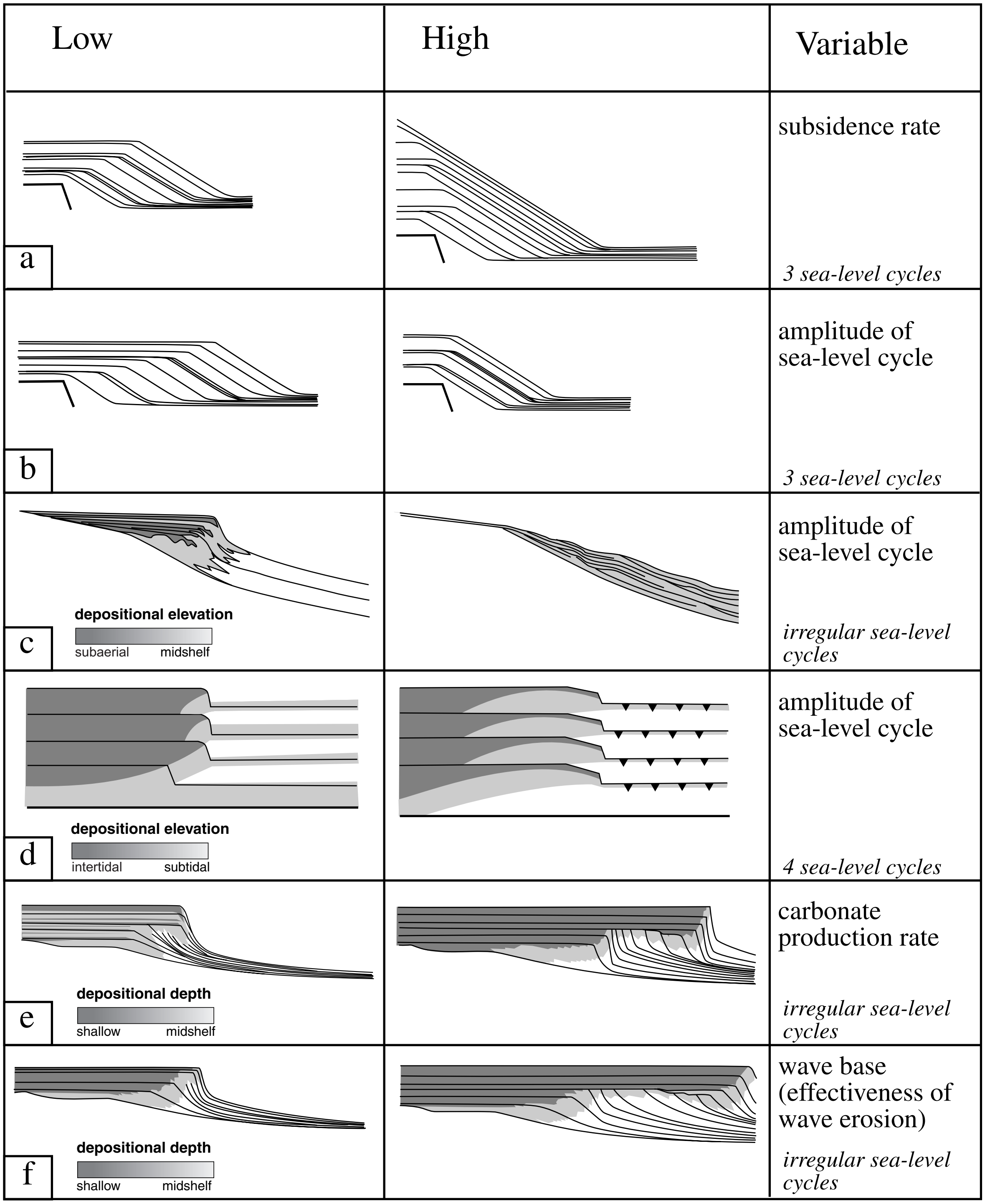
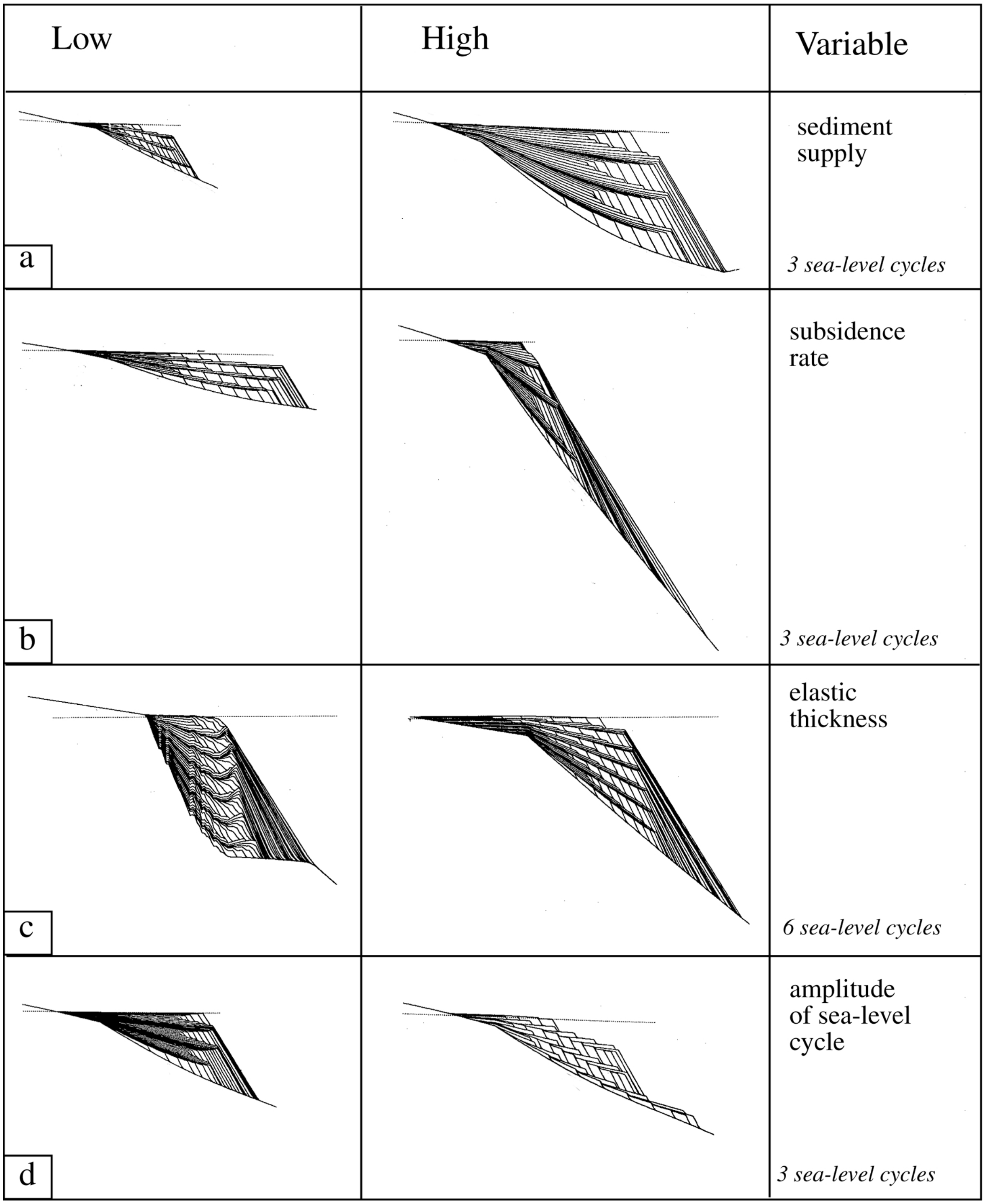


Fig. 7. A catalogue of carbonate stratal geometries produced from various models under various scenarios. In each case the variable being changed between the left image and the right image is given on the far right along with (in italics) the imposed forcing. Sea-level cycles are sinusoidal unless stated otherwise. The continuous lines are equally spaced time lines. Where an additional variable has been mapped onto the time lines it is shown in grey, with a relative scale bar below. The figures are redrawn and simplified from model results to give an idea of the stratigraphic pattern; readers should consult the original works for details. (a) Effect of changing subsidence rate (Bice, 1991); (b) effect of changing amplitude of sea-level fluctuation (Bice, 1991); (c) effect of changing amplitude of sea-level fluctuation (Read *et al.*, 1991); (d) effect of changing amplitude of sea-level fluctuation (Spencer & Demicco, 1989) – sawteeth indicate subaerial exposure; (e) effect of changing carbonate production rate (Bosence *et al.*, 1994); (f) effect of changing wave base (Bosence *et al.*, 1994).



Perlmutter *et al.* (1998) provide a good example of using SEDPAK to formalize and sharpen a conceptual model. The idea is that climatically driven eustatic cycles are inherently linked to various other changes, especially in sediment supply (Perlmutter & Matthews, 1990). The latter may vary in phase relative to sea-level over 180°. The SEDPAK application allows one to visualize the effects of changing the phase of sediment supply relative to sea-level (Fig. 8g). As would be expected, the main effect is to change the partitioning of deposition between fluvial and submarine, with the maximum fluvial deposition occurring when the supply maximum coincides

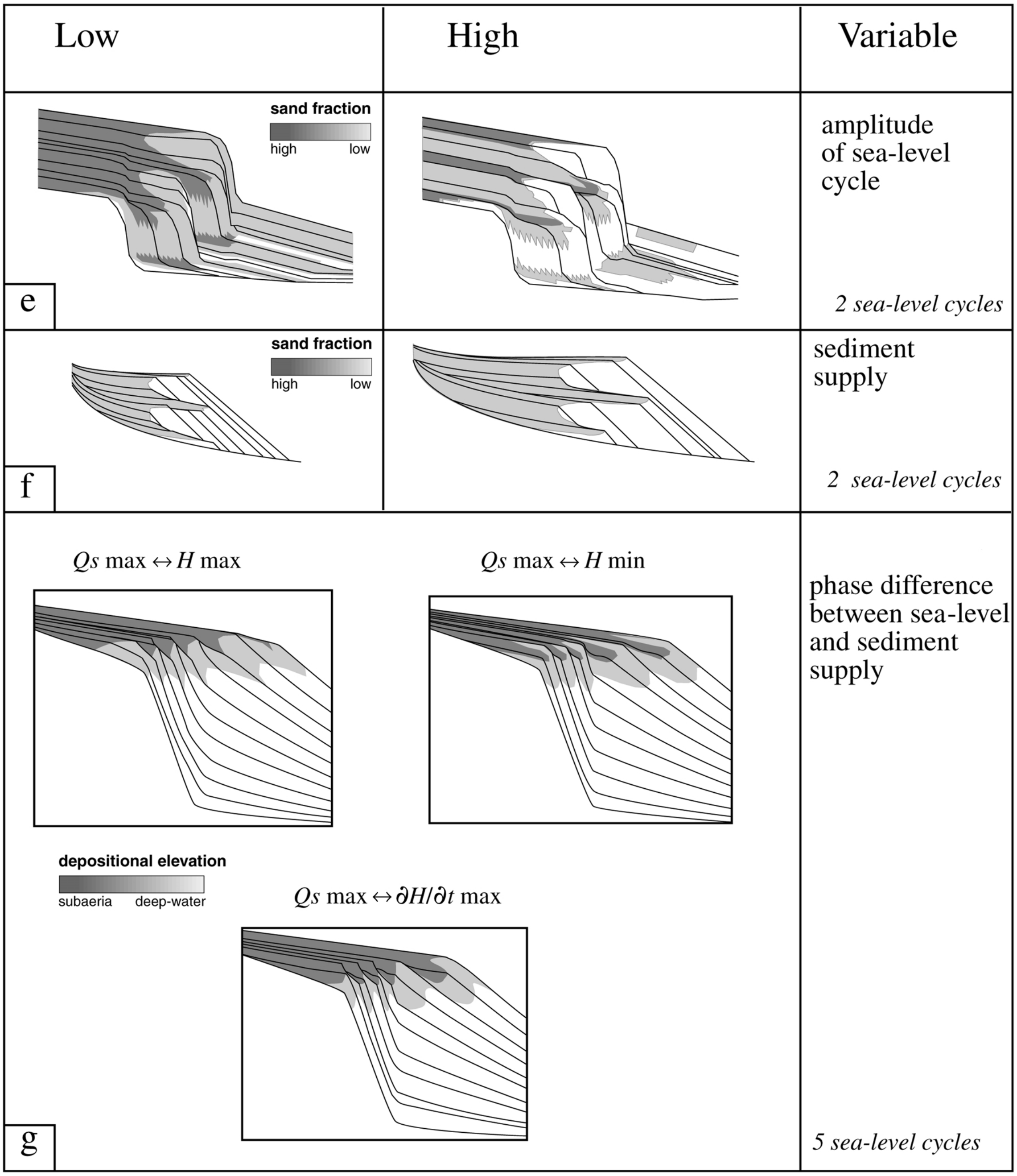


Fig. 8. A catalogue of clastic stratal geometries produced using geometric surface models under various scenarios. Layout, disclaimer and explanation are the same as for Fig. 7. (a) Effect of changing sediment supply; (b) effect of changing subsidence rate; (c) effect of changing elastic thickness; (d) effect of changing amplitude of sea-level cycle (a–d from Steckler *et al.*, 1993); (e) effect of changing amplitude of sea-level cycle (Ross *et al.*, 1995); (f) effect of changing sediment supply (Jervey, 1988); (g) effect of changing relative phase of sediment-supply and sea-level cycles (Perlmutter *et al.*, 1998).

with the maximum rate of sea-level rise. A more recent, and more adventurous, SEDPAK application is an attempt (Sen *et al.*, 1999) to date sedimentary sequences by pattern matching with the model, along the lines of magnetic-reversal stratigraphy. The idea is to generate model stratigraphies from the Haq *et al.* (1987) eustatic sea-level curve over various time intervals, then to date the section by choosing the interval that gives the best match to the real stratigraphic pattern. This is potentially a useful approach but requires a good deal of faith: in the reliability of the sea-level curves, in the accuracy of the model and in the assumption that changes in other forcing factors, especially subsidence rate, can be neglected.

Cant (1991) and Nummedal *et al.* (1993) have presented geometric models of sequence development on clastic continental margins aimed at, among other things, resolving ravinement surfaces formed by shallow submarine erosion. Cant does this by using curved depositional surfaces. He also suggests changing the geometry for times when the surfaces represent 'disequilibrium' conditions. This is a step in the direction of dynamic models that also illustrates the limitations of geometric modelling. In Nummedal *et al.* (1993) the shoreface migrates via 'Bruun's rule' (Bruun, 1962), a mass-conserving geometric model of shoreline retreat. This rule, although developed for engineering rather than stratigraphic purposes, is nonetheless a particularly nice example of using a geometric approach to capture the main effects of very complex physical processes, in this case sediment redistribution in the surf and shallow-offshore zones.

The SEQUENCE model of Steckler (1999) is the latest in a series of quantitative sequence-stratigraphy models that evolved from the rigid-lid approach. They are still 'geophysical' in spirit, coupling a dynamic treatment of subsidence, including both isostatic and flexural effects, to a geometric model of the sediment surface. A geometric clinoform model was used in this way by Reynolds *et al.* (1991) to show that variation in rigidity could lead to different stratigraphic sequences for the same eustatic forcing. Reynolds *et al.* (1991) and Steckler *et al.* (1993) clarified the independent roles of sea-level, tectonics (subsidence) and sediment supply in producing sequences (Fig. 8a–d). Their model also generates phase shifts in the timing of sequence-boundary formation reminiscent of the original Pitman (1978) result, although in this case dependent on amplitude and cycle time of the eustatic fluctuations, and on sediment supply. Two important additions in SEQUENCE, the most recent development in this line (Steckler, 1999), are a provision allowing for separation between the shoreline and the shelf-slope break (Ross *et al.*, 1995), and explicit accounting for the time needed to reach (flexural) isostatic equilibrium. The latter results from the need to accommodate changes in loading by viscous flow on timescales of the order of 10^4 years (e.g. glacioeustatic sea-level fluctuation). This is a timescale on which the relaxation effects originally examined by Beaumont (1978) are important, and it introduces another mode for dependence of stratigraphic response on cycle time. Separation of shoreline and the main clinoform break leads to phase differences between the two, for instance progradation of the shelf-slope break while shoreline is transgressing (Steckler, 1999).

Another model that uses geometric sedimentation coupled to dynamic subsidence is the Shell model described by Lawrence *et al.* (1990). This model also treats both carbonate and clastic sedimentation. Inclusion of subsidence mechanics allows it to produce some of the dynamical features of the early 'rheological' rigid-lid models, but with the addition of a full suite of sedimentary environments. Lawrence *et al.* (1990) and Aigner *et al.* (1989, 1990) show that by tuning their model (i.e. adjusting parameter values to achieve a best fit to observations) they can reproduce a variety of observed stratigraphic sections (e.g. Fig. 9b). This approach is continued by Shuster & Aigner (1994), who also generate synthetic seismic sections from the model output. Lawrence (1994) describes a more analytical application of the same model in which stratigraphic sequences are produced solely by varying sediment supply, as opposed to the usual eustatic forcing. This case also nicely illustrates one of the key differences between geometric and dynamic models: a dynamic simulation of variable sediment supply would also change the fluvial slope as a function of sediment supply, potentially producing updip unconformities during intervals of low supply (Fig. 4; Paola *et al.*, 1992). Nonetheless, the Shell model is one of the most complete stratigraphic models that exists today. It is unfortunate that it is not available for general use.

It is clear that one can model a number of stratigraphic features using predefined surface geometries. Particularly nice summaries of the physical basis for doing so are provided by Ross *et al.* (1995) and Nummedal *et al.* (1993). Nonetheless, there is no avoiding the fact that the surface morphology of basins is determined dynamically by the interplay of transport processes and subsidence. One of the immediate consequences of modelling sediment-surface dynamics (so-called 'morphodynamics') is that in a system with steady forcing and subsidence, the surface topography evolves towards a steady state in which sedimentation everywhere balances subsidence. This condition is the depositional

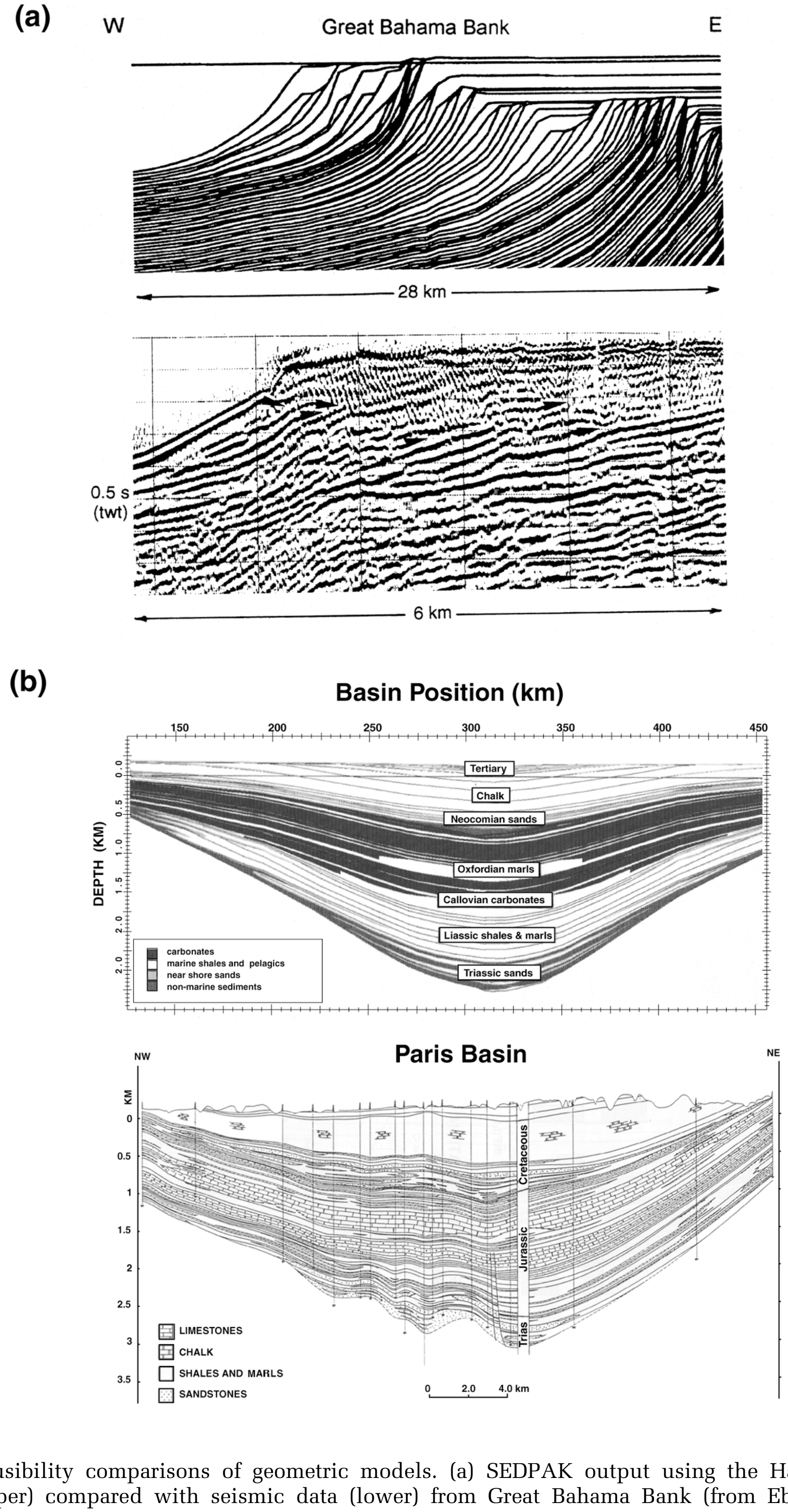


Fig. 9. Plausibility comparisons of geometric models. (a) SEDPAK output using the Haq *et al.* (1987) sea-level curve (upper) compared with seismic data (lower) from Great Bahama Bank (from Eberli *et al.*, 1994; © 1994, reprinted by permission of the American Association of Petroleum Geologists, whose permission is required for further use); (b) Shell model output (upper) using the Haq *et al.* (1987) sea-level curve and two iterations of tuning sediment parameters vs. stratigraphy (lower) from the Paris Basin (from Aigner *et al.*, 1989).

equivalent of the classical graded river. Let us call the timescale required to reach equilibrium T_{eq}; techniques for estimating it and numerical values will be discussed later. The point here is that dynamic variation in surface geometry becomes relatively less important as the period of forcing approaches and exceeds T_{eq} (strictly speaking this also depends on the amplitude of the forcing). Hence the formal justification for the geometric modelling approach is strongest for timescales greater than T_{eq}. As will be discussed below, in many cases higher-order (i.e. third order and above; Emery & Myers, 1996) eustatic oscillations occur on timescales that are below typical T_{eq} values, especially for large basins. This leads to potentially important effects associated with nonequilibrium response, which can be seen both theoretically (Jordan & Flemings, 1991; Swenson *et al.*, in press) and experimentally (Wood *et al.*, 1993, 1994; Koss *et al.*, 1994).

Dynamic models

Fluvial systems and the diffusion equation

Not surprisingly, the first dynamic stratigraphic models were for fluvial systems. The most commonly used equation has the form of the standard equation for the diffusion of heat or material and hence is usually called the diffusion equation:

$$\sigma + \frac{\partial \eta}{\partial t} = \frac{\partial \eta}{\partial x}\left(\nu \frac{\partial \eta}{\partial x}\right) \qquad (1)$$

where σ is the subsidence rate, η the elevation of the sediment surface, t time, x downstream distance and ν the transport (diffusion) coefficient (Fig. 10). It is worth stressing that use of this equation does not mean that the sediment diffuses in a real physical sense. Rather, the combination of one-dimensional mass balance

$$\sigma + \frac{\partial \eta}{\partial t} = \frac{-1}{C_0}\frac{\partial q_s}{\partial x} \qquad (2)$$

where q_s is the mean sediment flux per unit width of basin and C_0 the volume sediment concentration in the deposit, with a slope-dependent sediment flux

$$q_s = -\nu' \frac{\partial \eta}{\partial x} \qquad (3)$$

where $\nu' = \nu C_0$ gives an equation that has the same form as that of true diffusion by random motion.

Various forms of the diffusion equation have long been used to model fluvial morphology in engineering and geomorphology (Adachi & Nakatoh, 1969; Begin *et al.*, 1981; Garde *et al.*, 1981; Jain, 1981; Soni, 1981; Gill, 1983; Jaramillo & Jain, 1984; Ribberink & van der Sande, 1985; Zhang & Kahawita, 1987). The first workers to couple diffusion and subsidence to generate stratigraphy were Flemings & Jordan (1989). Their work built on the flexural model of foreland-basin subsidence of Jordan (1981), and represents the first fully dynamic basin-filling model. Before we look at the results of this and subsequent work, it is worth reviewing the approximations involved in deriving the diffusion equation as it applies to fluvial systems, both to appreciate its limitations and to make clear that it can be derived from basic principles of channel flow and sediment transport and does not, as is sometimes claimed, come from unrelated models of hillslopes or deltas. We will follow the derivation of Paola *et al.* (1992). Consider a fluvial system such as that shown in Fig. 10. The flow is distributed in one or more channels, whose planform is unspecified. The important assumptions are as follows:

1 The flow can be represented at each downstream location x by an average depth, velocity, shear stress, etc. This ignores the effects of both spatial and temporal variation in these parameters. For time variability, one can run a model repeatedly, imposing individual floods according to some specified distribution (e.g. Mackey & Bridge, 1995). This works, but does not provide any direct way of relating properties of the flood hydrograph to overall transport properties of the river system. One means of handling internally generated variability is a statistical–mechanics approach, based on the Reynolds decomposition from turbulence theory (Paola, 1996; Paola *et al.*, 1999). Unfortunately, as is the case for turbulence, the decomposition adds new unknowns to the analysis, which must be constrained independently. Whatever approach is used, it is clear that there is a great need for data on spatially averaged sediment fluxes in relation to channel pattern, water supply, sediment properties and biota.

2 Within particular fluvial regimes, the dimensionless boundary shear stress is constant. This is a substantial generalization of the approach presented in Paola *et al.* (1992). The dimensionless (or Shields) stress is defined as

$$\tau_* = \tau_0/[\rho g(s-1)D] \qquad (4)$$

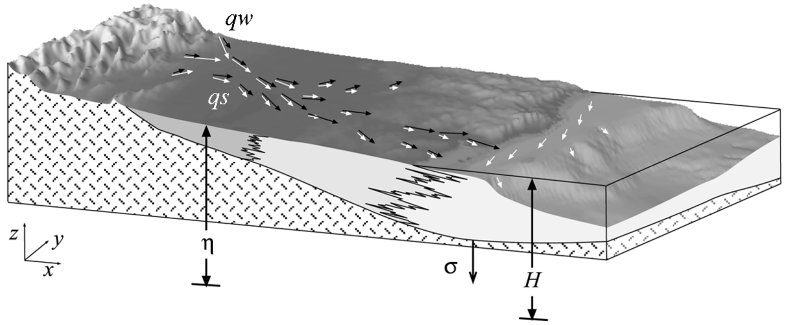


Fig. 10. Definition sketch for basin variables.

where τ_0 is the bottom shear stress, ρ the fluid density, s the sediment-specific gravity and D the median grain size. The set of basic equations for mean quantities in a fluvial system (e.g. velocity, depth) is underdetermined by one equation. The equation needed to close the system would impose a physical control on channel width. Setting τ_* constant (or, more generally, relating it to other variables in the system) supplies the additional equation and so is a way of indirectly setting the width. The idea is that limits on shear stress can be formulated directly in terms of bank erosion: if the stress is too high, bank erosion widens the channel and brings it back to its limiting value (Parker, 1978a,b). A more general stress condition would involve a dynamic equilibrium in which local bank erosion would balance local bank deposition on average.

In any case, Paola & Seal (1995) showed that τ_* estimated from mean depth, slope and grain size in coarse-grained braided rivers (data from Church & Rood, 1983) is indeed constant at about 1·4 times the critical value for initiating sediment movement (0·05) as long as the stream is not degrading and is not confined by cohesive banks or bedrock (Fig. 11). More surprisingly, Dade & Friend (1998) and Parker *et al.* (1998a) have presented data suggesting that for alluvial sand-bed streams τ_* is also constant, with a mean value between 1 and 2 (Fig. 11). The $\tau_* =$ constant approximation is rejected by Robinson & Slingerland (1998), who did not find the correlation between shear stress and grain size that would be expected if τ_* were constant. In part, this is because they applied a condition that is restricted to gravel rivers (Parker, 1978b; Paola *et al.*, 1992) to sand-bed ones (Parker, 1978a) as well. Nonetheless, their work nicely illustrates the ongoing debate about what controls river width, as well as the abrupt jump in τ_* at the gravel–sand transition in several rivers. It is worth stressing that the $\tau_* =$ constant condition in sand-bed rivers is not predicted by any existing theory.

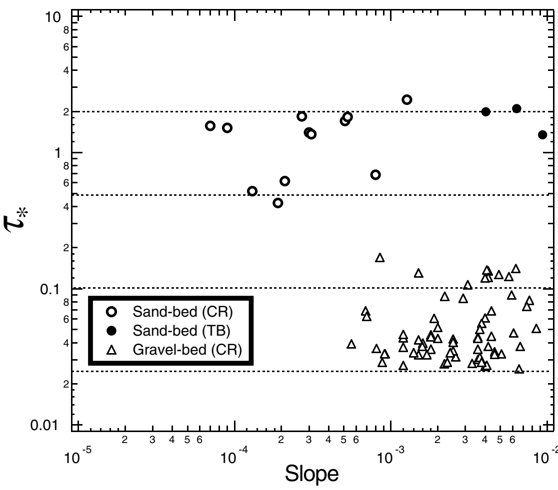


Fig. 11. Dimensionless shear stress (τ_*) for gravel-bed and sand-bed streams, suggesting two distinct regimes for fluvial diffusion models. Open points (CR) are from the data set of Church & Rood (1983). Points for both stream types were filtered as follows: characteristic flow condition, 2-year flood (preferred) or bankfull flood; channel type, meandering or higher; bank type, cohesive or higher; process code, not paved, degrading, or nonalluvial. The sand-bed data are for rivers with median size < 1 mm, the gravel-bed data from rivers with median size > 20 mm and a dimensionless stress high enough (> 0·025) to indicate at least some level of modern channel activity. Filled points (TB) are from a mine tailings basin (Paola *et al.*, 1999). Dashed lines define a factor of 2 variation about stress values of 1 and 0·05 for the sand-bed and gravel-bed cases, respectively.

3 The momentum equation for the fluid can be reduced to a balance between bed shear stress and depth–slope product: $\tau_0 = \rho g S h$ where S is the mean topographic slope and h the mean depth.

This equation can be found in most introductory sedimentology texts, but it is not always clear under what conditions it applies. Basically, the following two conditions must be satisfied: $U/gST << 1$ and $H/SL << 1$ where U, H, S, T and L are representative scales for velocity, depth, slope, time variation and length variation, respectively (there is a more complete discussion in Paola, 1996). The first condition is satisfied for nearly any realistic meteorological flood. In the second condition, H/S serves as a scale for the distance over which spatial (convective) accelerations are important. The most common manifestation of these acceleration terms is a backwater profile, so the ratio H/S may be termed the 'backwater length'. Note that the product HS scales the bed shear stress. If for a given stress H is large and S small, the backwater length is large, and vice versa. Evidently the depth–slope product as an estimate of shear stress works best for shallow, high-slope streams and worst for deep, low-gradient streams. For problems in which satisfying $H/SL << 1$ is overly restrictive, one must keep the backwater term (e.g. Snow & Slingerland, 1987; Slingerland *et al.*, 1994). The cost of this is greater computational complexity. In particular, the system of governing equations no longer reduces to a single morphodynamic equation for the sediment surface.

4 The boundary shear stress can be related to the local mean velocity via a constant drag coefficient c_f. This is not strictly correct, because the drag coefficient depends on the ratio of depth to local bed roughness. The effect of varying roughness is most likely to be important for dune-covered beds (Smith & McLean, 1977). On the whole, compared with the first three assumptions, constant drag coefficient is a relatively minor simplification. Leaving in the weak depth dependence makes the system slightly nonlinear (Parker *et al.*, 1998a).

With these four assumptions, it is easy to show that the morphodynamic equation has the diffusional form given above. The transport (diffusion) coefficient is given by:

$$\nu = \left(\frac{q_{s*}}{\langle\tau_*\rangle^{3/2}}\right)\frac{c_f^{\ 1/2}\langle q_w\rangle}{(s-1)C_0} \tag{5}$$

where q_{s*} is a dimensionless unit sediment flux determined by τ_* and a standard sediment-transport formula, and $\langle q_w\rangle$ is the average unit water discharge. An additional correction term can be incorporated in the coefficient to account for transport fluctuations (Paola *et al.*, 1999). The linear dependence of the transport coefficient on discharge means that the diffusion approximation is equivalent to setting the sediment flux proportional to the unit stream power. Apart from revealing the structure of the transport coefficient, deriving the diffusion equation from first principles also shows how, once the topography has been found, it can be used to back-calculate the depth, velocity, etc. Fluvial diffusion basically works like this: decreasing the total sediment flux Q_s decreases the total stream width because of assumption 2. Since the velocity is kept constant by assumptions 2 and 4, this means that the depth increases to conserve water. The depth increase, in turn, forces the slope to decrease to maintain the constant shear stress. Hence decreasing Q_s forces the channel system to reduce its slope and form narrower and deeper channels, all changes that in nature would tend to shift the system from braiding to meandering.

The two most important objections to the above analysis are to assumptions 2 and 3. The limitations of 3 were described above and become important chiefly for small length scales and/or large, low-gradient rivers. With regard to 2, the main alternative to imposing a stress condition is to set the width directly. This is the approach advocated by Snow & Slingerland (1987), Slingerland *et al.* (1994) and Robinson & Slingerland (1998). There is considerable empirical support for a relation of the form

$$b \propto Q_w^{\ 1/2} \tag{6}$$

where b is the width and Q_w the water discharge. The problem is that Eq. (6) has never been explained mechanistically. The use of empirical equations whose physical basis is unknown seems particularly risky for stratigraphic analysis, which may involve systems different from the modern ones on which such equations are based. However, if the stress-closure approach described above is correct, then Eq. (6) would require that sediment and water delivery systems organize themselves such that effective sediment flux Q_s varies as $Q_s \propto Q_w^{\ 1/2}$ within the gravel- and sand-dominated regimes. The lack of an explanation for how this self-organization might come about is a major weakness of the constant-τ_* approach.

It is important to note that the diffusion equation as formulated above is independent of grain size. In the diffusional context, it is tempting but incorrect to reason that the downstream slope decrease forced by deposition is directly responsible for downstream fining, on the grounds that the slope decrease reduces flow competence. Grain size comes into the picture only via τ_*, which as discussed above is some 25 times larger for sand-bed than for gravel-bed streams but seems to be roughly constant within each of these regimes. This implies that (1) within either regime downstream fining does not change τ_* and thus does not strongly influence the transport coefficient, (2) the slope decrease in the diffusional solution is forced only by deposition and is independent of grain size, and (3) that the most important grain-size change in fluvial systems is the rapid transition from gravel to sand. This transition typically occurs over a distance that is short compared with the entire length of the river (Sambrook Smith & Ferguson, 1995; Robinson & Slingerland, 1998); our group has previously approximated it as a front (Paola, 1988, 1990; Paola *et al.*, 1992). Within the gravel and sand regimes, downstream grain-size change is controlled mainly by the size distribution of the sediment supply. The τ_* constant condition implies that, to first order, it might be possible to decouple downstream grain-size variation from hydraulics, within each of the two size regimes (gravel and sand). This approach, which was adopted by Paola & Seal (1995), would be a powerful simplification of present size-sorting models (Parker, 1991; Bridge & Bennett, 1992; van Niekerk *et al.*, 1992; Hoey & Ferguson, 1994; Cui *et al.*, 1996). These require knowledge of the width and hence are difficult to apply on geological timescales, on which the width is a dependent variable.

An alternative treatment of grain-size variation in fluvial systems is to assign individual transport coefficients to sediment types (Rivenæs, 1992, 1997) and then solve a set of coupled diffusion equations for each type simultaneously. This approach produces a smooth, continuous downstream increase in the fraction of the more mobile type (i.e. higher transport coefficient). It also predicts that the effective transport coefficient (i.e. the composite value resulting from interaction of the various sediment types) increases smoothly and continuously downstream. This is because the transport coefficient is modelled as being a property of the material, rather than the stream system, and thus the coefficient tracks the material that carries it. For gravel–sand sorting, this continuous downstream variation is directly contrary to observed well-defined gravel–sand transitions. Since the size dependence in the transport coefficient enters via τ_* (Eq. 5), continuous downstream change in the overall transport coefficient also implies continuous downstream change in τ_* and thus is directly contradicted by the data presented in Fig. 11. The multiple-diffusivity approach might work for sand–mud sorting, particularly for settings like the continental shelf where it can be argued that the transport process really is diffusional (discussed below). In most river systems, it seems more likely that overall sand–mud sorting is controlled by the interaction of channels and floodplains (e.g. Mackey & Bridge, 1995; Ross *et al.*, 1995). This process is not treated physically in any existing diffusional models.

Robinson & Slingerland (1998) have for the first time coupled a physically based sorting model (van Niekerk *et al.*, 1992) to basinal fluvial aggradation. The depositional model in this case is not diffusional, because it uses a direct width closure (point 2 above), and because it retains the full backwater term (point 3 above). This work nicely illustrates the roles of both the rate and the spatial distribution of subsidence in controlling fining rate for a known input size distribution, and it is particularly valuable for showing how a physically based model can be used to compute quantities (here sedimentation rate and channel width) that can be estimated independently from the stratigraphic record. This is a significant step towards more testable stratigraphic models.

Results from diffusion-based fluvial models

Diffusional fluvial models have primarily been applied to the stratigraphic evolution of foreland basins. These have been coupled models that built on the approach developed by Flemings & Jordan (1989). In their model, both erosion of the mountain belt and the associated basin filling occur via linear diffusion, with a much lower transport (diffusion) coefficient in the erosional area than in the depositional basin. This leads to steeper slopes in the erosional area. The basement deformation that controls uplift and subsidence is controlled by the classical flexure equation, and the advancing orogenic wedge is assumed to be critically tapered. Viscous relaxation is ignored, for the reasons explained earlier. The Flemings and Jordan model is analytical as defined here; it was used primarily to explore fundamental behaviours rather than to simulate specific situations. It remains one of the most widely cited papers in basin modelling, and for good reason: it presents a model that has just enough of the relevant physics to be useful, and to reveal important general trends, but not so much detail that it has to be recalibrated for every situation. The most important overall result is that the model evolves to a state in which the thrust belt and basin form a wave that propagates with constant form, the basin forming its leading edge. The system is thus at steady state in a coordinate system moving at the shortening rate. Sedimentation balances subsidence, and input of material into the orogenic wedge balances erosion. The timescale for evolution to this steady state would be a useful reference point for studies of foreland-basin dynamics, but apparently has not yet been computed. Some of the key results are summarized in Fig. 12(a–d). At steady state, increasing the erosion coefficient makes the deposit wider, not only because more sediment is produced but because the sediment load widens the basin flexurally. Increasing the depositional transport coefficient makes the basin go from underfilled to overfilled and eventually buries the forebulge. Increasing the flexural rigidity produces the expected effect of widening the basin; increasing the thrusting rate narrows it, leading to underfilling, because basin deposits quickly become incorporated into the thrust. Flemings and Jordan were also able to produce a partially constrained comparison between their steady-state model and the basement and surface topographies of the sub-Andean foreland.

Flemings & Jordan (1990) and Jordan & Flemings (1990) went on to look at time-dependent aspects of their foreland-basin model. Key results included: (1) significant changes in basin wavelength, which might be interpreted as changes in lithospheric rheology but rather are due to sediment redistribution after an episode of thrusting; (2) the formation of piggyback basins; and (3) strong time variation in stratigraphy resulting from erosion and sediment redistribution after a thrust episode. Sinclair *et al.* (1991) used a similar coupled model to show how thickening and slowing of a critically tapered thrust wedge could produce unconformities that deepen approaching the forebulge.

The uncoupled behaviour of a diffusional system with simple geometric subsidence was examined by Paola *et al.* (1992) with the aim of understanding general aspects of diffusional basin filling. The addition of subsidence to the diffusion equation allows for a steady state in which the surface topography is adjusted so that sedimentation rate balances subsidence rate, i.e. $\partial\eta/\partial t = 0$ in Eq. (1). The time T_{eq} required to reach this condition is of the order of L^2/ν, where L is the length of the basin. T_{eq} is the basin analogue of the 'graded time' proposed by Schumm & Lichty (1965). Basin response to external forcing depends on the timescale of the forcing relative to T_{eq} (Fig. 12f–h). This result can be seen as a generalization of the findings of Pitman (1978) and Angevine (1989), and indeed it is easily shown that Pitman's tectonic timescale is approximately equivalent to T_{eq} as defined above. The main implication is that different basins can respond to the same external forcing in qualitatively different ways. Computed T_{eq} values for typical basin conditions fall within timescales of geological interest, as shown in Fig. 13.

Results from several other models with diffusional components will be discussed below under 'Coupled models'. A computer code with a graphical interface and documentation that allows one to explore many of the ideas discussed above has been made freely and readily available by Flemings & Grotzinger (1996). Beyond the computer code, Flemings and Grotzinger's model of openness in research is an important contribution to the geological community in its own right.

Alluvial architecture

In principle, there is no reason to separate the modelling of what is termed 'alluvial architecture' from alluvial basin filling in general. However, the work of Leeder (1978) discussed above set off a whole line of research specifically aimed at modelling the distribution of channel-belt sand bodies, first in vertical panels and later in three-dimensional blocks. The problem does seem to invite analysis on its own terms: fluvial sand bodies are usually distinct from enclosing floodplain material, and modelling avulsion as a stochastic process seems reasonable based on the apparently random distribution of sand bodies in outcrop. Leeder's original model was followed closely by Allen (1978), who incorporated the effect of buried fluvial topography, while later workers added effects of compaction, lateral tilting and lateral variation in floodplain sedimentation rate (Allen, 1979; Bridge & Leeder, 1979; Alexander & Leeder, 1987; Leeder & Gawthorpe, 1987; Bridge & Mackey, 1993a,b). We will refer to this whole series of theoretical models as the 'Leeder–Allen–Bridge' (LAB) models. Shanley & McCabe (1991, 1993) have coupled LAB-type models to sequence stratigraphy.

Bryant *et al.* (1995) showed experimentally that the avulsion rate can be a strong function of sedimentation rate, and Heller & Paola (1996) evaluated the effect of this within the LAB framework. At present, the most fully elaborated member of the LAB series is the model of Mackey & Bridge (1995), who have provided the first analysis of channel-belt architecture in three dimensions. The model is semi-empirical and serves to highlight both the value of considering the third dimension and the number of hard-to-constrain variables that must be added to make this possible. Unlike some cases, in which large-scale models seem to overwhelm their creators, Mackey and Bridge were able to extract a relatively simple and testable implication from their model: they predict that progessive upstream shift in avulsion points should lead to an 'avulsion sequence' in which sand bodies thin and become more widely dispersed going upward.

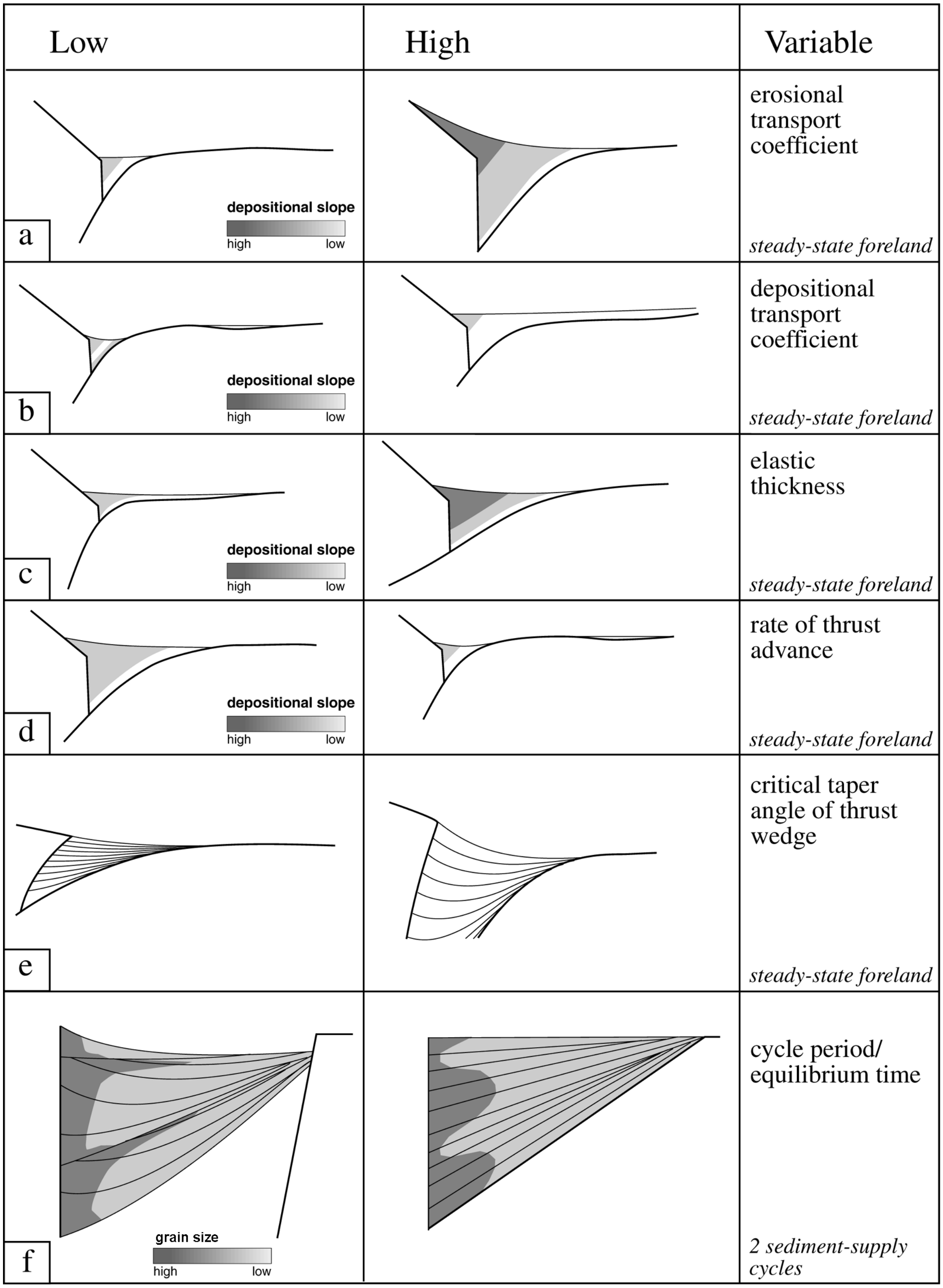
It seems likely that eventually 'architecture models' *per se* will disappear, becoming merged into overall basin-filling models. Avulsion and other channel-belt processes would retain their place as the dominant mesoscale process, but would occur spontaneously as part of the dynamics of the fluvial system. This approach would allow a unified approach to modelling the interplay between alluvial architecture and external forcings such as base-level change and variation in subsidence and sediment supply.

Outstanding problems in fluvial modelling

There are still several important pieces missing from first-order fluvial basin models. Current models lack a physical model for floodplain development, and although there have been a number of detailed studies of floodplain deposition, these do not yet provide a framework for developing a large-scale morphodynamic floodplain model. This is a first-order deficiency, since without floodplains it is impossible to model accurately the trapping of fines within fluvial systems, the downstream sorting of sandy and finer material, or avulsion. Another important point has to do with the so-called 'wash load', traditionally defined either as material finer than a cut-off small grain size (e.g. 40 μm), or as material in transport that is not present in the bed. Paola *et al.* (1999) have suggested a new definition more appropriate for morphodynamics: that the wash load is that portion of the load that does not affect the average bed slope. At present no method is available to determine the wash load defined this way, but its first-order importance is clear: any morphodynamic treatment that relates slope to sediment flux could greatly overestimate the slope by including in the sediment flux material that is in effect getting a free ride. In addition, in this view of wash load, downstream reduction in shear stress associated with both deposition and downstream fining would cause progressive conversion of wash load into effective (i.e. slope-forming) load, adding what would look like a continuously distributed sediment source to the flow. Finally, another outstanding problem in modelling fluvial systems is taking account of the effects of fluctuations in local flow conditions, either due to floods or to simple flow variability within channel systems. This was discussed earlier in the context of diffusion; the methods considered so far cover only spatial and flood-related variability. This is a specific instance of the more general problem of how to abstract the essential features of dynamics at one timescale so that they may be incorporated appropriately to the dynamics at longer timescales. We will return to this subject in the Discussion.

The continental shelf

The development of stratigraphic-scale models of fluvial systems has been paralleled by similar efforts for the continental shelf. The first step was to apply diffusional models to the continental shelf. Jordan & Flemings (1991), Johnson & Beaumont (1995) and Rivenæs (1992, 1997) have used standard linear diffusion for the shelf, assigning a lower diffusivity to represent the observed generally higher depositional slopes on shelves. Kaufman *et al.* (1991) used a depth-dependent diffusivity to account for attenuation of wave-induced bottom velocities in deep water. This depth dependence produces a rollover in the topography that was interpreted as representing the shelf–slope break. In none of these studies was any attempt made to derive the morphodynamic equation from a detailed analysis of flow and sediment transport on continental shelves. Niedoroda *et al.* (1995) have proposed a morphodynamic model of sedimentation by shallow-submarine processes. Their idea is to sum the effects of numerous short-term transport events, such as storms or river floods. These events are individually advective, but their net effect can be thought of as diffusional in aggregate. In this case,



unlike the fluvial diffusion approximation, the transport process is actually diffusive in a physical sense. The model also includes a component of advective onshore transport in shallow water. The Niedoroda *et al.* approach to diffusion works best when the events are uncorrelated, which is reasonable for shelf transport. The effective diffusion coefficients turn out to be strong functions of position and are dependent on the input meteorological and

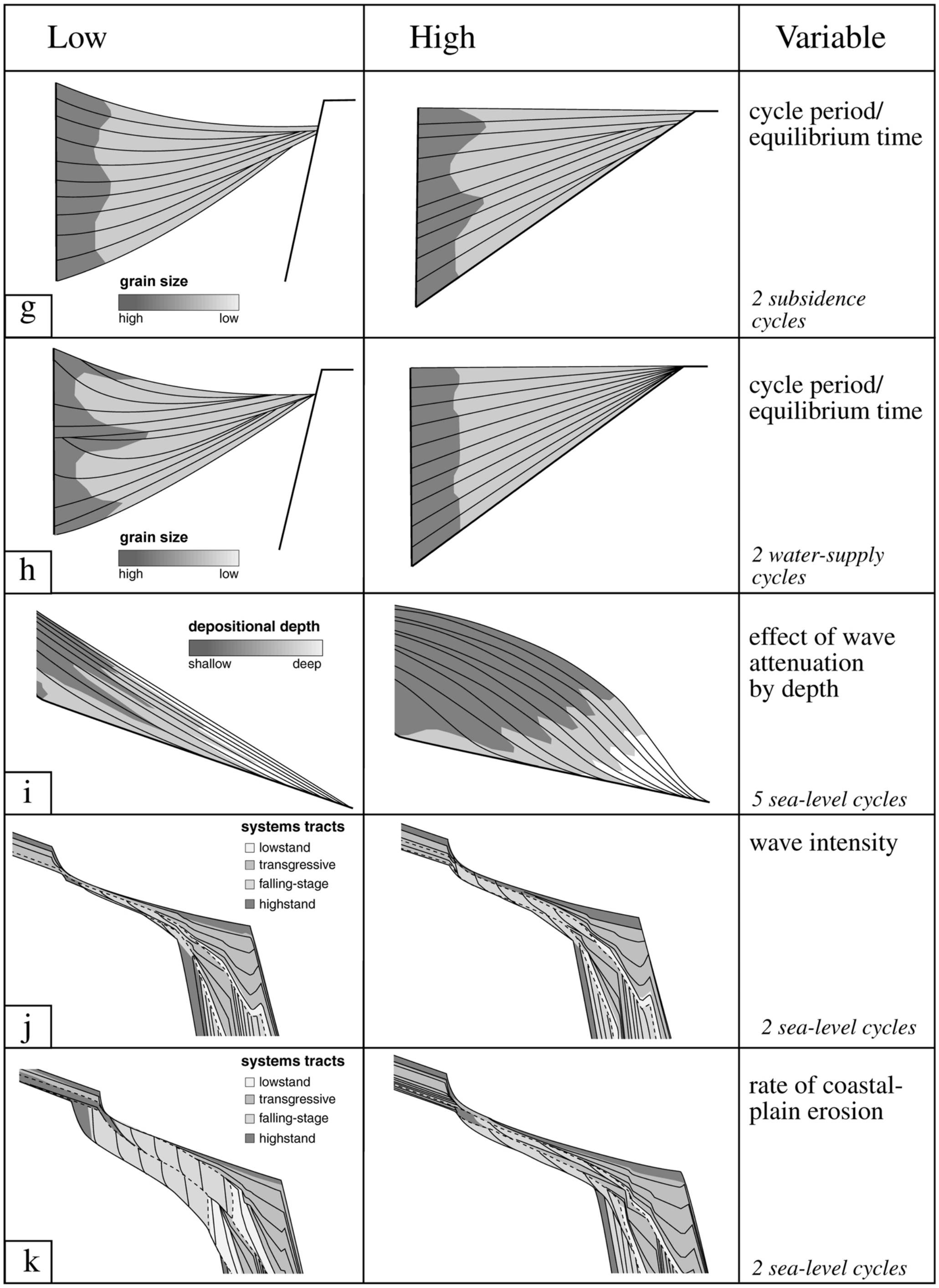


Fig. 12. A catalogue of clastic stratal geometries produced using diffusional surface models under various scenarios. Layout, disclaimer and explanation are the same as for Fig. 7. (a) Effect of changing erosional transport coefficient in thrust belt; (b) effect of changing depositional transport coefficient in foreland basin; (c) effect of changing elastic thickness; (d) effect of changing rate of thrust advance (a–d: Flemings & Jordan, 1989); (e) effect of changing critical taper angle of orogenic wedge (Sinclair *et al.*, 1991); (f) effect of changing period of imposed sediment-supply cycle relative to basin equilibrium time; (g) effect of changing period of imposed subsidence cycle relative to basin equilibrium time; (h) effect of changing period of imposed water-supply cycle relative to basin equilibrium time (f–h: Paola *et al.*, 1992); (i) effect of changing strength of depth dependence in depth-dependent diffusion (Kaufman *et al.*, 1991); (j) effect of changing wave strength in shelf advection–diffusion model; (k) effect of changing rate of coastal-plain erosion rate in shelf advection-diffusion model (j–k; Carey *et al.*, 1999).

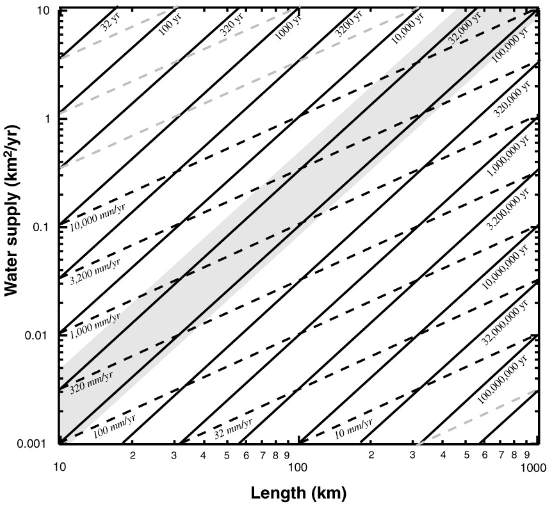


Fig. 13. Diffusional T_{eq} values (solid diagonal lines, plain labels) as a function of effective water supply and basin length. The effective water supply $\langle q_w \rangle$ is the product of rainfall rate and catchment length. The transport (diffusion) coefficient was calculated from Eq. (5) using $c_f = 0{\cdot}01$, $s = 1{\cdot}67$, $C_0 = 0{\cdot}6$, $\tau_* = 2$, and the sediment-transport formula of Engelund & Hansen (1967). The Milankovich frequency band is shown in light grey. The dashed diagonal lines show the rainfall values (italic labels) that would be required to produce the given values of $\langle q_w \rangle$ if the catchment length were equal to the basin length. Greyed rainfall lines represent implausible values.

hydrological event data used to compute them. In the present case, the greatest difficulty is that for most of the shelves for which abundant data are available (e.g. North American east coast), the present topography is substantially out of equilibrium, owing to the recent rapid rise in sea-level. The effect of the underlying topography on the computed effective diffusion coefficients is unclear but could be substantial. The Niedoroda *et al.* model has been coupled to a flexural subsidence model to produce a model of the coastal plain by Carey *et al.* (1999). The result is a semidynamic model: the shelf segment is treated dynamically, but the shoreline migrates via the geometric rule proposed by Bruun (1962) and the fluvial system is handled geometrically. One can understand dealing with the complexity of the nearshore region geometrically, but now that the physical basis of diffusion in river systems is clear, there is no reason to treat that side of the coastal plain geometrically while using a dynamic model for the submarine side. In any case, the Carey *et al.* model seems to produce reasonable shelf topography and stratigraphy but remains difficult to test, like most stratigraphic models.

A prominent feature of seismic sections of clastic shelves is prograding clinoforms. The simplest way to make a clinoform is by delta progradation (Kenyon & Turcotte, 1985; Syvitski & Daughney, 1992; Driscoll & Karner, 1999). However, the clinoform reconstructions of Steckler *et al.* (1999) suggest that the clinoform break is often in water depths of several tens of metres and hence is not a shoreline. An advective–diffusive submarine clinoform model was proposed by Pirmez *et al.* (1998). Here the advective component of the flow is a residual river current that drops sediment as it expands into deepening water, so the clinoform is still essentially deltaic. This produces the desired morphology, and is a step towards a more realistic shelf model in which waves and marine currents are the main agents that transport sediment. The advective–diffusive model of Niedoroda *et al.* (1995) can also produce clinoforms, but apparently only under conditions in which the rate of sea-level change is high relative to sediment supply (Carey *et al.*, 1999).

Since several shelf models appear to have adopted the diffusion equation with little or no dynamical justification, it is worth pointing out that there is no compelling *a priori* reason to choose diffusion as a first-order morphodynamic model. One can derive a diffusion equation either by invoking transport mechanics that is actually diffusive (Niedoroda *et al.*, 1995) or by showing that sediment flux is controlled by bottom slope, i.e. $q_s = f(S)$. The latter approach makes sense for systems like rivers (or turbidity currents) that are gravity-driven. For cases like the continental shelf or large aeolian systems in which the flow is driven by externally imposed pressure gradients, there is no particular reason why the bottom slope should be the main control on sediment flux. For clastic shelves, the dominant role played by clinoforms (Pirmez *et al.*, 1998; Steckler *et al.*, 1999) suggests that a wave equation might be a more natural choice than a diffusion equation. The Pirmez *et al.* (1998) clinoform model is really fluviodeltaic rather than marine, but it is not hard to see how one might develop a comparable wave-current model for shelves by simplifying the dynamics considered by Niedoroda *et al.* (1995). Assume that the characteristic event that moves sediment across the shelf is storm-induced downwelling. The cross-shelf flow in this case is due to the wind stress applied to the surface. Ignoring waves for the moment and confining attention to the inner shelf, where the effects of stratification and the Coriolis force can be neglected, the bottom stress is proportional to the wind stress. If the latter is constant, so are the bottom stress and (approximately) the near-bottom current, independent of depth or slope. As in the Niedoroda *et al.* model, it is waves coupled to this downwelling flow that give rise to spatial stress variation, via the exponential dependence of bottom orbital velocity on water depth. This suggests that to first order the structure of the sediment flux on the shelf could be $q_s = f(h)$, rather than $q_s = f(S)$.

Since the water surface is nearly horizontal, variations in water depth h and bed elevation η are linked via $d\eta = -dh$. Thus with a depth-controlled flux q_s we have

$$\frac{\partial \eta}{\partial t} = \frac{1}{C_0}\left(\frac{dq_s}{dh}\right)\frac{\partial \eta}{\partial x} \qquad (7)$$

This describes a wave that propagates in one direction, with constant form if dq_s/dh is constant. In contrast, diffusion always tends to spread and flatten topography. The wave speed c in Eq. (7) is given by $c = (1/C_0)dq_s/dh$, which would be negative on the shelf. Thus the wave propagates offshore. The equilibrium time T_{eq} for the wave case is given by $T_{eq} = L/c$ where L is the length of the shelf. On the shelf, q_s should decrease faster than linearly with h, in which case the shallower parts of the bed wave move faster than the deeper ones. This produces a kinematic instability in which the wave steepens until some independent slope-related failure criterion is exceeded, at which point a failure-dominated front forms. The same instability has been invoked in the case of bedforms by Middleton & Southard (1984) to explain why they have steep lee faces. In the shelf case, the wave is a clinoform and the front is its steep leading edge.

The question of wave-like (advective) vs. diffusive behaviour has been somewhat muddied because diffusional morphodynamics can also produce clinoforms (Jordan & Flemings, 1991; Kaufman *et al.*, 1991; Rivenæs, 1997; Carey *et al.*, 1999). For dual-diffusivity models (Jordan & Flemings, 1991; Rivenæs, 1997), the clinoform is produced by the change in transport coefficient at the shoreline and is thus basically deltaic. The origin of the wave-like behaviour in fully submarine diffusion is hidden in the spatial variability of the diffusion coefficient. Applying the product rule to (1) above, we have

$$\sigma + \frac{\partial \eta}{\partial t} = \frac{\partial \nu}{\partial x}\left(\frac{\partial \eta}{\partial x}\right) + \nu\left(\frac{\partial^2 \eta}{\partial x^2}\right) \qquad (8)$$

The first term on the right is a wave term similar to that in Eq. (7), and the second is the diffusive term. The wave term is zero if the diffusion coefficient is constant. If not, the overall behaviour of the equation is determined by which term dominates. In most cases the two terms are of the same order of magnitude, and in no case can the wave term dominate everywhere because it is not the highest-order term in the equation. The diffusive term tends to disperse wave-like topography, so the wave-like behaviour associated with spatial variation in a diffusional transport coefficient is significantly different from that produced by a wave equation. Hence, there really are two different schools of thought on shelf morphodynamics, and they should be distinguishable based on observations of shelf strata. On the one hand, there is the 'clinoform' view proposed by Pirmez *et al.* (1998), in which wave propagation dominates shelf morphodynamics. On the other hand, there is the 'diffusive' view of Niedoroda *et al.* (1995), in which clinoform development requires special circumstances (high rate of change of sea-level relative to sediment supply; Carey *et al.*, 1999). If the 'clinoform' viewpoint is correct, the wave term should be treated separately and not appear as a byproduct of spatially variable diffusion. One could retain a diffusive component but it would be uncoupled from the clinoform term. If the 'diffusive' viewpoint is correct, then the presence of clinoforms in shelf strata, which can be readily detected in seismic sections, is not a generic event but rather a byproduct of dominance of the wave-like part of Eq. (8) owing to particular conditions.

A detailed model of tidally driven flow in an epeiric sea (a shelf system interior to a continent, possibly lacking an associated continental slope and rise) was developed by Ericksen & Slingerland (1990). The approach was to adapt an existing numerical code to the boundary conditions appropriate to the Cretaceous interior seaway of west-central North America. The goal was not to produce a morphodynamic model but rather to see how effective tidal action could have been given the known palaeogeography of the interior seaway. This approach is ideally suited to a problem such as this one where attention is focused on a specific question (how strong might tidal currents have been in the seaway?). The same approach was used later to constrain the timing of uplifts that would have restricted entry of the tidal wave to a narrow Miocene seaway in southern Europe (Martel *et al.*, 1994). Apart from its importance to understanding the target deposits, this work is a reminder that there are many problems for which specialists have already developed sophisticated computational algorithms that could be adapted to stratigraphic purposes.

Outstanding problems in shelf modelling

The outstanding problem of continental-shelf research from a stratigraphic perspective is lack of a range of dynamical examples. Where for rivers there are hundreds of case studies showing how they respond to a wide spectrum of changes and inputs, for shelves we have only the single 'case' of the Holocene rapid rise in sea-level and, on shorter timescales, a handful of well-documented storms and storm beds. Morphodynamic insight on shelves will have to come from physical–oceanographic models, physical experiments, and careful analysis of the rock record – the past as the key to the present.

Excellent summaries of the state of understanding of transport dynamics on continental shelves are provided in Cacchione & Drake (1990) and Nittrouer & Wright (1994). It should be evident from the above discussion that it is not clear even what general form a first-order morphodynamic equation for siliciclastic continental shelves would take. Likewise, there does not seem to be general agreement within the physical–oceanography community about the basic processes that transport sand across shelves, even at a descriptive level. Is there a 'unit event' (presumably a representative storm, analogous to a characteristic flood in rivers) that, if iterated, could account for the first-order topography of a constructional shelf? Or is shelf sedimentation *inherently* dependent on variability of transport events, as implied in the Niedoroda *et al.* (1995) approach? Much of the uncertainty about these questions is due to the lack of data on sediment movement and associated flow dynamics for a large, sandy shelf the overall morphology of which is clearly of modern origin. Indeed, it is not even clear that such a shelf exists in the world today, though the Bering Sea, Arafura Sea and the shelf west of Taiwan seem to be possibilities. The Eel River site currently being studied in the STRATAFORM project (Nittrouer, 1999) provides an example with substantial Holocene deposition, but it is predominantly mud. Thus we have very little data on which to base a stratigraphic model of shelf sand transport and deposition. It is also worth noting that no first-order morphodynamic model has yet been proposed for tidally dominated shelves, although tidal effects were the main focus of the work on epeiric seas by Ericksen & Slingerland (1990).

Compared with the lack of a general consensus as to even the general form of the equations for overall shelf morphodynamics, the details that are missing seem less critical at this point. In almost every respect, our understanding of wave-influenced transport systems lags behind that of current-dominated systems. Physically based models for wave–current interaction and associated sediment transport, which must form the basis for any theory of sedimentation on shelves, did not even appear until the late 1970s, and these models have still not been comprehensively tested (Cacchione & Drake, 1990; Nittrouer & Wright, 1994). There is also considerable basic work to be done on wave-related bedforms, particularly for combined wave–current flows, and full-spectrum and directionally variable waves. Mud-dominated shelves are even less well understood, particularly the dynamics of so-called 'fluid muds' that may be able to transport large quantities of fine material near the bottom without complete suspension. There is also still much to be done on organism–sediment interactions in the shallow-marine realm.

Coupled models

Earlier, we considered models in which eroding compressional uplifts were coupled to an adjacent foreland basin via the diffusion equation. Here we look at models in which the primary emphasis is on the stratigraphic effects of coupling itself.

Humphrey & Heller (1995) studied the behaviour of a simplified coupled system of an eroding uplift and an associated depositional fluvial system. A unique aspect of their work is that the erosion was modelled as a first-order (wave) rather than a second-order (diffusion) process. The location of the erosion–deposition boundary was assumed to be pinned by a fault. The coupling of the two leads to a kind of

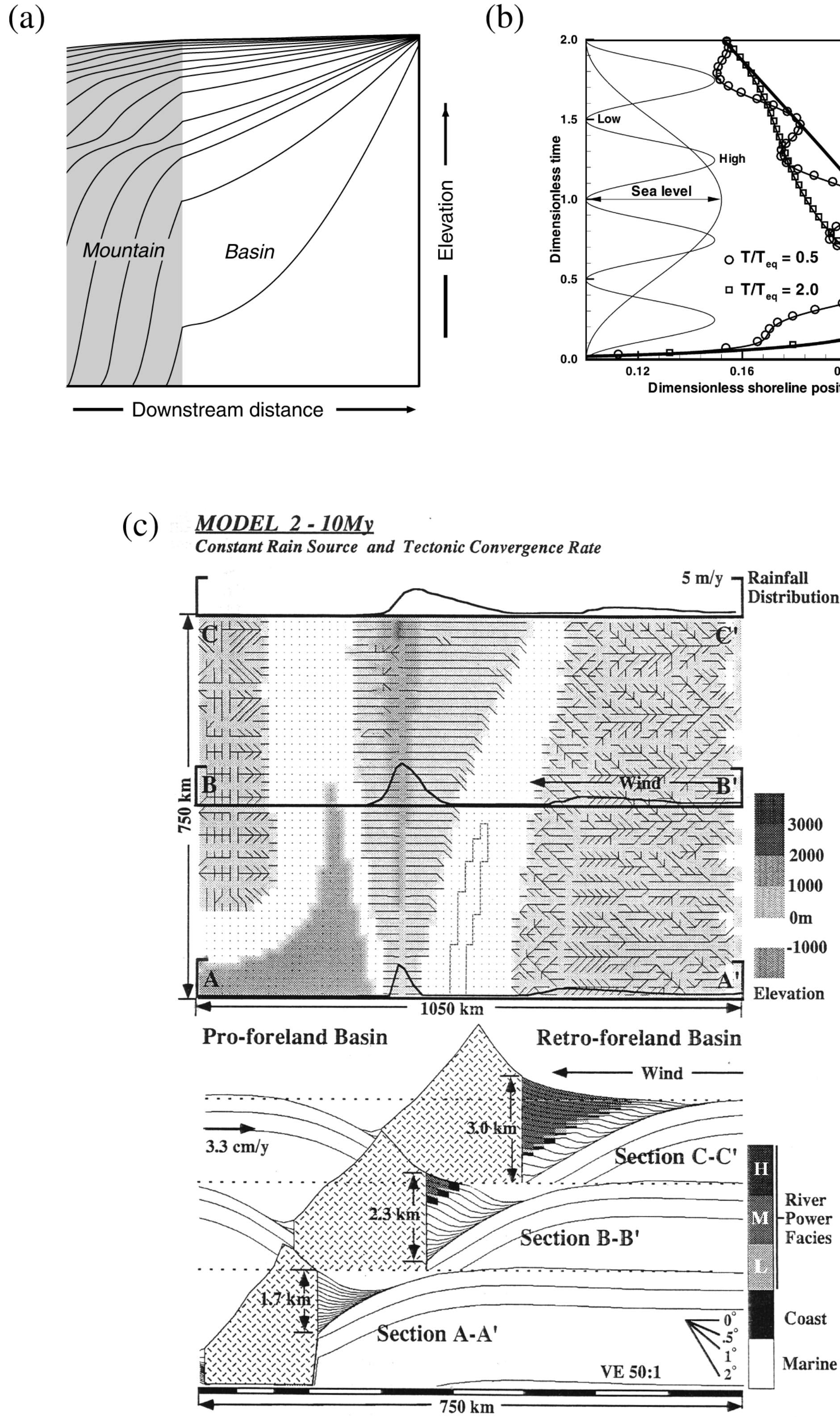


Fig. 14. Effects produced by coupled nonlinear models. (a) 'Ringing' cycles of deposition produced by a single, abrupt base-level rise (Humphrey & Heller, 1995); (b) shoreline autoretreat (heavy line) with superimposed shoreline migration cycles for low- (squares) and high- (circles) frequency sea-level variation with back-tilted linear subsidence (Swenson *et al.*, in press); (c) three-dimensional thrust-belt and foreland-basin topography produced by a coupled geodynamics/sediment-transport/rainfall model (from Johnson & Beaumont, 1995).

'ringing' in which the system responds to a step change in base level through a set of damped waves of aggradation (Fig. 14a).

A coupled model for the fluvial system and continental shelf was proposed by Jordan & Flemings (1991). The idea was to treat both the fluvial system and the shelf via linear diffusion, but with a lower transport coefficient on the shelf. As mentioned above, this model is essentially deltaic, and no detailed explanation of the coupling condition at the shoreline is given. The model produces phase shifts between shoreline position and sea-level reminiscent of those found by Pitman (1978). The results suggest that phase shifting becomes more important with increase in the ratio of cycle time to a characteristic time analogous to the T_{eq} defined above, but the paper presents only a set of case studies, from which the general form of the phase-shift function cannot be deduced. A more complete treatment of the shoreline coupling problem is presented by Swenson *et al.* (in press), who show that for a deltaic system with angle-of-repose foresets and a linear-diffusive fluvial system, shoreline migration is formally analogous to migration of a melting front in a two-phase (liquid–solid) medium heated from one end. The shoreline (phase transition) is a migrating internal boundary whose location is determined as part of the solution by specifying explicit coupling conditions at the boundary. The Swenson *et al.* model shows only minor phase shifting of shoreline relative to sea-level (Fig. 14b). There is no indication of a reduction of amplitude of shoreline response for long-period fluctuations, as was predicted by Angevine (1989). Another important result is that the model has no steady state; the shoreline tends to drift landward ('autoretreat') owing to lengthening of the submarine surface. This work highlights two key aspects of modelling coupled systems: (1) the coupling conditions are as important as the equations used to model the separate parts of the system, and (2) a system that couples linear parts with nonlinear coupling conditions can be strongly nonlinear.

The Dalhousie geodynamics group has developed the most complete model yet proposed for a foreland-basin system (Beaumont *et al.*, 1992; Kooi & Beaumont, 1994; Johnson & Beaumont, 1995). The latest version (Fig. 14c) is three-dimensional and couples erosion, deposition in fluvial and submarine systems, orographic rainfall and a geodynamic model of a convergent orogen. Hillslope transport is handled by assuming local linear diffusion. For channellized flow, it is assumed that at each point there is an equilibrium sediment flux proportional to the local stream power, i.e. equivalent to the diffusional approximation given above. Erosion is modelled as being detachment-limited in the sense that the system approaches its equilibrium sediment flux according to a first-order reaction law, i.e. exponentially over distance. The distance scale is a measure of the erodibility of the substrate. For erosion of unconsolidated material, or deposition, the distance scale tends to zero, making the model formally equivalent to diffusion as described above, but following channel axes. Submarine deposition is again handled with ordinary diffusion. In effect, the surface-process model is a composite of a detachment-limited drainage basin model, of which a number have been proposed (Willgoose *et al.*, 1991; Chase, 1992; Howard, 1994; Tucker & Slingerland, 1994), and diffusional fluvial and submarine depositional systems. The latter approach is similar to that proposed by Jordan & Flemings (1991). Apart from being the first to couple erosion to fluvial and submarine deposition in a three-dimensional model, the Dalhousie group's approach embodies two major advances: (1) it treats spatial variation in effective fluvial transport coefficient in a natural way, by letting water discharge accumulate and disperse over the three-dimensional grid; and (2) water is introduced by distributing an incident vapour flux over the terrain according to elevation and wind direction, thereby allowing for orographic precipitation effects. Having evolved from earlier geophysical subsidence models, this model also treats uplift and subsidence dynamically, using a simple elastic rheology. It is the first fully dynamic three-dimensional model of a foreland basin system. Including orographic effects highlights the importance of coupling climate and tectonics: the asymmetry between the basins on the leading and trailing edges of the orogen can be either reinforced or reduced depending on the relative wind direction.

The Stanford model

One of the most ambitious basin modelling programs to date is a three-dimensional model called SEDSIM that was developed at Stanford University. The basic algorithms and a first version of the code are presented in Tetzlaff & Harbaugh (1989). The approach is based on an adaptation of the marker-in-cell method originally developed for the study of complex, unsteady flows. The method tracks fluid units through a fixed computational grid; erosion and deposition occur as the units accelerate and decelerate owing to interactions with topography and each other (via pressure variations). Although it is very computationally intensive, this method has the advantage of allowing the flow to follow topography in a very natural way. This feature allows for development of remarkably realistic-looking patterns of flow and topography (Fig. 15). In addition to being fully dynamic and three-dimensional, the Stanford model is highly coupled, including modules for wave-influenced transport and turbidity currents along with compaction and hydrocarbon genesis and migration. Unfortunately, with the exceptions of the early version presented in Tetzlaff & Harbaugh (1989) and the nearshore model presented in Martinez & Harbaugh (1993), only general sketches of the workings of SEDSIM have been made available in the literature (Tetzlaff, 1990; Lee & Harbaugh, 1992; Martinez, 1992; Wendebourg & Ulmer, 1992; Harbaugh, 1993). In particular, no well-constrained tests of the accuracy of the model, with respect either to fluid mechanics or morphodynamics, appear to have been carried out. As such it remains an intriguing 'black box' on the basin modelling landscape.

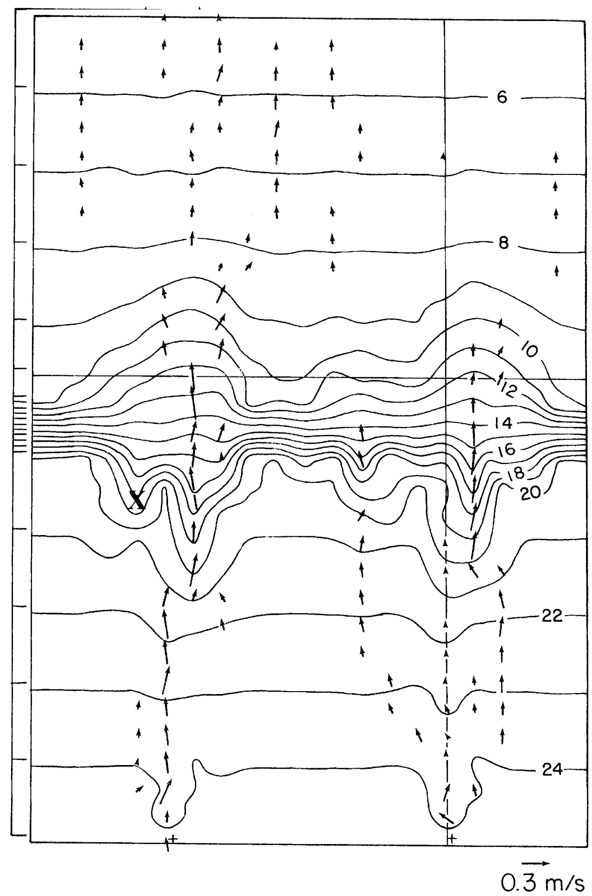


Fig. 15. Example of SEDSIM output (from Tetzlaff & Harbaugh, 1989) for development of alluvial fans and erosional valleys from an initial topographic step. Topography shown by contour lines in metres, flow by vector arrows.

The Penn State models

A wide-ranging summary of techniques for basin modelling is presented in Slingerland *et al.* (1994). Unlike SEDSIM, this is not a single, integrated model, but rather a collection of techniques for modelling specific subsystems. These include erosion and sediment supply, river flow in channels of variable width, size-selective sediment transport as bedload and in suspension, deltaic sedimentation, shelf transport by tides and storms, and compaction and subsurface flow. Aspects of the fluvial and shallow-marine parts of this collection have been discussed in previous sections. The collected work is noteworthy for including excellent introductory discussions of the derivation and physical meaning of the equations, a good deal of field data to go with the analytical material and all of the necessary computer code. Slingerland *et al.* (1994) is a toolbox, rather than a black box. It is an excellent summary of process-based basin modelling, with emphasis on doing a complete and careful job on individual components of the system.

The INSTAAR models

A comprehensive basin model called SEDFLUX has been developed at the Institute of Arctic and Alpine Research (INSTAAR), University of Colorado (Syvitski & Daughney, 1992; Syvitski & Alcott, 1993; Syvitski & Andrews, 1994; Syvitski & Alcott, 1995; Syvitski *et al.*, 1998a,b). SEDFLUX is aimed at simulating the transport and delivery of sediment on continental margins over timescales of up to tens of thousands of years. It includes the effects of eustatic variation, floods, storms and other relevant environmental factors (climate trends, random catastrophic events), using relatively short (daily to yearly) time steps. SEDFLUX combines several individual process–response models into one coupled model, distributing a multisized sediment load over a continental margin (Fig. 16). Processes currently included are river mouth dynamics, buoyant surface plumes, hyperpycnal flows, ocean storms, slope instabilities, turbidity currents and debris flows. The model includes compaction and load-induced isostatic subsidence; externally forced subsidence and other tectonic processes (faults, uplift) are predefined by the user. The modelled stratigraphy has a typical vertical resolution of 0·01–0·25 m, and a typical horizontal resolution of between 1 and 100 m.

At the moment, SEDFLUX is two-dimensional and marine. The INSTAAR group has a larger plan to develop a comprehensive three-dimensional model, at present called EARTHWORKS, that would integrate the fluvial system (erosional and depositional) as well as hydrology, thus allowing for the modelling of climatic effects. Such a model would be to sedimentary geology what global climate models are to atmospheric science. One of the best features of the INSTAAR group's approach is that it has been entirely open: thorough explanations of the algorithms, computer code, model output and the results of numerous tests of the component pieces against field data have been made available throughout development of the model series.

Unconventional models

The only explicitly stochastic basin-filling models proposed so far were introduced by Tipper (1991, 1992) and developed further by Pelletier & Turcotte (1997). Both are essentially diffusion models, but instead of invoking diffusion as a mean behaviour (i.e. Eq. 1), they model the morphodynamics as a random walk in two dimensions that behaves diffusionally in aggregate. These models are able to produce plausible-looking surface topography (Tipper, 1991) and power-law distributions of sedimentation rate as a function of averaging time (Pelletier & Turcotte, 1997).

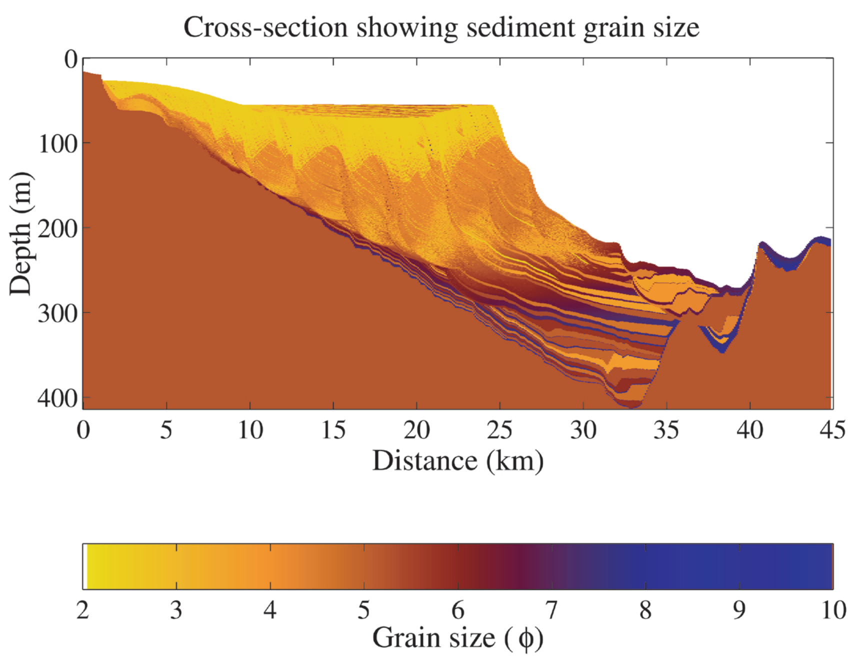


Fig. 16. Deltaic deposition as modelled with the INSTAAR model (references given in the text). The section is parallel to flow and the colours indicate grain size. The example is from a 12 000 year (daily) simulation for Knight Inlet, British Columbia. It includes the retreat of the glaciers from the ocean (about 10 kyr BP), an early to mid-Holocene warm period (to 6 kyr BP), slow cooling to the Little Ice Age (a few hundred years before present), rapid warming to the present, and then a hypothetical slower warming for 2 kyr into the future. The grain size shows the influence of a variety of processes. Hemipelagic drape from surface plumes can be seen across the entire basin. Debris flows (layers of uniform grain size) and graded turbidites occur in the deeper portions of the sub-basins, downslope from failure scarps. Delta front progradation of coarse particles from the dumping of the bedload seaward of the river mouth characterizes the upper portion of the deposits. Changes in sea-level caused by eustatic and isostatic influences of changing ice volume affect the position of the river mouth, forming delta plain and lagoonal deposits during relative sea-level rises.

FUZZIM (Norldund, 1999) is a rule-based, three-dimensional basin-filling model using fuzzy logic that has been developed and made freely available by Nordlund (1996). 'Fuzzy' logic replaces what would in traditional logic be binary choices by continuous variables. Thus 'false' (0) and 'true' (1) are replaced by a real number between 0 and 1 that expresses the degree of truth. In the basin model, this approach is implemented by defining state variables (e.g. 'far_from_source') that are true in varying degrees. Since FUZZIM is entirely rule-based, it is best thought of as a way of implementing conceptual ideas rather than a true model. Nonetheless, FUZZIM is fast, and accessible to people who are unfamiliar with numerical analysis. If nothing else, for these reasons it has great potential for introducing students to modelling. On a research level, it is not clear what is gained by replacing readily solvable equations with fuzzy-logic rules. On the other hand, there are aspects of depositional systems, such as those involving biota, where semi-empirical rules are a good starting point.

Model testing

Here 'testing' means a genuine attempt to falsify a model, as opposed to fitting it to a particular data set by varying its parameters. A true test requires that model predictions be compared with data that are independent of those used to constrain the model parameters. This rather exacting standard is very difficult to attain in historical geology, even for analytical models with relatively few parameters.

Geometric models

Because their depositional geometry must be specified by the user, geometric models are difficult to test. Hence, as described above, most of the effort in geometric modelling has gone to showing that the models can be calibrated to produce results that match real stratigraphy. One of the few geometric models that can in principle be tested in the strict sense is the original one of Pitman (1978). Although the main model prediction, phase-shifted long-term shoreline response, is relatively straightforward, it has been surprisingly difficult to test. Miller *et al.* (1993) report evidence of phase shifting of sequence boundaries relative to eustatic sea-level, but did not check quantitatively whether their cycles fit the temporal criterion for application of the model (Angevine, 1989).

Diffusional models

Again, despite their relative simplicity, the results of diffusion-based stratigraphic models have proven difficult to test. That many preserved basins show broad upward consistency in the locations of major facies boundaries (e.g. the gravel–sand transition) indicates that overall the idea of an average steady-state balance between subsidence and sedimentation is reasonable, but this is hardly a test. Flemings & Jordan (1989) showed that their coupled diffusion model could correctly reproduce the present-day surface and basement topography of the Andean foreland using reasonable parameter values. Heller & Paola (1992) provided some examples in which model-based inferences were consistent with previous interpretations of basin history based on classical geological evidence. Sinclair *et al.* (1991) modelled the development of a major unconformity with their diffusional foreland-basin model, and were able to show good agreement between computed and observed time–space diagrams for the unconformity. Although the quantitative details could not be constrained geologically, there was at least independent evidence for the nature and timing of the thickening event associated with the unconformity. This sort of constrained comparison between predicted and observed stratigraphic patterns is an important step towards real quantitative testing. On the whole, though, applications of diffusional modelling to the stratigraphic record amount to a series of arguments for plausibility, rather than true tests.

An unusual opportunity to test a diffusion-based basin model under highly constrained conditions was presented to our group several years ago. The problem involved a fluvial fan in a mine-tailings basin. The area of the basin was fixed, and accommodation was provided by raising the perimeter walls of the basin. Hence the system functioned as if it were in a basin with steady, uniform subsidence. The basin was supplied with water and sediment at a constant rate, and it developed a self-similar topography in which the sedimentation rate was everywhere constant (Fig. 17), as predicted by steady-state diffusional theory. Since all the inputs to the system were known and constant, it was an ideal case study for testing diffusion-based models. The engineering solution described by Parker

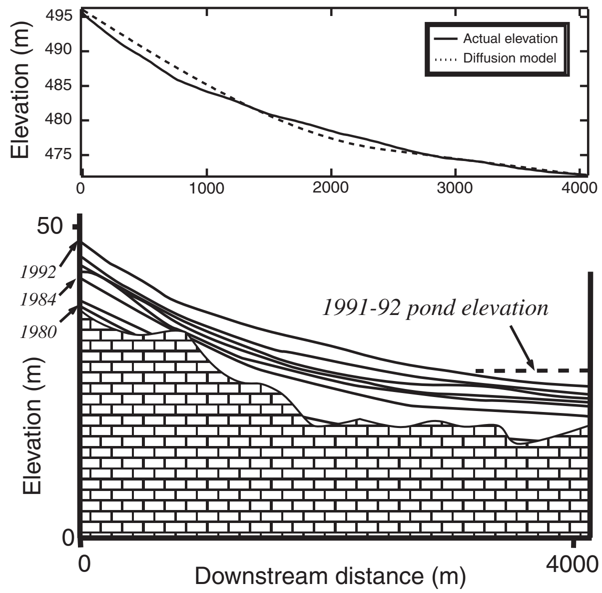


Fig. 17. Steady-state depositional profile developed in a mine-tailings basin (lower) and calculated profile from a diffusional solution (upper) using a transport coefficient computed from independent flow data (Paola *et al.*, 1999).

et al. (1998a,b) for channellized flow on the fan is, in effect, a linear diffusion equation in radial coordinates. It was fitted to the observed topography by adjusting a single parameter related to the nonlinear effect of zones of concentrated flow (e.g. channel confluences) in enhancing sediment transport. Later Paola *et al.* (1999) showed that this parameter could be estimated with reasonable accuracy from the observed distributions of flow velocity, depth and grain size (Fig. 17). Hence for this case at least, simple diffusion could be used to predict steady-state basinal topography on the basis of independently constrained input parameters.

PHYSICAL MODELS OF BASIN FILLING: A LOOK THROUGH THE DOOR

In contrast to all this theoretical modelling, very little experimental work has been done on basin filling. Most of the early experimental work in sedimentary geology focused on understanding outcrop-scale sedimentary structures. While there is still a good deal of work to be done in this area, in recent years experimental work has taken the same general direction as theory and shifted to larger-scale problems. There are now active programmes in experimental stratigraphy at Utrecht University, the University of Leeds and the University of Minnesota. Results from the latter two will be discussed below; reports of the Utrecht group's work can be found online at http://www.geo.uu.nl/Research/Sedimentology/Staff/gpostma/gpostma.html.

The main motivation for experimentation in stratigraphy is the same as in all other branches of science: it allows us to generate results under precisely known, repeatable conditions. This is especially valuable in the study of sedimentary basins, for which ancient examples usually involve boundary conditions that cannot be reconstructed precisely. Stratigraphic experiments have a particular advantage in compressing time. Much of the incompleteness of the stratigraphic record represents time when the sedimentary systems are inactive. In an experiment, the system is active all the time; for instance, in a fluvial experiment it is as if the rivers were constantly in flood. This represents a major compression of time. Continuing with the river example, if a typical natural river is activated by a flood that occurs every 1 or 2 years and lasts for a few days, the time compression from this effect alone in an experiment is more than two orders of magnitude. The price is that one loses effects such as plant growth and soil development that take place during the 'inactive' periods.

Previous work

One of the longest-running experimental programmes aimed at large-scale sediment processes is that at Colorado State University. A comprehensive summary of many of the main results is given by Schumm *et al.* (1987). Most of the Colorado State group's work has been on fluvial systems and is geomorphic rather than stratigraphic. A noteworthy exception is the work of Wood *et al.* (1993, 1994) and Koss *et al.* (1994), who studied the behaviour of fluvial systems and associated shelves in response to varying base level. Although the deposits were not analysed, the work was nonetheless aimed directly at sequence stratigraphy. Key results were: (1) shelf angle plays a major role in determining the total sediment volume derived from incision during relative fall, and hence the fraction of (presumably finer) shelf-derived material in the final deposit; (2) there are significant lags in delivery of fluvial sediment to offshore deltas associated with the finite time required for incision to propagate upstream; and (3) fluvial response to base-level fluctuation shows strong upstream decay. Points (2) and (3) also nicely illustrate the limitations of geometric theoretical models, which cannot capture this kind of wave-propagation effect, when applied to relatively high-frequency sea-level changes.

Alluvial architecture is another area in which stratigraphic experimentation is just beginning. The first experimental study directly aimed at architecture was that of Bryant *et al.* (1995), who in an unscaled experiment showed avulsion frequency to be a strongly increasing function of sedimentation rate, contradicting one of the key assumptions of the early LAB models. A much more detailed study is that of Ashworth *et al.* (1999), who modelled the internal architecture of a gravel-bed braided river (Ashburton River, New Zealand) using a 1:50 Froude-scale model (Fig. 18). The coarse-grained nature of the Ashburton system makes it an ideal target for physical modelling, and the experiments replicate both the observed distribution of subfacies in the field and the response of the Ashburton to changes in sedimentation rate. This response is a loud silence: no change in facies architecture was detectable for changes in sedimentation rate of a factor of two (laboratory) and nearly 10 (field).

There have been two major obstacles to experimentation as a tool in stratigraphy. The first is that until recently there has been no means of incorporating the effects of spatially variable subsidence into physical experiments. A solution to this is presented below. The second is that physical modelling of processes that in nature occur over large time and space scales seems to present formidable scaling problems. Since experimentation in stratigraphy is in its infancy, it is worth examining scaling issues before continuing any further.

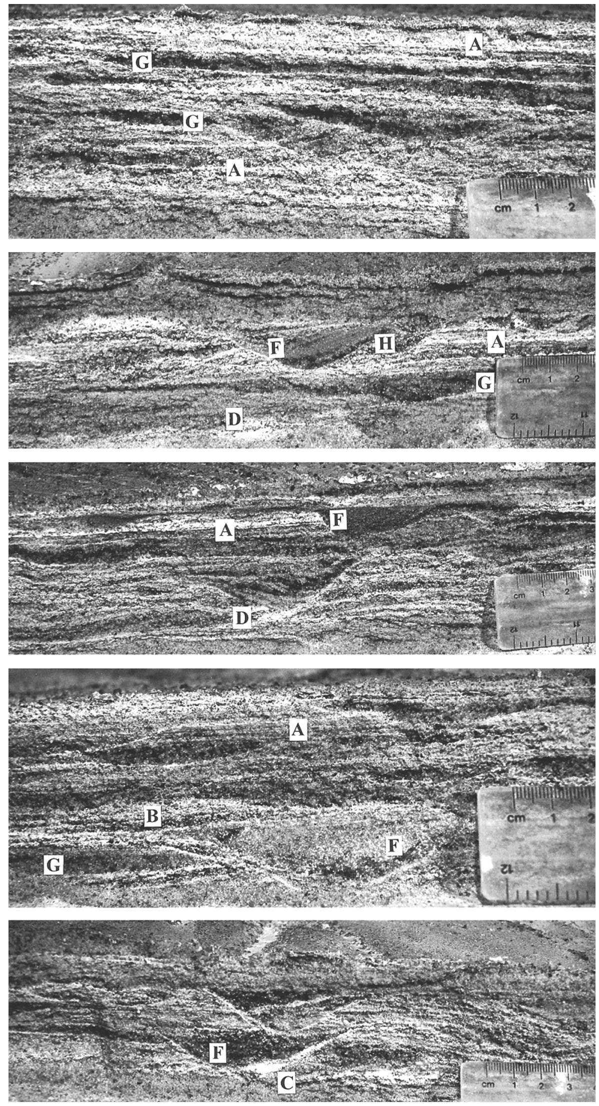


Fig. 18. Cross-sections showing depositional niches in a physical model of braided-river architecture (see Ashworth *et al.*, 1999, for details). Depositional niches identified in the flume sections are classified according to grain size for fine-grained (niches A–D) and coarser grained (niches F–H) sediments A: bar-top fines; B: bar margins; C: channel fills; D: erosional remnants; F: coarse-grained channel fills; G: bar cores; H: bar margins.

Scaling and the use of experimental results in stratigraphy

Although there is a substantial literature on the construction of dynamical scale models (Yalin, 1971; Peakall *et al.*, 1996, and References therein), in practice it is rarely possible to construct an exact scale model of any system involving fluid flow and sediment transport. For all practical purposes, water is kinematically the least viscous fluid available, which means that experimental models nearly always have significantly lower Reynolds numbers than their prototype systems. In loose-boundary experiments, it is difficult to scale down the ratio of particle size to system size because granular materials become cohesive as particle size decreases. There are a number of ways around these problems, but the key point is that scaling effects are by no means unique to stratigraphic experiments. The theoretical impossibility of creating exact scale models has not prevented engineers from solving countless design and management problems in rivers, estuaries and coastal areas with the aid of models. Beyond a certain amount of judgement and experience, the key to interpretation of the model results is that the experiments must be done in conjunction with physically based theory.

The dynamics of a system as complex as a sedimentary basin has many facets. At this point, a complete set of dimensionless numbers for basins, covering the whole range of relevant time and space scales, is not available. For a simplified basin model, Swenson *et al.* (in press) provide a set of dimensionless numbers that govern model response and stratal geometry. As the equations governing basin systems become better known, so will the relevant dimensionless numbers. Among these, some will be numbers that can readily be reproduced at laboratory scales, and some will not. The latter will tell us what aspects of the system we are missing in our experiments. At a more general level, basin processes and effects can be divided into three broad classes: (1) those that are scale-independent, (2) those that are present but distorted at laboratory scales and (3) those that cannot be represented at laboratory scales. At present, many of the first-order predictions of theoretical models are scale-independent. This is true for all geometric models (e.g. the shoreline response predicted by Pitman, 1978), which are scale-independent by definition. Neither are the existing dynamical models fundamentally scale-dependent, although some of them include effects (e.g. deposition of fine-grained sediments or floodplain formation) that are difficult to represent at small scales. In these cases, if a model is really robust, the modifications needed to apply it to a laboratory experiment should be dictated in a straightforward way by the model itself.

Experiments and theory are fundamentally interrelated in basin analysis, just as they are in engineering. Theoretical models that can successfully predict experimental results will have passed an important first hurdle. Although success in predicting experimental results does not guarantee that a model can be trusted in the field, it is very difficult to imagine trusting a model in the field if it cannot predict the results of a simplified, well-constrained experiment without a lot of tuning. Once a model has been tested, it becomes a vehicle for examining the effects of processes that cannot be scaled down to the experiments, for evaluating the effects of scaling distortions, and for examining new scenarios and suggesting further work. On the other hand, if experiments are done without a decent theoretical framework, one has no recourse but the naïve and often misleading approach of treating the experimental system as a miniature analogue.

One can also think of an experiment simply as being itself a model (Hooke, 1968). In this sense, it is similar to a theoretical model in that it involves simplified versions of the fundamental processes. The weakness of experiments is that they often involve parameters or conditions that are obviously different from natural systems. The strength, and the weakness, of theoretical models is that their parameters can be set to whatever values we want. If a model has enough of these 'knobs', it can be tuned to make its results look superficially like reality, whether it has actually captured reality or not. If we had complete confidence that the representations of natural processes embodied in our theoretical models were correct and complete, there would be no need for experiments. In my opinion, such confidence would be misplaced. Even a scaled-down water–sediment system, free to find its own transport paths and mechanisms, and to shape its own topography, is wonderfully complex. Seen as models (and not as miniatures), the advantage of experiments is that they are real, albeit simplified, physical systems. Their internal dynamics is not programmed, and they offer the possibility of new and unexpected behaviours. It is then necessary to determine, by analysis and/or field observation, if these are scale-related artefacts or not. Although experimental models of sedimentary systems are indeed simplifications, it is worth stressing that at present there is no theoretical model capable of capturing all of their dynamics.

An example experiment

Theoretical and philosophical arguments aside, the most direct way to appreciate the possibilities that experimentation has to offer in stratigraphy is to see the results of an actual experiment. Over the past 3 years the University of Minnesota group has been working to build an experimental basin with a programmable subsiding floor (Fig. 19). The subsidence mechanism is based on a honeycomb of independent hexagonal cells. The honeycomb forms the floor of an open basin. At the beginning of an experiment, the basin is filled with dry, well-sorted gravel. The top of the gravel is covered with a thin rubber membrane. Subsidence is induced by withdrawing gravel from the bottoms of the hexagonal cells. Each hexagon forms the top of a cone that tapers into an elbow pipe. Subsidence is induced by firing a

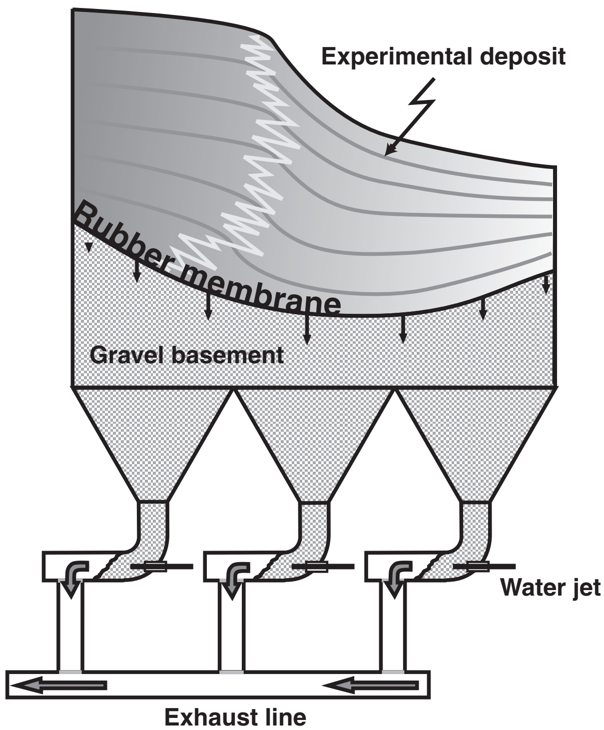


Fig. 19. Schematic diagram of the mechanism used to produce subsidence in the experimental stratigraphy facility discussed in the text. The subsidence cells are 0·4 m wide.

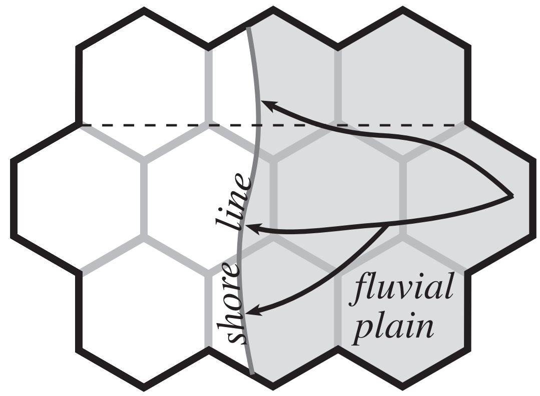


Fig.20. Plan view of the base-level experiment described in the text. The dashed line indicates the section shown in Fig. 21.

pulse of high-pressure water into the gravel in the elbow. A small volume of gravel is knocked out of the elbow and into an exhaust line. Each pulse produces about 120 µm of subsidence. The subsidence is spatially continuous because the cells are separated only at floor level. Firing a single cell, for instance, produces a smooth bowl-shaped subsidence pattern that extends over the six adjoining cells. The full-size basin is 11·4 m long, 5·8 m wide and has about 1·3 m of usable accommodation. Further details of the experimental system can be found at http://www.umn.edu/safl/X_strat.htm

Stratigraphic signature of base-level cycles

This experiment was carried out in a small subsiding-floor basin that is 1·3 m long and 1 m wide, with 10 subsidence cells (Fig. 20). The experiment was designed to study the effects of slow and rapid falls in base level on shoreline position and the associated stratigraphy. Water and a 50:50 mixture (by volume) of quartz and coal sand were fed from a single source point. Base level was controlled from the opposite side of the basin. Subsidence was induced in a bowl-shaped pattern with a maximum in the centre of the basin. The rates of subsidence and water and sediment discharge were maintained constant throughout the run. The sediment discharge was set so that the initial shoreline position would be about in the centre of the basin.

First, the basin was run at equilibrium for 10 hours to stabilize the shoreline. Then the first, slow cycle of base-level fall and rise was begun. The time history of base level is shown in Fig. 21. The slow cycle took a total of 30 hours and had an amplitude of 0·1 m. The rate of fall was never greater than the maximum subsidence rate. The delta-front sediments were thus never exposed. The cycle time was longer than the equilibrium time for the basin, as defined in Paola *et al.* (1992). After base level had returned to its equilibrium position, the system was left undisturbed for 6 hours to re-equilibrate. It is clear from the deposit that during this period a series of growth faults developed near the shoreline, and that they produced space that accommodated a significant fraction of the total sedimentation. There was no visible surficial evidence of the faults during the experiment, however. After the equilibration period a rapid cycle of base-level fall and rise was begun. The amplitude of this cycle was the same as in the slow case, but the cycle time was reduced to 2 hours. Base-level fall caused exposure of the delta front and development of a major channel incision. Exposure of the delta front also led to 10–20 mm of visible gravitational collapse of the front, which occurred along the pre-existing growth faults. During the rise, all the

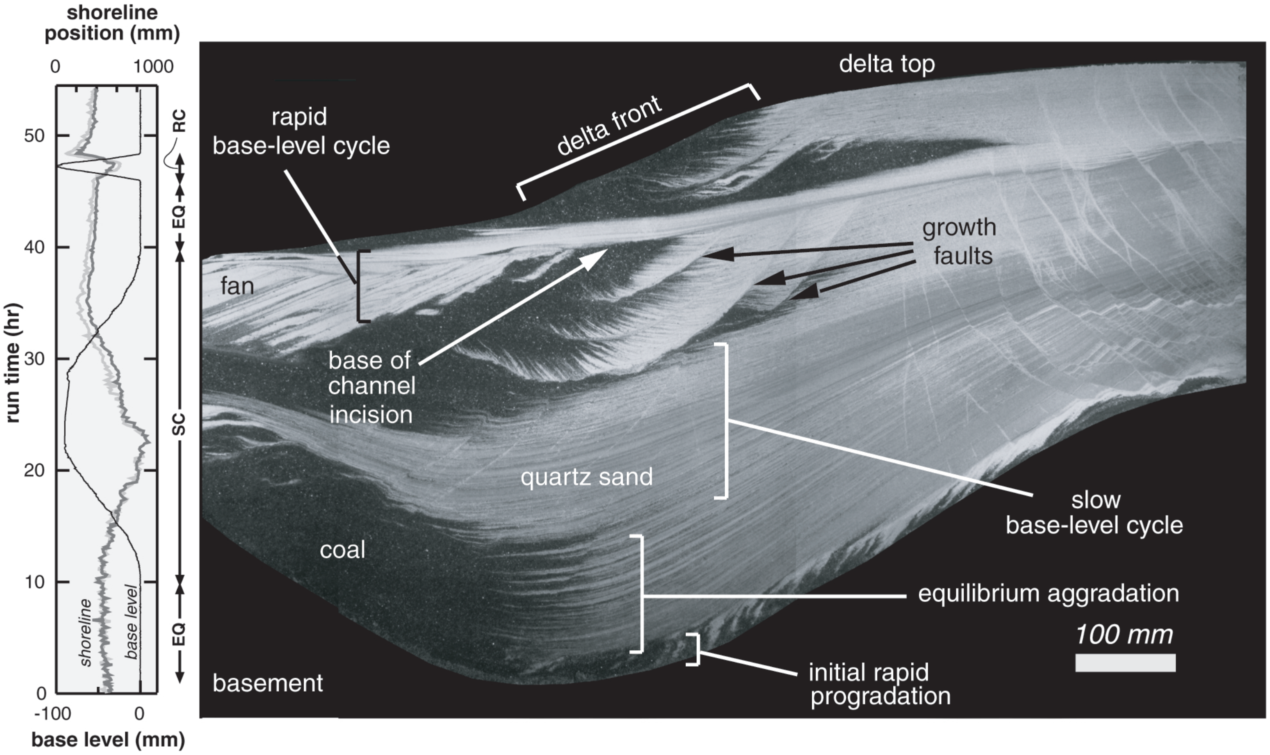


Fig. 21. Labelled photographic panel of the deposit produced in the prototype basin run. Transport was from right to left. Light tones are quartz sand, dark tones crushed coal. Base-level history is shown in the panel at left (black line), with the slow and rapid base-level cycles (SC and RC, respectively) and periods of equilibrium (EQ) marked. All other variables were held constant. Shoreline positions measured on longitudinal transects within and outside of the rapid-cycle incision are also shown at left (light and dark grey lines respectively).

sediment supplied to the system went to fill the incision, and as a result the shoreline substantially overshot its initial position at the end of the rise. The run ended with a regression as the shoreline migrated back toward the basin centre, during which time a new set of growth faults began to form.

The stratigraphic results of the experiment are shown in Fig. 21. The first, slow, base-level fall and rise produced a stratigraphically symmetric regression and transgression of the shoreline. This cycle produced only minor, localized incision, and the sand body produced was laterally extensive. On the other hand, the rapid base-level fall produced a narrow incised valley and an accompanying depositional lobe. Deposition across the entire system did not resume until the end of the cycle, when the incised valley was filled. Filling of the incised valley was accompanied by widening of the valley as sedimentation triggered active bank erosion. Outside of the valley, the subaerial part of the basin experienced the entire fall and rise cycle as a disconformity.

This experiment, despite its simplicity, produced a rich and complex stratigraphy. Other data from the experiment include a time-lapse video record of the surface pattern, and some 200 photomosaics of cross-sections parallel to the one in Fig. 21, spaced 2·5 mm apart. These have been assembled into a three-dimensional block diagram of the deposit, available for viewing at the web site given above. The scientific highlights of the run include: (1) the stratigraphic signature of slow base-level cycles is quite different from that for rapid cycles, as existing theory predicts; (2) the amplitude of response to the slow base-level cycle was larger than that for the rapid one, and neither showed strong phase-shifting of shoreline relative to base level; and (3) both base-level cycles, especially the rapid one, showed significant shoreline 'overshoots' at the end of the cycle, a feature not predicted by existing theory.

The use of stratigraphic experiments in model testing is illustrated in Fig. 22, which shows the measured base-level change and shoreline response in comparison with the theory of Swenson

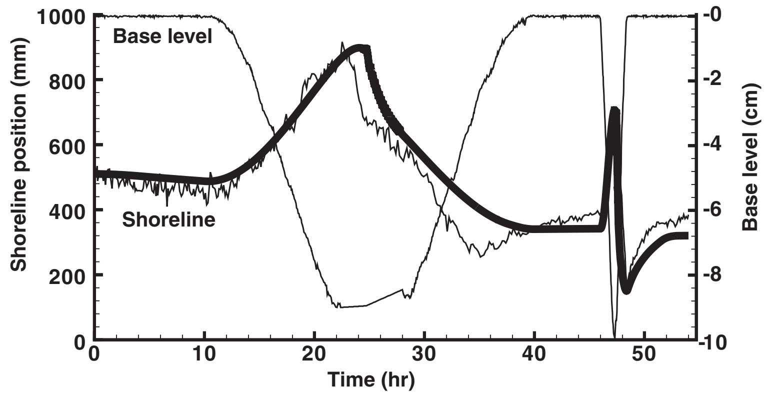


Fig. 22. Theoretical (heavy line) and experimental (light line) shoreline behaviour for two base-level cycles for the experiment described in the text. Theory from Swenson *et al.* (in press).

Table 1. Characteristic equations used in two-dimensional dynamic modelling and some of their basic properties. Symbols are defined in the text, except for ω and θ, which are growth coefficients with dimension time^{-1}

Characteristic equation	Environment(s)	Equilibrium condition	Equilibrium time (T_{eq})	Characteristic response	Stratigraphic signal
$\sigma + \frac{\partial \eta}{\partial t} = \omega(H - \eta)$	carbonate platform: shallow water (h< few m)	$\sigma = \omega(H - \eta)$	$\frac{1}{\omega}$	damped depth variation	stable palaeodepth
$\sigma + \frac{\partial \eta}{\partial t} = -\theta(H - \eta)$	carbonate platform: deep water (h> few m)	unstable	none	abrupt shallowing, drowning	shallowing-upward cycles; abrupt deepening
$\sigma + \frac{\partial \eta}{\partial t} = c\frac{\partial \eta}{\partial x}$	clastic shelf, erosional mountain belt	$\sigma = c\frac{\partial \eta}{\partial x}$	$\frac{L}{c}$	wave propagation	clinoforms
$\sigma + \frac{\partial \eta}{\partial t} = \frac{\partial}{\partial x}\left(\nu\frac{\partial \eta}{\partial x}\right)$	fluvial, clastic shelf, erosional mountain belt	$\sigma = \frac{\partial}{\partial x}\left(\nu\frac{\partial \eta}{\partial x}\right)$	$\frac{L^2}{\nu}$	spatially decaying	boundary-driven unconformities

et al. (in press). The only calibration involved estimating the effective transport coefficient from the supplied sediment flux and the measured fluvial slope. Thus the results shown constitute a true test, not a curve-fitting exercise.

One of the most useful aspects of experimental results such as those shown in Fig. 21 is that they provide a unique opportunity to apply one's interpretive techniques to a case for which the answer is known. We have done this exercise with a variety of interpreters: academic and industrial, experienced and inexperienced. The most interesting general result is not that almost no one gets it right – it is a very hard problem, after all – but that the error is nearly always on the side of overcomplicating the interpretation. We clearly have a lot to learn about how simple forcing can give rise to complex stratigraphy.

DISCUSSION

A synthesis of three basic model forms

If we consider the three main environments we have looked at in this review – carbonate shelf, clastic shelf and fluvial – we find that each has its own characteristic form of morphodynamic equation. These form a simple sequence if we express them in a common form, as given in Table 1. As discussed above, it is still not clear what is the best first-order morphodynamic equation for clastic shelves. I have chosen the wave equation simply because clinoforms are one of the things that make the shelf distinctive, and a wave equation is the most direct way to represent them. It is important to remember that the two clastic cases both make use of the general Exner (sediment-conservation) equation (Eq. 2). This equation applies as long as the deposited material is brought in (advected) from somewhere else. The carbonate equation does not conserve sediment because it refers to material produced within the water column. Physical transport of carbonates can be described with one of the clastic forms.

Each of these equations can be linearized to give a simple solution. The form of this solution can be interpreted as the mathematical expression of a characteristic depositional motif for each system, as illustrated in Fig. 23. The characteristic carbonate solution is exponential. If production depends positively on depth the solution tends to a stable equilibrium point where sedimentation (production) balances subsidence. If production depends negatively on depth an equilibrium point exists but is unstable, so the surface tends either to fill to sea-level or deepen rapidly if perturbed. The stratigraphic expressions of these two trends are shallowing-upward cycles and abrupt deepening, respectively. The characteristic clastic-shelf solution is an offshore-migrating wave, and the stratigraphic expression of this is clinoform deposition. The characteristic fluvial solution is an error function. The stratigraphic expression of this is response to forcing that tends to decay spatially away from the boundary where the forcing is applied.

Combining all three terms and extending them to two horizontal dimensions (x, y) we have the following generalized morphodynamic equation:

$$\sigma + \frac{\partial \eta}{\partial t} = p(H - \eta) + \left[c_x \frac{\partial \eta}{\partial x} + c_y \frac{\partial \eta}{\partial y}\right] + \left[\frac{\partial}{\partial x}\left(\frac{\nu \partial \eta}{\partial x}\right) + \frac{\partial}{\partial y}\left(\frac{\nu \partial \eta}{\partial y}\right)\right] \tag{9}$$

where p is a production function with units of (volume of sediment produced)/(bed area)/(time), and H is water level relative to some fixed datum. This is an advection–diffusion equation with one source term (the production rate p) and one sink term (the subsidence rate σ). So far p has been used only for carbonate production, but it could as well represent production of evaporites, biogenic silica or (if modelling a particular grain size) production or destruction of particles by aggregation–disaggregation processes. In the other two terms H does not appear explicitly, although the coefficients c and ν depend on at least the sign of $H - \eta$. They may depend independently on x and y as well. The second and third terms are both mass-conserving, and come from the Exner Eq. (2). The second term is a first-order wave equation (Eq. 7), also called an 'advective term', and describes the effect of dependence of sediment flux q_s on elevation η. The wave speed c has two vector components corresponding to the two horizonatal components of the sediment flux: $c_x = -(1/C_0)\mathrm{d}q_{sx}/\mathrm{d}\eta$ and $c_y = -(1/C_0)\mathrm{d}q_{sy}/\mathrm{d}\eta$. The third term is diffusive (Eq. 1), although if the diffusion coefficient ν depends on position, differentiation of this term produces additional wave-like terms, as discussed earlier. Overall, the ratio of the wave (second) term to the diffusive (third) term scales as $\mathbf{c}L/\nu$ where $\mathbf{c}$ is a characteristic value of the wave speed and L is a length scale. This number, which has the form of a Reynolds or Péclet number, is a measure of the importance of wave-like (advective) behaviour relative to diffusion. In the stratigraphic context it can be thought of as a ratio of clinoform to spatially decaying response styles.

The future of basin modelling

In the preceding sections, I have suggested some specific problems that deserve attention in the coming years. More generally, an entire area that has apparently been overlooked so far is large-scale quantitative modelling of aeolian systems. This could be done building on the conceptual framework provided by Kocurek & Havholm (1994). Other natural next steps involve better integration of various subsystems now being looked at separately. One of the grand challenges in this regard will be the continued development of fully coupled dynamic models for entire sedimentary systems, such as the continental margin. This will require much closer integration of carbonate and clastic modelling, which are often still pursued independently. There is also still plenty of work to be done on coupling erosional and depositional systems following the initial steps discussed earlier. Another integration problem is coupling models for basinal fluid flow and reaction (Harrison, 1990; Person *et al.*, 1996; Lee, 1997) to dynamic stratigraphic models. The Stanford SEDSIM group has, it is claimed, already accomplished some of this. Unfortunately, the development of SEDSIM has been closed, allowing for neither evaluation, testing nor use by other researchers. A much better approach is Syvitski *et al.*'s (1997) call for development of a 'community' stratigraphic model along the lines of community climate models. This approach is surely the way to develop the kind of large-scale, highly integrated models that will be needed to simulate sedimentary basins in detail.

As interesting as it will be to solve the many remaining technical problems in basin modelling, I hope we can do better than this. I hope we can also think of new and more creative ways to approach modelling itself. Here are a few thoughts on this.

Unlimited computing power

How does the advent of nearly unlimited computing power change our way of thinking about modelling in the earth sciences, or for that matter in science in general? Until the late twentieth century, simplification of problems was a necessary and universally accepted first step in modelling physical processes. One could argue that the most essential element of scientific skill was the ability to simplify insightfully, to extract the real essence of a problem. As computers become more powerful, simplification becomes less necessary. The obvious (and simple) response is that if we do not have to simplify, we should not do it. We all know that nature is complex, and so, according to this line of reasoning, our models of nature must be similarly complex. A model that includes more processes and/or does so in more detail is always better than a simpler model. The ideal model would, by extrapolation, include everything. I term this viewpoint 'literalism': the idea that theoretical models should always aim for the most exact, literal and detailed representation of natural systems that current knowledge and computing power allow. It is surprising how common this view of things is, so perhaps it is worth noting that the only entirely acceptable model in the literalist world would be an exact replica of the Earth itself (in an identical Solar System, of course).

In areas of science in which quantitative theory is part of the disciplinary tradition, the behaviour

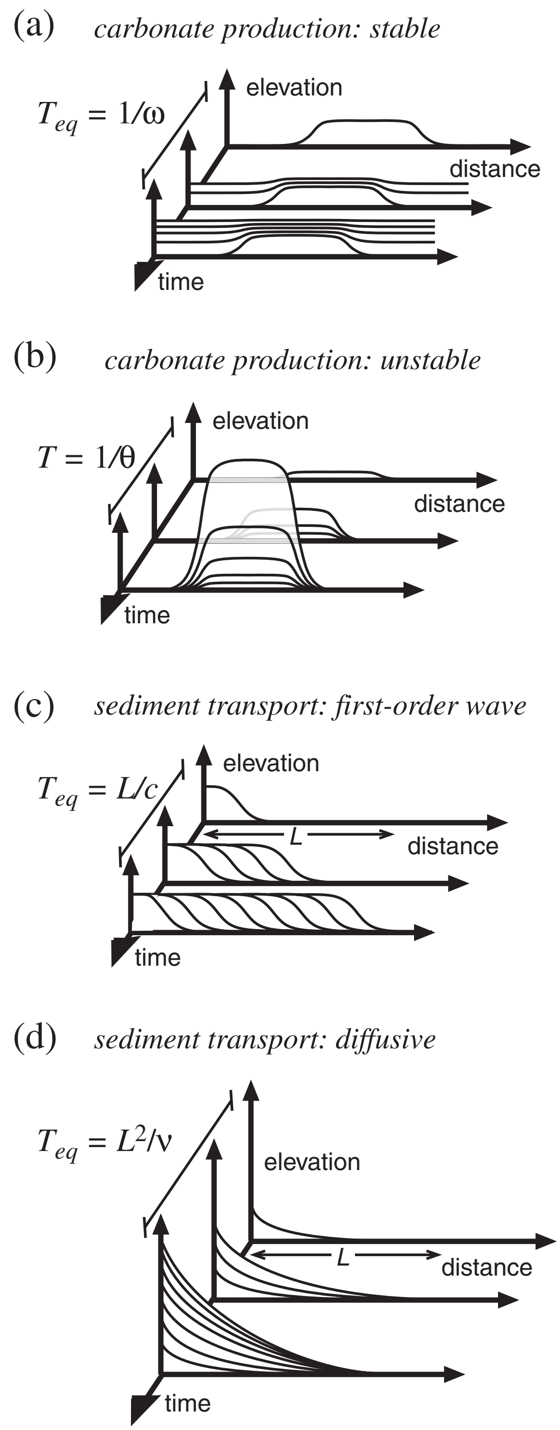


Fig. 23. Stratigraphic expressions of linearized solutions to the main equation forms used so far to model basin filling (Table 1). (a) Carbonate, shallow-water form (increasing production with depth). Stable equilibrium; the system responds to perturbation by returning to equilibrium on a timescale of the order ω^{-1}. (b) Carbonate, deep-water form (decreasing production with depth). Unstable equilibrium; the system responds to perturbation by exponential shallowing or deepening on a timescale of the order θ^{-1}. (c) Clastic shelf, first-order wave equation. Stable equilibrium; the system responds to perturbation by wave propagation on a timescale of the order L/c. (d) Fluvial system or clastic shelf, second-order linear diffusion equation. Stable equilibrium; the system responds to perturbation by a damped response on a timescale of the order L^2/ν. For perturbations with periods T significantly shorter than L^2/ν the response is damped over a distance of the order of $(T\nu)^{1/2}$.

and limitations of simplified models are well known. Large-scale computational models are developed not simply because they are feasible, or because of a generic belief that natural systems are complicated, but because it is known that they are required to solve a particular problem. There is no question that comprehensive stratigraphic models will be large, complex and very useful. But in sedimentary geology, it has happened that the conceptual breakthroughs that made analytical basin models possible have coincided with the manufacturing breakthroughs that made large-scale computational power available at low cost. Obviously, this is a good thing. However, it also means that it is now easy to create models that we ourselves do not understand, because we have skipped the initial 'simplistic' steps, or because we know more about programming than we do about analysis. There is a great temptation to program to the limits of our computers, rather than to the limits of our comprehension. Perhaps this is because, while it is easy to justify putting things into a model, it takes real understanding to justify leaving things out. It would be a shame if, in the pursuit of bigger and more comprehensive models, we were to lose sight of the equally valuable pursuit of the elegant, the insightful and the essential.

Model runs as experiments

Available computing power may be increasing exponentially, but the grasp of the human mind is not. As models become less comprehensible to their creators, running them becomes less like the final step in an analysis and more like an exploratory physical experiment. In what sense is a complicated model run analogous to an actual physical experiment? The criteria that a model must meet to be thought of in this way are: (1) the *governing equations* for the process must be completely known, and (2) the *representation* of those equations in the numerical model must have been thoroughly tested (e.g. by comparison with known solutions of the relevant equations), so that it is certain that the model output really represents the solution of the equations. Only if both conditions are satisfied is it safe to consider a computer simulation as equivalent to a physical experiment. At present, large-scale models of sedimentary basins do not pass criterion (1), because we do not know the complete governing equations for any sedimentary system. As discussed above, scaled-down physical experiments can provide a useful bridge between untested theories and hard-to-constrain field cases. Experimental testing is a crucial step in building confidence that our models correctly capture dynamics that is either scale-independent or scale-dependent in ways that we can account for. Nonetheless, experimental testing clearly is not enough. Ultimately, all geological models must be tested against natural systems. How can this be done?

Model testing

Time is to historical geology what distance is to cosmology. Models in both fields cannot be tested by direct experimentation on the objects under study because they are removed from us by impossibly large dimensions. Passage through both deep space and deep time progressively destroys information, but what reaches us from space lends itself far more readily to interpretation via physics than what time has left us. Ideally, we would look back into the past using only physical laws that we are certain are invariant. The fact that this approach is so hard to apply has to do both with the kind of information we have in geology and with the irreversible evolution of the Earth; it is why geology is not a branch of physics. Nonetheless, the more closely our models come directly from basic physical laws, the more confident we can be about applying them to systems where they cannot be tested directly.

After we have done our best to base our models on known physical laws and to test them experimentally, what remains must take the form of an integrated approach to narrowing the constraints on field cases as much as possible. We must make use of every available type of information, descriptive as well as quantitative. A crucial limitation in this regard is the lack of highly integrated basin-scale data sets. What could we learn if the best techniques we have, from all the relevant subdisciplines, were brought to bear in a concerted way on the best exposed and constrained basins in the world? The culture of geology suffers in this regard from the curse of easy access: we divide the world into little areas, and each of us works on 'our rocks', largely because we can. Integrated programmes, in which a number of researchers with different skills and viewpoints work together for something larger than any of them could achieve on their own, are much more common in fields like oceanography, where often there is no choice. One of the great challenges of the next few decades will be to develop in sedimentary geology a complementary culture of large-scale integrated field studies, without destroying the diversity and creative potential of our existing 'small-science' culture. This will involve learning from colleagues in fields like oceanography and geophysics, and co-operation with industrial colleagues as well.

Model design should also take field testing into account. So far, we have largely concentrated on making plausible-looking simulations and explaining general patterns. 'Testable design' would involve designing models to predict relations between aspects of preserved deposits that can be independently measured (e.g. Paola, 1990; Robinson & Slingerland, 1998), rather than just reproducing basin fills. In the fluvial realm, we have now been using diffusional models for years. As mentioned earlier, derivation of the equation from first principles allows us to predict downstream changes in variables like channel width and depth, but so far this opportunity has hardly been exploited. Little has been done to see how well one might constrain the transport (diffusion) coefficient from the stratigraphic record, based on careful analysis of grain size, sedimentary structures and other transport indicators in the deposits. Palaeohydraulic reconstructions of ancient rivers had a vogue in the 1970s and 1980s (e.g. Cotter, 1971; Baker, 1974; Ethridge & Schumm, 1978; Kochel & Baker, 1982; Gardner, 1983; Steer & Abbott, 1984) but have so far not been integrated into the fabric of sedimentary geology. Part of the problem has been a lack of context for the numbers: of what use is it to say that the discharge of a Cambrian river was $X\,\mathrm{m^3\,s^{-1}}$ and leave it at that? Now, however, quantitative basin models have given us a framework in which such numbers might be pieced together to help constrain transport coefficients and hence predictions of large-scale stratal geometry.

In this regard, an area that has received insufficient attention is careful analysis of what can actually be determined quantitatively from the stratigraphic record. Lessenger & Cross (1996) have argued that inversion of stratigraphy is possible, while Burton *et al.* (1987) and Burgess & Allen (1996) have stressed the nonuniqueness of the stratigraphic record. Heller *et al.* (1993) have outlined a method for constraining the space of possible forcing functions for a given stratigraphic pattern. It should be possible to compute the loss of information from ancient systems, and use this to design more testable models. In this regard, the gap between Quaternary geology and general sedimentology and stratigraphy still represents an underexploited opportunity. The review by Blum and Törnqvist (2000) shows the value of Quaternary studies in evaluating the stratigraphic effects of eustatic and climate change. Quaternary studies could be equally valuable for quantitative modelling; for instance, a systematic approach to model testing might involve starting with the shortest intervals and going back in time progressively. There also may be an analogy between the abstraction of dynamics from one level to the next in a multiple-timescale model and the loss of information as one goes deeper into the stratigraphic record.

Learning from climatology

There are some useful parallels between modelling climate and modelling sedimentary basins. The atmosphere and sedimentary basins are both complex, nonlinear systems; both are primarily physical but involve significant biological effects; and both involve dynamics over a wide range of timescales. In this analogy, sedimentary processes on human timescales, typically the domain of engineering and geomorphology, constitute the 'weather'. This points to a crucial aspect in which the analogy breaks down: weather modelling uses an agreed-upon set of basic equations, and weather models are tested many times every day. For all the major sedimentary environments, there are still fundamental questions about how to set up the governing flow and transport equations, and the database of case studies is much more limited. Another important way in which the analogy fails is that while we have only one atmosphere, we have many different basins. The general problem of modelling sedimentary basins is comparable with modelling not just the Earth's atmosphere but those of all other planets as well.

These differences make climate modelling easier than basin modelling. Because of this, and because it is topical, climate modelling is much more advanced than basin modelling. Thus we may be able to learn something from it. In particular, climate modellers have given considerable thought to dealing with multiple time and space scales. When we run a discretized model of a basin, we choose a time step $\mathrm{d}t$, and a grid size $\mathrm{d}x$. Processes with timescales less than $\mathrm{d}t$ and length scales less than $\mathrm{d}x$ are, by definition, below the resolution of the model. How do we extract the important dynamics of those processes so that their effects are correctly represented in our basin model?

The main way this is handled in climate modelling is through so-called ‘parameterization’, in which events or processes below the model resolution are represented as functions of variables that are resolved in the model (e.g. average temperature at a point). The most familiar example of this kind of parameterization is the representation of turbulent momentum flux via an eddy viscosity. In clastic basin modelling, the Exner sediment-continuity equation (Eq. 2) is the fundamental equation that is discretized and solved. Synthetic relations between sediment flux and, for instance, mean topographic slope (Eq. 3) are examples of parameterization. The transport coefficient ν is the parameter. In climate modelling, a good deal of effort has gone into calculating the parameter values by suitably averaging the underlying equations. In basin analysis, this process is only just beginning (Paola *et al.*, 1992, 1999; Niedoroda *et al.*, 1995; Paola, 1996).

The opposite problem, of resolving subgrid-scale local effects, has been approached via so-called ‘nested’ models (Fig. 24; Giorgi & Mearns, 1991). The idea is to have a series of grids, of increasingly fine resolution, one inside the other like a Russian doll. In the simplest kind of nested model, information flows downward, i.e. the solution at the large scale provides a set of boundary conditions for a more detailed model of processes within a particular grid cell. In the basin context, an overall diffusional model could, for instance, provide the boundary conditions for a detailed model of channel processes within a grid cell. However, information can also propagate upward in so-called ‘adaptive multigrid’ methods, in which the results of the fine-scale simulation can change parameter values at the larger scale (Barros & Lettenmaier, 1993).

Another useful idea that has been suggested in climate modelling is that the behaviour of a complex system over long timescales may be simpler and more predictable than its behaviour over short timescales. Mahlman (1998) has stressed the distinction between weather forecasting and climate modelling. Weather forecasting is a classical initial-value problem in a chaotic system and therefore cannot in principle be solved very far into the future. Modelling the effect of a secular change (e.g. a rise in CO_2 concentration) on climate is a boundary-value problem that in principle has a definite solution. (Here the climate represents a statistical steady state of the atmosphere.) Mahlman has proposed a particularly apt metaphor that also works for basin analysis: think of the atmosphere (basin) as a pinball machine. Forecasting the weather (evolution, for instance, of a single river channel belt) is like trying to forecast the precise path of a single ball, which is practically impossible. Modelling climate change (long-term basin response) is like trying to model the mean playing time over many balls, as a function of the tilt angle of the machine. This long-term problem is solvable in a way that the unpredictable short-term problem is not.

Emergent behaviour, self-organization and chaos

From an intellectual standpoint, the most far-reaching result to come out of climate research is the theory of complex systems, with its core result that even determinisitc systems can be fundamentally unpredictable (Lorenz, 1963). Unpredictable behaviour in deterministic

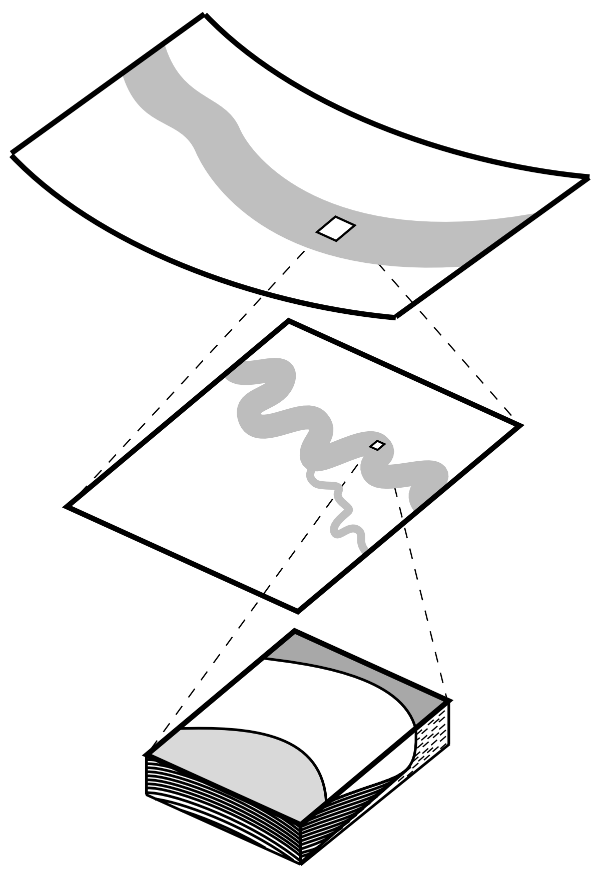


Fig. 24. How ‘nested’ numerical models might be used in stratigraphy. In the simplest scheme, the solution at each higher level provides boundary conditions for a more detailed solution at the lower level. In more complicated schemes, the lower-level solution can also feed information back up to higher levels.

systems was originally termed 'chaos', but this term has now become somewhat ambiguous. We will retain it here with the understanding that it carries no implications beyond the original definition. Although sedimentary basins are certainly complex systems, it is noteworthy how little impact complex-systems theory has had on modelling them. Slingerland (1990) and Nicolis & Nicolis (1991) have provided brief descriptions of complex-systems dynamics and chaos, but only a very broad outline of how these ideas might apply to sedimentary systems. At this point, most large-scale models for stratigraphy are of conventional (prechaos) form, evidently reflecting a belief on the part of their creators that the longest-period stratigraphic signals directly reflect external forcing. The Stanford SEDSIM model discussed earlier shows signs of unpredictability (Tetzlaff, 1990), but this has evidently not been analysed systematically. Even stochastic effects, certainly present in stratigraphic systems and themselves a form of chaos, have been incorporated explicitly in basin models only by Tipper (1991, 1992) and Pelletier & Turcotte (1997). If we step back and look at sedimentary processes as a whole, the general pattern that emerges is one of alternation between unpredictability and predictability as a function of scale. At the finest scale of interest, turbulence is the 'type example' of deterministic chaos. There are also signs of chaos at the level of bedform interactions (Rubin, 1992; Werner, 1995; Werner & Kocurek, 1997). Moving up in scale, sedimentary-system response to single events, like floods or storms, has been modelled using conventional predictive methods. This is reflected in engineering models of river or coastline behaviour over human timescales. In the fluvial case, this window of predictability occupies what might be called the 'in-channel' scale, i.e. the range of timescales on which the dynamics of a channel is determined primarily by local processes within that channel. In fluvial systems a new level of unpredictability arises for timescales on which interaction with other channels or the floodplain must be taken into account. Braiding defines the shortest of these timescales, nodal avulsion the longest. We might call this the 'multichannel' range of timescales. The unpredictability is reflected in the stochastic aspects of modelling avulsion (Leeder, 1978; Mackey & Bridge, 1995), and models of braiding that exhibit chaos (Murray & Paola, 1994; Murray & Paola, 1997) and power-law scaling (Sapozhnikov & Foufoula-Georgiou, 1996, 1997). Avulsion on deltas and submarine fans is presumably comparable, but beyond that it is not clear whether analogues of this mesoscale chaotic interval exist in other environments. In any case, at the largest scales, the view implicit in the modelling we have reviewed is that sedimentary systems are once again predictable.

Even within the apparently chaotic scale intervals, there seems to be little indication that the formal mathematical theory of chaos, with its emphasis on rigorous analysis of low-dimensional systems, is going to be of major relevance for basin analysis (for a possible counterexample see Gaffin & Maasch, 1991). But unpredictability is only one aspect of nonlinearity; other types of nonlinear behaviour are everywhere in basin modelling. Even apparently simple geometric models need not be linear, and coupled models involving linear components become nonlinear if any of the coupling conditions are nonlinear. Stratigraphic models that focus specifically on nonlinear effects include Gaffin & Maasch (1991), Humphrey & Heller (1995) and Swenson *et al.* (in press).

The idea of emergent behaviour is one of the more interesting and relevant developments in the theory of complex dynamical systems. 'Emergent' describes behaviours that seem to emerge from the interactions of the parts of a nonlinear system. Emergent behaviour occurs in nonlinear systems with many interacting parts, and cannot be derived directly from the equations governing the parts. A common type of emergent behaviour is formation of organized patterns through internal interactions; this is called 'self-organization'. Basins are obviously complex systems with many interacting parts, so it is reasonable to expect emergent behaviours in them. Knowing that emergence is possible and perhaps inevitable in this type of system raises the possibility that modelling them from the bottom up (i.e. by summing many small-scale or short-term events) might be inefficient, misleading, or as Werner (1999) has argued, fundamentally inconsistent. This is the basis for the 'synthesist' approach to basin modelling that was described earlier.

At the fine scale, self-organization models have been applied to bedforms and other geomorphic patterns (Werner & Fink, 1993; Werner & Hallet, 1993; Werner, 1995; Werner & Kocurek, 1997). All of these are highly organized patterns that clearly do develop spontaneously without external forcing. Likewise, in the model of stream braiding proposed by Murray & Paola (1994, 1997), the braid pattern emerges from a highly simplified model of the dynamics. In this realm there is still vigorous debate about whether the pattern formation requires an approach explicitly based on emergence, or can be better modelled with conventional 'reductionist' approaches.

There are aspects of emergence and self-organization in sedimentary systems that have not been identified as such. Autocyclicity (Beerbower, 1964) can be seen as an example of mesoscale self-organization in channellized systems. 'Autocyclic' has come to be used for internally generated stratigraphic cycles on length scales large enough that one might plausibly think they were externally forced. (Although it was used for good reason, the 'cycle' in 'autocycle' should not be taken too literally. Self-generated fluctuations need not be periodic, and may have very broad spectra. They are, however, a persistent, intrinsic feature of the system.) Typical autocyclic mechanisms include meander migration and cut-off, and channel-belt or lobe switching. In these cases, as in bedforms and bars, the transport system organizes itself into a highly structured spatial pattern that generates strong variations in sediment flux even with steady forcing. The spatial pattern leaves a corresponding signature in the stratigraphic record. If the essence of autocyclicity is internally generated variability in sedimentation owing to spatial pattern formation, then it encompasses a wide range of scales and forms, from bedforms through bars and small-scale channel switching, and up to large-scale lobe switching and avulsion. These can all be seen as aspects of self-organization in sedimentary systems.

The traditional question of how to tell autocyclic from allocyclic responses could be rephrased as that of distinguishing self-organized (internally generated) sedimentary patterns from externally forced ones. The better we understand the internally generated signals, the more accurately we can constrain the external forcing. Traditional wisdom says that autocycles are restricted in both spatial and temporal scale, and this is the main basis for identifying them. So at this point, it seems that self-organization, in channellized systems at least, is restricted to scales ranging from those of bedforms to the mesoscale of channel belts, lobes and avulsion. (The only hint so far of self-organization beyond the level of bedforms in the shallow-submarine or aeolian realms is the autocyclic drowning mechanism of Ginsburg (1971).) There is an element of circularity in our reasoning. What kinds of large-scale self-organized patterns might there be in sedimentary systems that we have not recognized because we have always assigned them external causes?

Another interesting aspect of self-organization that has received little attention so far is the interplay between self-organized and externally forced behaviours. One can see, however, the possibility that external forcing within the range of autocyclic timescales might either be completely masked in the autocyclic 'noise' or produce a very unintuitive response. Indeed, some aspects of what Schumm (1977) calls 'complex response' could be viewed as the superposition of self-generated and externally forced dynamics. Though 'complex response' is a bit vague, the basic concept of exploring system response to forcing on timescales comparable with those of self-generated variability is potentially a very fruitful area of research. It would bear revisiting with modern dynamical-systems concepts in mind.

Where are we now?

Is sedimentary geology a quantitative, analytical science, or a descriptive one? Is it model-driven or observation-driven? Inductive or deductive? Should we look at rocks or study mathematics? Of course, the answer is 'all of the above'. It is not clear whether basins are chaotic, but the state of modelling them certainly is. Within just the last 10 years we have seen publication of everything from the simplest of one-dimensional geometric models to three-dimensional, computationally intensive dynamical models, all aimed at explaining stratigraphy. This chaos is a good thing, and we should not try to escape from it too soon. Overall, we have learned two things:

1 It is not hard to invent plausible stratigraphic models.

2 It is very hard to test them.

This is not meant to be dismissive. Given how much we still do not understand about morphodynamics, the fact that we have been able to produce useful models at all is a sign that at least some of what we are interested in stratigraphically is not all that sensitive to the details of basin-filling processes. One of the common threads that has emerged from basin modelling using a variety of approaches is that even relatively simple forcing can produce wonderfully complex effects within basin systems. This should not be discouraging; rather, it is a motivation to redouble our efforts to be economical in invoking causes for observed stratigraphic patterns. We do not need a separate explanation for every observed stratigraphic feature. Rather what we need is a general framework for defining the space of possible forcings and parameters that could produce a given stratigraphic pattern, along the lines suggested by Heller *et al.* (1993) and Lessenger & Cross (1996). The model results reproduced in this review form an abbreviated catalogue of some of the stratigraphic patterns that modellers have seen. They can be used as a starting point in the search for the right combination of parameters to explain a given real sequence.

Another general result is that basin systems have intrinsic timescales, and that the nature of stratigraphic response to external forcing can depend strongly on the timescale of the forcing relative to these timescales. Some of the timescales that have been identified for depositional processes are given in Table 1. We have now formulated, and can estimate magnitudes for, timescales associated with erosion, deposition and isostatic relaxation. More generally, we also now have a number of theoretical cases in which systems with different timescales respond very differently to the same forcing (Angevine, 1989; Reynolds *et al.*, 1991; Paola *et al.*, 1992; Steckler *et al.*, 1993; Rivenæs, 1997). This is a possibility that must be borne in mind when trying to separate global from local effects by looking for similarity of response over a wide area. It would be useful to unify all these results in a common framework, based on fundamental dimensionless parameters (e.g. Swenson *et al.*, in press). The dependence of basin response on local characteristics should also be testable, for instance by well-constrained comparison of the response of basins with different timescales (e.g. different lengths) to the same forcing.

Overall, however, the main point that emerges from the modelling done to date is that even simple models readily surpass our ability to test them. As discussed earlier, the strategy will have to be to constrain the problem from as many directions as possible, using a combination of tactics: incorporation of the best fundamental understanding of processes; continuing the search for simple, testable behaviours within complex models; playing off different forms of stratigraphic information against one another; emphasizing testability as we develop models; critically comparing results from models using different approaches; pushing experimental tests to their limits; using our models to analyse both the uniqueness of model results and the limits placed on us by the stratigraphic record; and above all gathering comprehensive descriptive and quantitative data sets from the best field sites in the world.

Educating analytical sedimentologists

So far we have discussed what has been done in quantitative modelling, and what might be done in the future. In my view, however, the single most important problem facing us now is human, rather than scientific. We have to change the culture of the discipline. Geology has often served as a haven for students seeking to do science without mathematics, and within geology, sedimentology and stratigraphy have formed one of the more sheltered parts of the harbour. It will do little good to develop better models if most of our students see them as black boxes, or worse, are either terrified or awed by them. It is not that everyone in the field must be able to develop models themselves. On the contrary, it seems clear that inasmuch as the weak link in the chain is already not developing models but rather using and testing them, there will continue to be a much greater need for field researchers than for modellers in sedimentary geology. Rather, what is needed is researchers who are skilled in the field but at the same time understand what quantitative modelling is about: why and how people make approximations, why approaches to modelling can and must differ, and, above all, what the mathematics in the models means physically. Just as there is no substitute for experience in learning to work in the field, there is no substitute for experience in developing physical insight. And there is no shortcut: we need researchers who are good at both traditional, descriptive geology and quantitative, analytical geology. For the 'modal' sedimentary-geology student, it is not sophisticated computational skills or training in advanced calculus that is lacking, but rather the routine application of basic quantitative reasoning. This means things like estimating scales and rates for key processes, knowing the magnitudes of basic physical properties, and being able to estimate the relative importance of various processes in a particular setting. Understanding scales, rates and relative magnitudes is to quantitative science what recognizing quartz and feldspar is to field geology. Neither requires years of sophisticated training, but both require repetition until they become habitual.

In the 1960s, research on the origin, mechanics and sedimentary signature of bedforms led to a new appreciation of the value of fluid and sediment mechanics in sedimentary geology. Landmark publications in this regard were the works of Allen (1970) and Middleton & Southard (1984). The modelling of sedimentary basins and strata has provided another avenue of application of physical principles to sedimentary geology. To what extent has all of this worked its way into the basic fabric of our discipline?

Some 30 years after the initial 'physics scare' associated with bedforms and sedimentary structures, a set of basic principles from fluid and sediment mechanics now appears routinely in introductory sedimentology textbooks. Popular items include settling velocity and Stokes's Law, the Reynolds and Froude numbers, and the basic force balance for steady, uniform channel flow. This material is typically presented somewhere near the beginning of the book and then is largely ignored (though progress has been made in the recent texts of Allen (1997) and Leeder (1999)). There remains a striking contrast between the role of fluid and sediment physics in sedimentary geology and that of thermodynamics in igneous and metamorphic geology. In 'ig-met' texts the underlying thermodynamic principles are introduced and then applied repeatedly. Whereas in hard-rock petrology, thermodynamics permeates the discipline, in sedimentary geology, sediment mechanics still seems a little like taking vitamins: it is surely good for you, but most people cannot say exactly why. There are several reasons for this. In current practice, process-based interpretation is often applied in a piecemeal, descriptive way, to no apparent end beyond providing the interpreter with one more adjective. In addition, the quantitative material that is traditionally taught is often not the most important. For instance, a real appreciation of the implications of the sediment-continuity equation as the governing relation for physical sedimentation is far more useful than the details of sediment-transport formulae or even the definition of the Reynolds number. Finally, and most important, those of us in the 'quantitative school' simply have not provided a comprehensive interpretive framework that is sufficiently compelling to convince the larger community to learn a new way of thinking. At the moment it appears that the crossover of quantitative thinking into the mainstream may happen first in the deep-marine environment, where the fluid-mechanical approach to turbidites and debris flows is in a period of rapid acceleration (Kneller & Buckee, 2000). The physics of sediment-gravity flows is one level more subtle than that of rivers – perhaps just enough so that we cannot even understand deep-marine deposits qualitatively without the help of quantitative analysis. Wherever the practical utility of the quantitative approach is first demonstrated conclusively, it will spread from there to the rest of sedimentary geology. At the basin scale, the models discussed here have opened a door. Following through represents one of the great challenges for quantitative sedimentary geology in the next century.

Deciphering the language of sediments is one of the great enterprises of the Earth sciences. We try to base our interpretations as closely as possible on basic principles because only these do not change with time. We try to apply these principles quantitatively because quantitative models yield sharper predictions and can be more readily tested than qualitative ones. It is still sometimes said that the complexity of sedimentary systems places them beyond the reach of analysis, in a domain where the only recourse is endless description and cataloguing. Whether these objections arise from legitimate scepticism or from a simple distaste for mathematics, they will ultimately be overcome not by philosophical arguments but by the development of quantitative models that are simply too useful to resist. Nonetheless, the beginning of a new millennium is a time for reflection, as the papers in this special issue attest. Here in closing are some general thoughts:

1 At this point it is not known whether the inverse problems we are trying to solve in sedimentary geology have unique solutions. Even if they are nonunique, this does not mean they are incomprehensible. For problems without a unique solution, the role of analysis is to find the space of possible solutions. It may well be qualitative constraints that determine which theoretical possibility is correct.

2 The multivariate character of the stratigraphic record means that there are many sources of information to play off against one another.

3 There is no law of nature that says that 'complexity + complexity = greater complexity'. The sum of complicated processes may be simple, especially if the processes are uncorrelated. The possibility of simple behaviour in complex systems should never be dismissed until it has been shown to be absent by observation.

4 Quantitative modelling is a complement to, and an extension of, traditional conceptual modelling in sedimentary geology. A quantitative model can refine and add predictive power to a conceptual model, or show that it is physically impossible. But a correct quantitative model cannot be developed from an incorrect conceptual model.

5 Physical sedimentary geology deals with processes that obey known laws. There is a great deal of work to be done to find out how to apply these laws over the immense span of geological time. But there is no sign anywhere of processes, events or conditions that are beyond the reach of human physical insight.

ACKNOWLEDGEMENTS

It is a pleasure to thank Phil Allen, Phil Ashworth, Jim Best and Dan Bosence for thoughtful, comprehensive reviews that considerably improved this paper. I also thank Tom Hickson, Lincoln Pratson, John Swenson, Paul Heller, Gary Parker and Rudy Slingerland for helpful informal reviews, and Marcia McNutt, Gary Kocurek, Neil Lundberg, Shun Karato, Brad Murray and Chris Kendall for useful suggestions. James Syvitski and Eric Hutton produced Fig. 16 and Lincoln Pratson helped with Fig. 21. Finally, I am deeply grateful to Paul Heller, Gary Parker, David Mohrig and Giovanni Seminara for innumerable discussions that have clarified my thinking on many of the topics discussed in this paper. This work has been supported by US National Science Foundation Grants EAR 95-12472 and EAR 97-09217, and by the members of the St. Anthony Falls Experimental Stratigraphy consortium: ARCO Exploration and Production Technology Company, Conoco Inc., Exxon Production Research Company, Japan National Oil Corporation, Mobil Technology Company and Texaco Corporation.

REFERENCES

Adachi, S. and **Nakatoh, T.** (1969) Changes of top-set-bed in a silted reservoir. In: *13th Congress, International Association for Hydraulic Research*, **5,** 269–271.

Aigner, T., Brandenburg, A., van Vliet, A., Doyle, M., Lawrence, D. and **Westrich, J.** (1990) Stratigraphic modelling of epicontinental basins: two applications. *Sedim. Geol.*, **69,** 167–190.

Aigner, T., Doyle, M., Lawrence, D., Epting, M. and **van Vliet, A.** (1989) Quantitative modeling of carbonate platforms: some examples. In: *Controls on Carbonate Platform and Basin Development* (Ed. by P. D.Crevello, *et al.*), *Spec. Publ. Soc. Econ. Paleont. Miner.*, **44,** 323–338.

Alexander, J. and **Leeder, M.R.** (1987) Active tectonic control on alluvial architecture. In: *Recent Developments in Fluvial Sedimentology* (Ed. by F. G. Ethridge, R. M. Flores and M. D. Harvey), *Spec. Publ. Soc. Econ. Paleont. Miner.*, **39,** 243–252.

Allen, J.R.L. (1965) A review of the origin and characteristics of recent alluvial sediments. *Sedimentology*, **5,** 89–191.

Allen, J.R.L. (1970). *Physical Processes of Sedimentation*. George Allen and Unwin Ltd, London.

Allen, J.R.L. (1974) Studies in fluviatile sedimentation: implications of pedogenic carbonate units, Lower Old Red Sandstone, Anglo-Welsh outcrop. *Geol. J.*, **9,** 181–208.

Allen, J.R.L. (1978) Studies in fluviatile sedimentation: an exploratory quantitative model for the architecture of avulsion-controlled alluvial suites. *Sedim. Geol.*, **21,** 129–147.

Allen, J.R.L. (1979) Studies in fluviatile sedimentation: an elementary geometrical model for the connectedness of avulsion-related channel sand bodies. *Sedim. Geol.*, **24,** 253–267.

Allen, P.A. (1997) *Earth Surface Processes*. Blackwell Science, Oxford.

Allen, P.A. and **Allen, J.R.** (1990) *Basin Analysis: Principles and Applications*. Blackwell Scientific Publications, Oxford.

Angevine, C.L. (1989) Relationship of eustatic oscillations to regressions and transgressions on passive continental margins. In: *Origin and Evolution of Sedimentary Basins and Their Energy and Mineral Resources* (Ed. by R. A. Price), pp. 29–35. American Geophysical. Union, Washington, DC.

Angevine, C.L., Heller, P.L. and **Paola, C.** (Eds) (1990) *Quantitative Sedimentary Basin Modeling. American Association of Petroleum Geologists, Continuing Education Course Note Series* **32**.

Ashworth, P.J., Best, J.L., Peakall, J. and Lorsong, J.A. (1999) The influence of aggradation rate on braided alluvial architecture: field study and physical scale-modelling of the Ashburton River gravels, Canterbury Plains, New Zealand. In: *Current Research in Fluvial Sedimentology: Proceedings of the 6th International Conference on Fluvial Sedimentology* (Ed. by N. D. Smith and J. J. Rogers), *Spec. Publ. Int. Assoc. Sedimentol.*, **28,** 333–346.

Aurell, M., Bádenas, B., Bosence, D.W.J. and **Waltham, D.A.** (1998) Carbonate production and offshore transport on a Late Jurassic carbonate ramp (Kimmeridgian, Iberian Basin, NE Spain): evidence from outcrops and computer modelling. In: *Carbonate Ramps* (Ed. by V. P. Wright and T. P. Burchette), *Spec. Publ. Geol. Soc. London*, **149,** 137–161.

Aurell, M., Bosence, D. and **Waltham, D.** (1995) Carbonate ramp depositional systems from a late Jurassic epeiric platform (Iberian Basin, Spain): a combined computer modelling and outcrop analysis. *Sedimentology*, **42,** 75–94.

Baker, V.R. (1974) Paleohydraulic interpretation of Quaternary alluvium near Golden, Colorado. *Quaternary Res.*, **4,** 94–112.

Barros, A.P. and **Lettenmaier, D.P.** (1993) Multiscale aggregation and disaggregation of precipitation for regional hydroclimatological studies. In: *Macroscale Modelling of the Hydrosphere* (Ed. by W. B. Wilkinson), *IAHS Publ.*, **214,** 183–193.

Beaumont, C. (1978) The evolution of sedimentary basins on a viscoelastic lithosphere: theory and examples. *Geophys J. R. Astronom. Soc.*, **55,** 471–497.

Beaumont, C. (1981) Foreland basins. *Geophys J. R. Astronom. Soc.*, **65,** 291–329.

Beaumont, C., Fullsack, P. and **Hamilton, J.** (1992) Erosional control of active compressional orogens. In: *Thrust Tectonics* (Ed. by K. R. McClay), pp. 1–18. Chapman & Hall, London.

Beerbower, J.R. (1964) Cyclothem and cyclic depositional mechanisms in alluvial plain sedimentation. *Kansas Geol. Surv. Bull.*, **169,** 31–42.

Begin, Z.B., Meyer, D.F. and **Schumm, S.S.** (1981) Development of longitudinal profiles of alluvial channels in response to base-level lowering. *Earth Surf. Process. Landforms*, **6,** 49–68.

Bice, D. (1988) Synthetic stratigraphy of carbonate platform and basin systems. *Geology*, **16,** 703–706.

Bice, D.M. (1991) Computer simulation of carbonate platform and basin systems. In: *Sedimentary Modeling: Computer Simulations and Methods for Improved Parameter Definition* (Ed. by E. K. Franseen, W. L. Watney, C. G. S. C. Kendall and W. Ross), *Kansas Geol. Surv. Bull.*, **233,** 431–447.

Blum, M.D. and **Törnqvist, T.** (2000) Fluvial response to climate and sea-level change: a review and look forward. *Sedimentology*, **47** (Suppl. 1) 2–48.

Bosence, D. and **Waltham, D.** (1990) Computer modeling of the internal architecture of carbonate platforms. *Geology*, **18,** 26–30.

Bosence, D.W.J., Pomar, L., Waltham, D.A. and Lankester, T.H.G. (1994) Computer modelling a Miocene carbonate platform, Mallorca, Spain. *AAPG Bull.*, **78,** 247–266.

Bosscher, H. and **Schlager, W.** (1992) Computer simulation of reef growth. *Sedimentology*, **39,** 503–512.

Bosscher, H. and **Schlager, W.** (1993) Accumulation rates of carbonate platforms. *J. Geol.*, **101,** 345–355.

Bosscher, H. and **Southam, J.** (1992) CARBPLAT – A computer model to simulate the development of carbonate platforms. *Geology*, **20,** 235–238.

Bridge, J.S. and **Bennett, S.J.** (1992) A model for the entrainment and transport of sediment grains of mixed sizes, shapes, and densities. *Water Resour. Res.*, **28,** 337–363.

Bridge, J.S. and **Leeder, M.R.** (1979) A simulation model of alluvial stratigraphy. *Sedimentology*, **26,** 617–644.

Bridge, J.S. and **Mackey, S.D.** (1993a) A revised alluvial stratigraphy model. In: *Alluvial Sedimentation* (Ed. by M. Marzo and C. Puigdefabregas), *Int. Assoc. Sedimentol. Spec. Publ.*, **17**, 319–336.

Bridge, J.S. and **Mackey, S.D.** (1993b) A theoretical study of fluvial sandstone body dimensions. In: *The Geological Modelling of Hydrocarbon Reservoirs and Outcrop Analogues* (Ed. by S. S. Flint and I. D.Bryant), *Int. Assoc. Sedimentol. Spec. Publ.*, **15**, 213–236.

Briggs, L.I. and **Pollack, H.N.** (1967) Digital model of evaporite sedimentation. *Science*, **155,** 453–456.

Bruun, P. (1962) Sea-level rise as a cause of shore erosion. *J. Waterways Harbors Division, ASCE*, **88,** 117–130.

Bryant, M., Falk, P. and **Paola, C.** (1995) Experimental study of avulsion frequency and rate of deposition. *Geology*, **23,** 365–368.

Burgess, P.M. and **Allen, P.A.** (1996) A forward-modelling analysis of the controls on sequence stratigraphic geometries. In: *Sequence Stratigraphy in British Geology* (Ed. by S. P. Hesselbo and D. N. Parkinson), *Spec. Publ. Geol. Soc. London*, **103,** 9–24.

Burton, R., Kendall, C.G.S.C. and **Lerche, I.** (1987) Out of our depth: On the impossibility of fathoming eustasy from the stratigraphic record. *Earth-Sci. Rev.*, **24,** 237–277.

Cacchione, D.A. and **Drake, D.E.** (1990) Shelf sediment transport: An overview with applications to the Northern California continental shelf. In: *The Sea*, Vol. 9*: Ocean Engineering Science* (Ed. by B. LeMehaute and D. M. Hanes), pp. 729–773. John Wiley and Sons, Inc., New York.

Cant, D.J. (1989) Simple equations of sedimentation: applications to sequence stratigraphy. *Basin Res.*, **2,** 73–81.

Cant, D.J. (1991) Geometric modelling of facies migration: theoretical development of facies successions and local unconformities. *Basin Res.*, **3,** 51–62.

Carey, J.S., Swift, D.J.P., Steckler, M., Reed, C.W. and **Niedoroda, A.** (1999) High resolution sequence stratigraphic modelling: effects of sedimentation processes. In: *Numerical Experiments in Stratigraphy* (Ed. by J. Harbaugh, L. Watney, G. Rankey, R. Slingerland, R. Goldstein and E. Franseen), *Spec. Publ. SEPM*, **62**, 151–164.

Chase, C.G. (1992) Fluvial landsculpting and the fractal dimension of topography. *Geomorphology*, **5,** 39–57.

Church, M. and **Rood, K.** (1983) *Catalogue of Alluvial River Channel Regime Data.* University of British Columbia, Department of Geography, Vancouver, BC.

Cisne, J.L., Gildner, R.F. and **Rabe, B.D.** (1984) Epeiric sedimentation and sea-level: synthetic ecostratigraphy. *Lethaia*, **17,** 267–288.

Cloetingh, S. (1986) Intraplate stresses: a new tectonic mechanism for fluctuations of relative sea-level. *Geology*, **14,** 617–620.

Cloetingh, S., McQueen, H. and **Lambeck, K.** (1985) On a tectonic mechanism for regional sealevel variations. *Earth Planet. Sci. Lett*, **75,** 157–166.

Cochran, J.R. (1983) Effects of finite rifting times on the development of sedimentary basins. *Earth Planet. Sci. Lett*, **66,** 289–302.

Collier, R.E.L., Leeder, M.R. and **Maynard, J.R.** (1990) Transgressions and regressions: a model for the influence of tectonic subsidence, deposition and eustasy, with application to Quaternary and Carboniferous examples. *Geol. Mag.*, **127,** 117–128.

Cotter, E. (1971) Paleoflow characteristics of a late Cretaceous river in Utah from analysis of sedimentary structures in the Ferron sandstone. *J. Sedim. Petrol.*, **41,** 129–138.

Cross, T.A. (Ed.) (1990) *Quantitative Dynamic Stratigraphy.* Prentice Hall, Englewood Cliffs, New Jersey.

Cui, Y., Parker, G. and **Paola, C.** (1996) Numerical simuation of aggradation and downstream fining. *J. Hydraulic Res.*, **34,** 185–204.

Dade, W.B. and **Friend, P.F.** (1998) Grain size, sediment-transport regime and channel slope in alluvial rivers. *J. Geol.*, **106,** 661–675.

Demicco, R.V. (1998) CYCOPATH 2D – a two-dimensional, forward model of cyclic sedimentation on carbonate platforms. *Computers & Geosciences*, **24,** 405–423.

Demicco, R.V. and **Spencer, R.J.** (1989) MAPS – a BASIC program to model accumulation of platform sediments. *Computers & Geosciences*, **15,** 95–105.

Demicco, R.V., Spencer, R.J., Waters, B.B. and **Cloyd, K.C.** (1991) Two dimensional computer models of a Cambrian carbonate shelf deposit. In: *Sedimentary Modeling: Computer Simulations and Methods for Improved Parameter Definition* (Ed. by E. K. Franseen, W. L. Watney, C. G. S. C. Kendall and W. Ross), *Kansas Geol. Surv. Bull.*, **233,** 463–472.

Dietz, R.S. (1963) Collapsing continental rises: an actualistic concept of geosynclines and mountain building. *J. Geol.*, **71,** 314–333.

Driscoll, N.W. and **Karner, G.D.** (1999) Three-dimensional quantitative modeling of clinoform development. *Mar Geol*, **154,** 383–398.

Eberli, G.P., Kendall, C.G.S.C., Moore, P., Whittle, G.L. and **Cannon, R.** (1994) Testing a seismic interpretation of Great Bahama Bank with a computer simulation. *AAPG Bull.*, **78,** 981–1004.

Emery, D. and **Myers, K.** (Eds) (1996) *Sequence Stratigraphy*. Blackwell Science, Oxford.

Engelund, F. and **Hansen, E.** (1967) *A Monograph on Sediment Transport in Alluvial Streams*. Teknisk Forlag, Copenhagen.

Ericksen, M.C. and **Slingerland, R.** (1990) Numerical simulations of tidal and wind-driven circulation in the Cretaceous Interior Seaway of North America. *Geol. Soc. Am. Bull.*, **102,** 1499–1516.

Ethridge, F.G. and **Schumm, S.A.** (1978) Reconstructing paleochannel morphologic and flow characteristics: methodology, limitations, and assessment. In: *Fluvial Sedimentology* (Ed. by A. D. Miall), *Can. Soc. Petrol. Geol. Mem.*, **5,** 703–721.

Flemings, P.B. and **Grotzinger, J.P.** (1996) STRATA: Freeware for Solving Classic Stratigraphic Problems. *GSA Today*, **6,** 1–7.

Flemings, P.B. and **Jordan, T.E.** (1989) A synthetic stratigraphic model of foreland basin development. *J. Geophys. Res.*, **94,** 3851–3866.

Flemings, P.B. and **Jordan, T.E.** (1990) Stratigraphic modeling of foreland basins: Interpreting thrust deformation and lithosphere rheology. *Geology*, **18,** 430–434.

Franseen, E.K., Watney, W.L., Kendall, C.G.S.C. and **Ross, W.** (Eds) (1991) *Sedimentary Modeling: Computer Simulations and Methods for Improved Parameter Definition. Kansas Geol. Surv. Bull.*, **233**.

Gaffin, S.R. and **Maasch, K.A.** (1991) Anomalous cyclicity in climate and stratigraphy and modeling nonlinear oscillations. *J. Geophys. Res.*, **96,** 6701–6711.

Galewsky, J. (1998) The dynamics of foreland basin carbonate platforms: tectonic and eustatic controls. *Basin Res.*, **10,** 409–416.

Garde, R.J., Ranga Raju, K.G. and **Mehta, P.J.** (1981) Bed level variation in aggrading alluvial streams. *International Association for Hydraulic Research Congress*, **19,** 247–253.

Gardner, T.W. (1983) Paleohydrology and paleomorphology of a Carboniferous, meandering, fluvial sandstone. *J. Sediment. Petrol.*, **53,** 991–1005.

Gildner, R.F. (1990) STRIPPER: An interactive backstripping, decompaction and geohistory program. In: *Quantitative Dynamic Stratigraphy* (Ed. by T. A. Cross), pp. 165–180. Prentice Hall, Englewood Cliffs, NJ.

Gildner, R.F. and **Cisne, J.L.** (1990) Quantitative modeling of carbonate stratigraphy and water-depth history using depth-dependent sediment accumulation function. In: *Quantitative Dynamic Stratigraphy* (Ed. by T. A. Cross), pp. 417–432. Prentice Hall, Englewood Cliffs, New Jersey.

Gill, M.A. (1983) Diffusion model for degrading channels. *J. Hydraulic Res.*, **21,** 369–378.

Ginsburg, R.N. (1971) Landward movement of carbonate mud: new model for regressive cycles in carbonates. *Am. Assoc. Petrol. Geol. Bull.*, **55,** 340.

Giorgi, F. and **Mearns, L.O.** (1991) Approaches to the simulation of regional climate change: a review. *Rev. Geophysics*, **29,** 191–216.

Goldenfeld, N. and **Kadanoff, L.P.** (1999) Simple lessons from complexity. *Science*, **284,** 87–89.

Goldhammer, R.K., Dunn, P.A. and **Hardie, L.A.** (1987) High-frequency glacio-eustatic sealevel oscillations with Milankovitch characteristics recorded in Middle Triassic platform carbonates in northern Italy. *Am. J. Sci.*, **287,** 853–892.

Goldhammer, R.K. and **Harris, M.T.** (1989) Eustatic controls on the stratigraphy and geometry of the Latemar buildup (Middle Triassic), the Dolomites of northern Italy. In: *Controls on Carbonate Platform and Basin Development, Spec. Publ. Soc. Econ. Paleont. Miner.*, **44,** 323–338.

Goldhammer, R.K., Lehmann, P.J. and **Dunn, P.A.** (1993) The origin of high-frequency platform carbonate cycles and third-order sequences (Lower Ordovician El Paso Group, West Texas): constraints from outcrop data and stratigraphic modeling. *J. Sedim. Petrol.*, **63,** 318–359.

Haq, B.U., Hardenbol, J. and **Vail, P.R.** (1987) Chronology of fluctuating sea-levels since the Triassic (250 million years ago to present). *Science*, **235,** 1156–1167.

Harbaugh, J., Watney, L., Rankey, G., Slingerland, R., Goldstein, R. and **Franseen, E.** (Eds) (1999) *Numerical Experiments in Stratigraphy. Spec. Publ. SEPM*, **62**.

Harbaugh, J.W. (1993) Simulating geologic processes that create sedimentary basins: incorporation of a carbonate-ecology model. In: *Computerized Basin Analysis: the Prognosis of Energy and Mineral Resources* (Ed. by J. Harff and D. F. Merriam), pp. 11–23. Plenum Press, New York.

Harbaugh, J.W. and **Bonham-Carter, G.** (1970) *Computer Simulation in Geology*. John Wiley & Sons, New York.

Harbaugh, J. W. (1966) Mathematical simulation of marine sedimentation. *Kansas Geol. Surv. Computer Contrib.*, **1**.

Hardy, S. and **Poblet, J.** (1994) Geometric and numerical model of progressive limb rotation in detachment folds. *Geology*, **22,** 371–374.

Harris, M.T. (1991) Carbonate progradation geometry and significance of sediment accumulation patterns and syndepositional relief: Comparison of modeling results with outcrop data. In: *Sedimentary Modeling: Computer Simulations and Methods for Improved Parameter Definition* (Ed. by E. K. Franseen, W. L. Watney, C. G. S. C. Kendall and W. Ross), *Kansas Geol. Surv. Bull.*, **233,** 509–517.

Harrison, W.J. (1990) Modeling fluid/rock interactions in sedimentary basins. In: *Quantitative Dynamic Stratigraphy* (Ed. by T. A. Cross), pp. 195–231. Prentice Hall, Englewood Cliffs, New Jersey.

Haxby, W.F., Turcotte, D.L. and **Bird, J.M.** (1976) Thermal and mechanical evolution of the Michigan Basin. *Tectonophysics*, **36,** 57–75.

Helland-Hansen, W., Kendall, C.G.S.C., Lerche, I. and **Nakayama, K.** (1988) A simulation of continental basin margin sedimentation in response to crustal movements, eustatic sea-level change, and sediment accumulation rates. *Mathemat. Geol*, **20,** 777–802.

Heller, P.L., Burns, B. and **Marzo, M.** (1993) Stratigraphic solution sets for determining the roles of sediment supply, subsidence, and sea-level on transgressions and regressions. *Geology*, **21,** 747–750.

Heller, P.L. and **Paola, C.** (1992) The large-scale dynamics of grain-size variation in alluvial basins, 2: application to syntectonic conglomerate. *Basin Res.*, **4,** 91–102.

Heller, P.L. and **Paola, C.** (1996) Downstream changes in alluvial architecture: an exploration of controls on channel-stacking patterns. *J. Sedim. Res.*, **66,** 297–306.

Hoey, T.B. and **Ferguson, R.** (1994) Numerical simulation of downstream fining by selective transport in gravel bed rivers: Model development and illustration. *Water Resour. Res.*, **30,** 2251–2260.

Hooke, R.L. (1968) Model geology: prototype and laboratory streams: discussion. *Geol. Soc. Am. Bull.*, **79,** 391–394.

Howard, A.D. (1994) A detachment-limited model of drainage basin evolution. *Water Resour. Res.*, **39,** 2261–2285.

Humphrey, N.F. and **Heller, P.L.** (1995) Natural oscillations in coupled geomorphic systems: an alternative origin for cyclic sedimentation. *Geology*, **23,** 499–502.

Jain, S.C. (1981) River bed aggradation due to overloading. *J. Hydraulics Division ASCE*, **107,** 120–124.

Jaramillo, W.F. and **Jain, S.C.** (1984) Aggradation and degradation of alluvial-channel beds. *J. Hydraulic Engineering*, **110,** 1072–1085.

Jarvis, G.T. and **McKenzie, D.P.** (1980) Sedimentary basin formation with finite extension rates. *Earth Planet. Sci. Lett*, **48,** 42–52.

Jervey, M.T. (1988) Quantitative geological modeling of siliciclastic rock sequences and their seismic expression. In: *Sea-Level Changes—an Integrated Approach* (Ed. by C. K. Wilgus, B. S. Hastings, H. W. Posamentier, J. Van Wagoner, C. A. Ross and C. G. S. C. Kendall), *Spec. Publ. Soc. Econ. Paleont. Miner.*, **42,** 47–69.

Jin, Y., McNutt, M.K. and **Zhu, Y.** (1994) Evidence from gravity and topography data for folding of Tibet. *Nature*, **371,** 669–674.

Johnson, D.D. and **Beaumont, C.** (1995) Preliminary results from a planform kinematic model of orogen evolution, surface processes and the development of clastic foreland basin stratigraphy. In: *Stratigraphic Evolution of Foreland Basins* (Ed. by S. L. Dorobek and G. M. Ross), *Spec. Publ. Soc. Econ. Paleont. Miner.*, **52,** 3–24.

Jordan, T.E. (1981) Thrust loads and foreland basin evolution. *Am. Assoc. Petrol. Geol. Bull.*, **65,** 2506–2520.

Jordan, T.E. and **Flemings, P.B.** (1990) From geodynamic models to basin fill – a stratigraphic perspective. In: *Quantitative Dynamic Stratigraphy* (Ed. by T. A. Cross), pp. 149–163. Prentice Hall, Englewood Cliffs, New Jersey.

Jordan, T.E. and **Flemings, P.B.** (1991) Large-scale stratigraphic architecture, eustatic variation, and unsteady tectonism: a theoretical evaluation. *J. Geophys. Res.*, **96,** 6681–6699.

Karner, G.D., Steckler, M.S. and **Thorne, J.A.** (1983) Long-term thermo-mechanical properties of the lithosphere. *Nature*, **304,** 250–252.

Kaufman, P., Grotzinger, J.P. and **McCormick, D.S.** (1991) Depth-dependent diffusion algorithm for simulation of sedimentation in shallow marine depositional systems. In: *Sedimentary Modeling: Computer Simulations and Methods for Improved Parameter Definition* (Ed. by E. K. Franseen, W. L. Watney, C. G. S. C. Kendall and W. Ross), *Kansas Geol. Surv. Bull.*, **233,** 489–508.

Kendall, C.G.S.C., Moore, P., Strobel, J., Cannon, R., Perlmutter, M., Bezdek, J. and **Biswas, G.** (1991a) Simulation of the sedimentary fill of basins. In: *Sedimentary Modeling: Computer Simulations and Methods for Improved Parameter Definition* (Ed. by E. K. Franseen, W. L. Watney, C. G. S. C. Kendall and W. Ross), *Kansas Geol. Surv. Bull.*, **233,** 9–30.

Kendall, C.G.S.C., Strobel, J., Cannon, R., Bezdek, J. and **Biswas, G.** (1991b) The simulation of the sedimentary fill of basins. *J. Geophys. Res.*, **96,** 6911–6929.

Kenyon, P.M. and **Turcotte, D.L.** (1985) Morphology of a delta prograding by bulk sediment transport. *Geol Soc. Am. Bull.*, **96,** 1457–1465.

Kneller, B. and **Buckee, C.** (2000) The structure and fluid mechanics of turbidity currents: a review of some recent studies and their geological implications. *Sedimentology*, **47** (Suppl. 1), 63–94.

Kochel, R.C. and **Baker, V.R.** (1982) Paleoflood hydrology. *Science*, **215,** 353–361.

Kocurek, G. and **Havholm, K.G.** (1994) Eolian sequence stratigraphy—a conceptual framework. In: *Siliciclastic Sequence Stratigraphy: Recent Developments and Applications* (Ed. by P. Weimer and H. Posamentier), *AAPG Mem.*, **58,** 393–409.

Kooi, H. and **Beaumont, C.** (1994) Escarpment evolution on high-elevation rifted margins: Insights derived from a surface-processes model that combines diffusion, advection, and reaction. *J. Geophys. Res.*, **99,** 12191–12209.

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

Krumbein, W.C. and **Pettijohn, F.J.** (1938) *Manual of Sedimentary Petrography*. Appleton-Century-Crofts, Inc., New York.

Kruse, S.E. and Royden, L.H. (1994) Bending and unbending of an elastic lithosphere: the Cenozoic history of the Apennine and Dinaride foredeep basins. *Tectonics*, **13,** 278–302.

Kusznir, N. and **Karner, G.** (1985) Dependence of the flexural rigidity of the continental lithosphere on rheology and temperature. *Nature*, **316,** 138–142.

Lawrence, D.T. (1994) Evaluation of eustasy, subsidence, and sediment input as controls on depositional sequence geometries and the synchroneity of sequence boundaries. In: *Siliciclastic Sequence Stratigraphy: Recent Developments and Applications* (Ed. by P. Weimer and H. Posamentier), *AAPG Mem.*, **58,** 337–367.

Lawrence, D.T., Doyle, M. and **Aigner, T.** (1990) Stratigraphic simulation of sedimentary basins: Concepts and calibration. *AAPG Bull.*, **74,** 273–295.

Lee, M.-K. (1997) Predicting diagenetic effects of groundwater flow in sedimentary basins: a modeling approach with examples. In: *Basin-Wide Diagenetic Patterns: Integrated Petrologic, Geochemical, and Hydrologic Considerations* (Ed. by I. P. Montañez, J. M. Gregg and K. L. Shelton), *Spec. Publ. Soc. Econ. Paleont. Miner.*, **57,** 3–14.

Lee, Y.-H. and **Harbaugh, J.W.** (1992) Stanford's SEDSIM project: dynamic three-dimensional simulation of geologic processes that affect clastic sediments. In: *Computer Graphics in Geology* (Ed. by R. Pflug and J. W. Harbaugh), pp. 113–127. Springer-Verlag, Berlin.

Leeder, M.R. (1978) A quantitative stratigraphic model for alluvium, with special reference to channel deposit density and interconnectedness. In: *Fluvial Sedimentology* (Ed. by A. D. Miall), *Can. Soc. Petrol. Geol. Mem.*, **5,** 587–596.

Leeder, M. (1999) *Sedimentology and Sedimentary Basins: from Turbulence to Tectonics*. Blackwell Science, Oxford.

Leeder, M.R. and **Gawthorpe, R.L.** (1987) Sedimentary models for extensional tilt block/half-graben basins. *Spec. Publ. Geol. Soc. London*, **28,** 139–152.

Lerche, I., Dromgoole, E., Kendall, C.G.S.C., Walter, L.M. and **Scaturo, D.** (1987) Geometry of carbonate bodies: a quantitative investigation of factors influencing their evolution. *Carbonates Evaporites*, **2,** 15–42.

Lessenger, M.A. and **Cross, T.A.** (1996) An inverse stratigraphic simulation model – Is stratigraphic inversion possible? *Energy Exploration & Exploitation*, **14,** 627–637.

Lorenz, E.N. (1963) Deterministic nonperiodic flow. *J. Atmospheric Sci.*, **20,** 130–141.

Mackey, S.D. and **Bridge, J.S.** (1995) Three-dimensional model of alluvial stratigraphy: theory and application. *J. Sedim. Res.*, **B65,** 7–31.

Mahlman, J.D. (1998) Science and nonscience concerning human-caused climate warming. *Annu. Rev. Energy Environ.*, **23,** 83–106.

Martel, A.T., Allen, P.A. and **Slingerland, R.** (1994) Use of tidal-circulation modeling in paleogeographical studies: an example from the Tertiary of the Alpine perimeter. *Geology*, **22,** 925–928.

Martinez, P.A. (1992) Three-dimensional simulation of littoral transport. In: *Computer Graphics in Geology* (Ed. by R. Pflug and J. W. Harbaugh), pp. 129–142. Springer-Verlag, Berlin.

Martinez, P.A. and **Harbaugh, J.W.** (1993) *Simulating Nearshore Environments*. Pergamon Press, Oxford.

McKenzie, D.P. (1978) Some remarks on the development of sedimentary basins. *Earth Planet. Sci. Lett.*, **40,** 25–32.

McNutt, M.K., Diament, M. and **Kogan, M.G.** (1988) Variations of elastic plate thickness at continental thrust belts. *J. Geophys. Res.*, **93,** 8825–8838.

Meckel, L.D. (1967) Origin of Pottsville conglomerates (Pennsylvanian) in the central Appalachians. *Geol Soc. Am. Bull.*, **78,** 223–258.

Middleton, G.V. and **Southard, J.B.** (1984) *Mechanics of Sediment Movement*. *SEPM Eastern Section Short Course*, **3**.

Miller, K.G., Thompson, P.R. and **Kent, D.V.** (1993) Integrated Late Eocene – Oligocene stratigraphy of the Alabama coastal plain: correlation of hiatuses and stratal surfaces to glacioeustatic lowerings. *Paleoceanography*, **8,** 313–331.

Mountney, N.P. and **Westbrook, G.K.** (1997) Quantitative analysis of Miocene to Recent forearc basin evolution along the Colombian convergent margin. *Basin Res.*, **9,** 177–196.

Murray, A.B. and **Paola, C.** (1994) A cellular model of braided rivers. *Nature*, **371,** 54–57.

Murray, A.B. and **Paola, C.** (1997) Properties of a cellular braided stream model. *Earth Surf. Process. Landforms*, **22,** 1001–1025.

Nicolis, G. and **Nicolis, C.** (1991) Nonlinear dynamic systems in the geosciences. In: *Sedimentary Modeling: Computer Simulations* and Methods for Improved Parameter Definition (Ed. by E. K. Franseen, W. L. Watney, C. G. S. C. Kendall and W. Ross), *Kansas Geol. Surv. Bull.*, **233,** 33–42.

Niedoroda, A.M., Reed, C.W., Swift, D.J.P., Arata, H. and **Hoyanagi, K.** (1995) Modeling shore-normal large-scale coastal evolution. *Mar. Geol.*, **126,** 181–199.

van Niekerk, A., Vogel, K.R., Slingerland, R.L. and **Bridge, J.S.** (1992) Routing of heterogeneous sediments over movable bed: model development. *J. Hydraulic Engineering*, **118,** 246–262.

Nittrouer, C.A. (1999) STRATAFORM: overview of its design and synthesis of its results. *Mar. Geol.*, **154,** 3–12.

Nittrouer, C.A. and **Wright, L.D.** (1994) Transport of particles across continental shelves. *Rev. Geophysics*, **32,** 85–113.

Nordlund, U. (1999) FUZZIM: forward stratigraphic modeling made simple. *Comput. Geosci.*, **25**, 449–456.

Nordlund, U. (1996) Formalizing geological knowledge—with an example of modeling stratigraphy using fuzzy logic. *J. Sedim. Res.*, **66,** 689–698.

Nummedal, D., Riley, G.W. and **Templet, P.L.** (1993) High-resolution sequence architecture: a chronostratigraphic model based on equilibrium profile studies. In: *Sequence Stratigraphy and Facies Associations* (Ed. by H. W. Posamentier, C. P. Summerhayes, B. U. Haq and G. P. Allen), *Spec. Publ. Int. Assoc. Sedimentol.*, **18,** 55–68.

Paola, C. (1988) Subsidence and gravel transport in alluvial basins. In: *New Perspectives in Basin Analysis* (Ed. by K. L. Kleinspehn and C. Paola), pp. 231–243. Springer-Verlag, New York.

Paola, C. (1990) A simple basin-filling model for coarse-grained alluvial systems. In: *Quantitative Dynamic Stratigraphy* (Ed. by T. A. Cross), pp. 363–374. Prentice Hall, Englewood Cliffs, New Jersey.

Paola, C. (1996) Incoherent structure: turbulence as a metaphor for stream braiding. In: *Coherent Flow Structures in Open Channels* (Ed. by P. J. Ashworth, S. J. Bennett, J. L. Best and S. J. McLelland), pp. 705–723. John Wiley & Sons, Chichester.

Paola, C., Heller, P.L. and **Angevine, C.L.** (1992) The large-scale dynamics of grain-size variation in alluvial basins, 1: Theory. *Basin Res.*, **4,** 73–90.

Paola, C., Parker, G., Mohrig, D.C. and **Whipple, K.X.** (1999) The influence of transport fluctuations on spatially averaged topography on a sandy, braided fluvial fan. In: *Numerical Experiments in Stratigraphy: Recent Advances in Stratigraphic and Sedimentologic Computer Simulations* (Ed. by J. Harbaugh, L. Watney, G. Rankey, R. Slingerland, R. Goldstein and E. Franseen), *Spec. Publ. Soc. Econ. Paleont. Miner.,* **62,** 211–218.

Paola, C. and **Seal, R.** (1995) Grain-size patchiness as a cause of selective deposition and downstream fining. *Water Resour. Res.*, **31,** 1395–1407.

Parker, G. (1978a) Self-formed straight rivers with equilibrium banks and mobile bed. Part 1. The sand-silt river. *J. Fluid Mechanics*, **89,** 109–125.

Parker, G. (1978b) Self-formed straight rivers with equilibrium banks and mobile bed. Part 2. The gravel river. *J. Fluid Mechanics*, **89,** 127–146.

Parker, G. (1991) Selective sorting and abrasion of river gravel. I: theory. *J. Hydraulic Engineering*, **117,** 131–149.

Parker, G., Paola, C., Whipple, K.X. and **Mohrig, D.C.** (1998a) Alluvial fans formed by channelized fluvial and sheet flow. I: theory. *J. Hydraulic Engineering*, **124,** 985–995.

Parker, G., Paola, C., Whipple, K.X., Mohrig, D.C., Toro-Escobar, C.M., Halverson, M. and **Skoglund, T.W.** (1998b) Alluvial fans formed by channelized fluvial and sheet flow. II: application. *J. Hydraulic Engineering*, **124,** 996–1004.

Peakall, J., Ashworth, P. and **Best, J.** (1996) Physical modelling in fluvial geomorphology: Principles, applications and unresolved issues. In: *The Scientific Nature of Geomorphology* (Ed. by B. L. Rhoads and C. E. Thorn), pp. 221–253. John Wiley & Sons, Chichester.

Pelletier, B.R. (1958) Pocono paleocurrents in Pennsylvania and Maryland. *Geol. Soc. Am. Bull.*, **69,** 1033–1064.

Pelletier, J.D. and **Turcotte, D.L.** (1997) Synthetic stratigraphy with a stochastic diffusion model of sedimentation. *J. Sedim. Res.*, **67,** 1060–1067.

Perlmutter, M.A. and **Matthews, M.D.** (1990) Global cyclostratigraphy — a model. In: *Quantitative Dynamic Stratigraphy* (Ed. by T. A. Cross), pp. 233–260. Prentice Hall, Englewood Cliffs, New Jersey.

Perlmutter, M.A., Radovich, B.J., Matthews, M.D. and **Kendall, C.G.S.C.** (1998) The impact of high-frequency sedimentation cycles on stratigraphic interpretation. In: *Sequence Stratigraphy — Concepts and Applications* (Ed.

by F. M. Gradstein, K. O. Sandvik and N. J. Milton), *Norwegian Petrol. Soc. Spec. Publ.*, **8,** 141–170.

Person, M., Raffensperger, J.P., Ge, S. and **Garven, G.** (1996) Basin-scale hydrogeologic modeling. *Rev. Geophysics*, **34,** 61–87.

Pettijohn, F.J. (1957) *Sedimentary Rocks*. Harper, New York.

Pettijohn, F.J. and **Potter, P.E.** (1964) *Atlas and Glossary of Primary Sedimentary Structures*. Springer-Verlag, Berlin.

Pettijohn, F.J., Potter, P.E. and **Siever, R.** (1973) *Sand and Sandstone*. Springer-Verlag, New York.

Pirmez, C., Pratson, L.F. and **Steckler, M.S.** (1998) Clinoform development by advection-diffusion of suspended sediment: modeling and comparison with natural systems. *J. Geophys. Res.*, **103,** 24141–24157.

Pitman, W.C. (1978) Relationship between eustacy and stratigraphic sequences on passive margins. *Geol. Soc. Am. Bull.*, **89,** 1389–1403.

Potter, P.E. and **Pettijohn, F.J.** (1977) *Paleocurrents and Basin Analysis*. Springer-Verlag, Berlin.

Read, J.F., Osleger, D. and **Elrick, M.** (1991) Two-dimensional modeling of carbonate ramp sequences and component cycles. In: *Sedimentary Modeling: Computer Simulations and Methods for Improved Parameter Definition* (Ed. by E. K. Franseen, W. L. Watney, C. G. S. C. Kendall and W. Ross), *Kansas Geol. Surv. Bull.*, **233,** 473–488.

Reynolds, D.J., Steckler, M.S. and **Coakley, B.J.** (1991) The role of the sediment load in sequence stratigraphy: the influence of flexural isostasy and compaction. *J. Geophys. Res.*, **96,** 6931–6949.

Ribberink, J.S. and **van der Sande, J.T.M.** (1985) Aggradation in rivers due to overloading – analytical approaches. *J. Hydraulic Res.*, **23,** 273–284.

Rivenæs, J.C. (1992) Application of a dual-lithology, depth-dependent diffusion equation in stratigraphic simulation. *Basin Res.*, **4,** 133–146.

Rivenæs, J.C. (1997) Impact of sediment transport efficiency on large-scale sequence architecture: results from stratigraphic computer simulation. *Basin Res.*, **9,** 91–105.

Robinson, R.A.J. and **Slingerland, R.L.** (1998) Origin of fluvial grain-size trends in a foreland basin: the Pocono Formation on the the central Appalachian Basin. *J. Sedim. Res.*, **68,** 473–486.

Ross, W.C. (1991) Cyclic stratigraphy, sequence stratigraphy, and stratigraphic modeling from 1964 to 1989: Twenty-five years of progress?. In: *Sedimentary Modeling: Computer Simulations and Methods for Improved Parameter Definition* (Ed. by E. K. Franseen, W. L. Watney, C. G. S. C. Kendall and W. Ross), *Kansas Geol. Surv. Bull.*, **233,** 3–8.

Ross, W.C., Watts, D.E. and **May, J.A.** (1995) Insights from stratigraphic modelling: Mud-limited versus sand-limited depositional systems. *AAPG Bull.*, **79,** 231–258.

Royden, L. and **Keen, C.E.** (1980) Rifting processes and thermal evolution of the continental margin of eastern Canada determined from subsidence curves. *Earth Planet. Sci. Lett.*, **51,** 343–361.

Rubin, D.M. (1992) Use of forecasting signatures to help distinguish periodicity, randomness, and chaos in ripples and other spatial patterns. *Chaos*, **2,** 525–535.

Sambrook Smith, G.H. and **Ferguson, R.I.** (1995) The gravel-sand transition along river channels. *J. Sedim. Res.*, **A65,** 423–430.

Sapozhnikov, V.B. and **Foufoula-Georgiou, E.** (1996) Self-affinity in braided rivers. *Water Resour. Res.*, **32,** 1429–1439.

Sapozhnikov, V.B. and **Foufoula-Georgiou, E.** (1997) Experimental evidence of dynamic scaling and indications of self-organized criticality in braided rivers. *Water Resour. Res.*, **33,** 1983–1991.

Sawyer, D.S., Hsui, A.T. and **Toksoz, M.N.** (1987) Extension, subsidence, and thermal evolution of the Los Angeles Basin — a two-dimensional model. *Tectonophysics*, **133,** 15–32.

Schlager, W. (1981) The paradox of drowned reefs and carbonate platforms. *Geol. Soc. Am. Bull.*, **92,** 197–211.

Schlee, J. (1957) Upland gravels of southern Maryland. *Geol. Soc. Am. Bull.*, **68,** 1371–1430.

Schumm, S.A. (1977) *The Fluvial System*. John Wiley, New York.

Schumm, S.A. and **Lichty, R.W.** (1965) Time, space, and causality in geomorphology. *Am. J. Sci.*, **263,** 110–119.

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

Schwarzacher, W. (1966) Sedimentation in a subsiding basin. *Nature*, **5043,** 1349–1350.

Sen, A., Kendall, C.G.S.C. and **Levine, P.** (1999) Combining a computer simulation and eustatic events to date seismic sequence boundaries: a case study of the Neogene of the Bahamas. *Sediment. Geol.*, **125,** 47–59.

Shanley, K.W. and **McCabe, P.J.** (1991) Predicting facies architecture through sequence stratigraphy—An example from the Kaiparowits Plateau, Utah. *Geology*, **19,** 742–745.

Shanley, K.W. and **McCabe, P.J.** (1993) Alluvial architecture in a sequence stratigraphic framework: a case history from the Upper Cretaceous of southern Utah, USA. In: *The Geological Modelling of Hydrocarbon Reservoirs and Outcrop Analogues* (Ed. by S. S. Flint and I. D. Bryant), *Spec. Publ. Int. Assoc. Sedimentol.*, **15,** 21–56.

Shuster, M.W. and **Aigner, T.** (1994) Two-dimensional synthetic seismic and log cross sections from stratigraphic forward models. *AAPG Bull.*, **78,** 409–431.

Sinclair, H.D., Coakley, B.J., Allen, P.A. and **Watts, A.B.** (1991) Simulation of foreland basin stratigraphy using a diffusion model of mountain belt uplift and erosion: an example from the Central Alps, Switzerland. *Tectonics*, **10,** 599–620.

Sleep, N.H. (1971) Thermal effects of the formation of Atlantic continental margins by continental break-up. *Geophys. J. Royal Astronom. Soc.*, **45,** 325–350.

Slingerland, R. (1990) Predictability and chaos in quantitative dynamic stratigraphy. In: *Quantitative Dynamic Stratigraphy* (Ed. by T. A. Cross), pp. 45–53. Prentice Hall, Englewood Cliffs, New Jersey.

Slingerland, R., Harbaugh, J.W. and **Furlong, K.** (1994) *Simulating Clastic Sedimentary Basins*. PTR Prentice Hall, Englewood Cliffs, New Jersey.

Sloss, L.L. (1962) Stratigraphic models in exploration. *J. Sedim. Petrol.*, **32,** 415–422.

Sloss, L.L. (1991) The tectonic factor in sea-level change: a countervailing view. *J. Geophys. Res.*, **96,** 6609–6617.

Smith, J.D. and **McLean, S.R.** (1977) Spatially averaged flow over a wavy surface. *J. Geophys. Res.*, **82,** 1735–1746.

Snow, R.S. and **Slingerland, R.L.** (1987) Mathematical modeling of graded river profiles. *J. Geol.*, **95,** 15–33.

Soni, J.P. (1981) Laboratory study of aggradation in alluvial channels. *J. Hydrol.*, **49,** 87–106.

Spencer, R.J. and **Demicco, R.V.** (1989) Computer models of carbonate platform cycles driven by subsidence and eustasy. *Geology*, **17,** 165–168.

Starfield, A.M. (1990) Qualitative, rule-based modeling. *Bioscience*, **40,** 601–604.

Starfield, A.M., Cumming, D.H.M., Taylor, R.D. and **Quadling, M.S.** (1993) A frame-based paradigm for dynamic ecosystem models. *AI Applications*, **7,** 1–13.

Starfield, A.M., Farm, B.P. and **Taylor, R.H.** (1989) A rule-based ecological model for the management of an estuarine lake. *Ecol. Modelling*, **46,** 107–119.

Steckler, M.S. (1999) High resolution sequence stratigraphic modeling: 1. The interplay of sedimentation, erosion and subsidence. In: *Numerical Experiments in Stratigraphy* (Ed. by J. Harbaugh, L. Watney, G. Rankey, R. Slingerland, R. Goldstein and E. Franseen), *Spec. Publ. Soc. Econ. Paleont. Miner.*, **62**, 139–150.

Steckler, M.S. (1985) Uplift and extension at the Gulf of Suez: indications of induced mantle convection. *Nature*, **317,** 135–139.

Steckler, M.S., Mountain, G.S., Miller, K.G. and **Christie-Blick, N.** (1999) Reconstruction of Tertiary progradation and clinoform development on the New Jersey passive margin by 2-D backstripping. *Mar. Geol.*, **154,** 399–420.

Steckler, M.S., Reynolds, D.J., Coakley, B.J., Swift, B.A. and **Jarrard, R.** (1993) Modelling passive margin sequence stratigraphy. In: *Sequence Stratigraphy and Facies Associations* (Ed. by H. W. Posamentier, C. P. Summerhayes, B. U. Haq and G. P. Allen), *Spec. Publ. Int. Assoc. Sedimentol.*, **18,** 19–41.

Steer, B.L. and **Abbott, P.L.** (1984) Paleohydrology of the Eocene Ballena Gravels, San Diego County, Caliornia. *Sedim. Geol.*, **38,** 181–216.

Sternberg, H. (1875) Untersuchungen über Längen und Querprofil geschiebeführnde Flüsse. *Z Bauwesen*, **25,** 483–506.

Strobel, J., Cannon, R., Kendall, C.G.S.C., Biswas, G. and **Bezdek, J.** (1989) Interactive (SEDPAK) simulation of clastic and carbonate sediments in shelf to basin settings. *Computers & Geosciences*, **15,** 1279–1290.

Swenson, J.B., Voller, V., Paola, C. and **Parker, G.** (in press) Modeling the stratigraphic response of fan-deltas to base-level forcing as a generalized Stefan problem. *European J. Appl. Mechanics*, in press.

Syvitski, J.P.M. and **Alcott, J.M.** (1993) GRAIN2: Predictions of particle size seaward of river mouths. *Computers & Geosciences*, **19,** 399–446.

Syvitski, J.P.M. and **Alcott, J.M.** (1995) RIVER3: Simulation of river discharge and sediment transport. *Computers & Geosciences*, **21,** 89–151.

Syvitski, J.P.M. and **Andrews, J.T.** (1994) Climate Change: Numerical modelling of sedimentation and coastal processes, Eastern Canadian Arctic. *Arctic Alpine Res.*, **26,** 199–212.

Syvitski, J.P.M. and **Daughney, S.** (1992) DELTA2: Delta progradation and basin filling. *Computers & Geosciences*, **18,** 839–897.

Syvitski, J.P.M., Morehead, M. and **Nicholson, M.** (1998a) HYDROTREND: a climate-driven hydrologic-transport model for predicting discharge and sediment to lakes or oceans. *Computers & Geosciences*, **24,** 51–68.

Syvitski, J.P.M., Nicholson, M., Skene, K. and **Morehead, M.D.** (1998b) PLUME1.1: Deposition of sediment from a fluvial plume. *Computers & Geosciences*, **24,** 159–171.

Syvitski, J.P.M., Pratson, L. and **Morehead, M.D.** (1997) EARTHWORKS: a large spatial scale numerical model to study the flux of sediment to ocean basins and reworking of deposits over various time scales. *EOS Suppl.*, **78,** F258.

ter Voorde, M., Ravnås, R., Færseth, R. and **Cloetingh, S.** (1997) Tectonic modelling of the Middle Jurassic synrift stratigraphy in the Oseberg-Brage area, northern Viking Graben. *Basin Res.*, **9,** 133–150.

Tetzlaff, D.M. (1990) Limits to the predictive ability of dynamic models that simulate clastic sedimentation. In: *Quantitative Dynamic Stratigraphy* (Ed. by T. A. Cross), pp. 55–65. Prentice Hall, Englewood Cliffs, New Jersey.

Tetzlaff, D.M. and **Harbaugh, J.W.** (1989) *Simulating Clastic Sedimentation*. Van Nostrand Reinhold, New York.

Tipper, J. (1991) Modelling the fill of sedimentary basins. *Exploration Geophys.*, **22,** 397–400.

Tipper, J.C. (1992) Landforms developing and basins filling: three-dimensional simulation of erosion, sediment transport, and deposition. In: *Computer Graphics in Geology. Lecture Notes in Earth Sciences, Vol. 41.* (Ed. by R. Pflug and J. W. Harbaugh), pp. 155–170. Springer-Verlag, Berlin.

Tucker, G.E. and **Slingerland, R.L.** (1994) Erosional dynamics, flexural isostasy, and long-lived escarpments: a numerical modeling study. *J. Geophys. Res.*, **99,** 12229–12243.

Turcotte, D.L. and **Schubert, G.** (1982) *Geodynamics, Applications of Continuum Physics to Geological Problems*. John Wiley & Sons, New York.

Van Wagoner, J.C., Mitchum, R.M., Campion, K.M. and **Rahmanian, V.D.** (Eds.) (1990) Siliciclastic Sequence Stratigraphy in Well Logs, Cores, and Outcrops: Concepts for High-Resolution Correlation of Time and Facies. *AAPG Methods in Exploration Series*, **7**.

Waltham, D. (1992) Mathematical modelling of sedimentary basin processes. *Mar. Petrol. Geol.*, **9,** 265–273.

Watney, W.L., Wong, J.-C. and **French,J.A. Jr** (1991) Computer simulation of Upper Pennsylvanian (Missourian) carbonate-dominated cycles in western Kansas. In: *Sedimentary Modeling: Computer Simulations and Methods for Improved Parameter Definition* (Ed. by E. K. Franseen, W. L. Watney, C. G. S. C. Kendall and W. Ross), *Kansas Geol. Surv. Bull.*, **233,** 415–430.

Watts, A.B., Karner, G.D. and **Steckler, M.S.** (1982) Lithospheric flexure and the evolution of sedimentary basins. *Phil. Trans. Royal Soc. London*, **305A,** 249–281.

Weimer, P. and **Posamentier, H.** (Eds) (1994) *Siliciclastic Sequence Stratigraphy: Recent Developments and Applications. AAPG Memoir*, **58**.

Wendebourg, J. and **Ulmer, J.W.D.** (1992) Modelling compaction and isostatic compensation in SEDSIM for basin analysis and subsurface fluid flow. In: *Computer Graphics in Geology* (Ed. by R. Pflug and J. W. Harbaugh), pp. 143–153. Springer-Verlag, Berlin.

Werner, B.T. (1995) Sand dune attractors. *Geology*, **23,** 1111–1114.

Werner, B.T. (1999) Complexity in natural landform patterns. *Science*, **284,** 102–104.

Werner, B.T. and **Fink, T.M.** (1993) Beach cusps as self-organized patterns. *Science*, **260,** 968–971.

Werner, B.T. and **Hallet, B.** (1993) Numerical simulation of self-organized stripes. *Nature*, **361,** 142–145.

Werner, B.T. and **Kocurek, G.** (1997) Bed-form dynamics: does the tail wag the dog? *Geology*, **25,** 771–774.

Willgoose, G., Bras, R.L. and **Rodriguez-Iturbe, I.** (1991) A coupled channel network growth and hillslope evolution model. I. Theory. *Water Resour. Res.*, **27,** 1671–1684.

Wood, L.J., Ethridge, F.G. and **Schumm, S.A.** (1993) The effects of rate of base-level fluctuation on coastal-plain, shelf and slope depositional systems; an experimental approach. In: *Sequence Stratigraphy and Facies Associations* (Ed. by H. W. Posamentier, C. P. Summerhayes, B. U. Haq and G. P. Allen), *Spec. Publ. Int. Assoc. Sedimentol.*, **18,** 43–53.

Wood, L.J., Ethridge, F.G. and **Schumm, S.A.** (1994) An experimental study of the influence of subaqueous shelf angles on coastal plain and shelf deposits. In: *Siliciclastic Sequence Stratigraphy: Recent Developments and Applications* (Ed. by P. Weimer and H. Posamentier), *AAPG Mem.*, **58,** 381–391.

Yalin, M.S. (1971) *Theory of Hydraulic Models*. MacMillan, London.

Yeakel, L.S. (1962) Tuscarora, Juniata and Bald Eagle paleocurrents and paleogeography in the central Appalachians. *Geol Soc. Am. Bull.*, **73,** 1515–1540.

Zhang, H. and **Kahawita, R.** (1987) Nonlinear model for aggradation in alluvial channels. *J. Hydraulic Engineering*, **113,** 353–369.

NOMENCLATURE

Symbol	Meaning	Dimensions (**L** length, **M** mass, **T** time)
b	stream width	$\mathbf{L}$
c	wave (phase) speed	$\mathbf{LT}^{-1}$
c_f	drag coefficient $\tau_0/(\pi u^2)$	
C_0	volumetic sediment concentration in the bed	
D	grain diameter	$\mathbf{L}$
f	is a function of	from context
$f_D(D)$	grain diameter probability density	$\mathbf{L}^{-1}$
g	gravitational acceleration	$\mathbf{LT}^{-2}$
h	flow depth	$\mathbf{L}$
H	base level elevation relative to a fixed datum; representative depth	$\mathbf{L}$
L	basin length; length scale	$\mathbf{L}$
p	*in situ* sediment production weight	$\mathbf{LT}^{-1}$
q_s	sediment flux (discharge) per unit width	$\mathbf{L}^2\mathbf{T}^{-1}$
q_{s^*}	dimensionless sediment flux $q_s/[g^{1/2}(s–1)^{1/2}D^{3/2}]$	0
q_w	unit water flux (discharge) per unit width	$\mathbf{L}^2\mathbf{T}^{-1}$
Q_s	volumetric sediment flux (discharge)	$\mathbf{L}^3\mathbf{T}^{-1}$
Q_w	volumetric water flux (discharge)	$\mathbf{L}^3\mathbf{T}^{-1}$
s	sediment specific gravity	0
S	downstream bed slope	
t	time	$\mathbf{T}$
T	representative time; cycle period	$\mathbf{T}$
T_{eq}	time required for system to reach equilibrium	$\mathbf{T}$
u	flow velocity	$\mathbf{LT}^{-1}$
U	representative velocity	$\mathbf{LT}^{-1}$
x	horizontal distance along the direction of mean sediment flux	$\mathbf{L}$
$\mathbf{x}$	two dimensional location (x, y)	($\mathbf{L}$, $\mathbf{L}$)
y	horizontal distance across the direction of mean sediment flux	$\mathbf{L}$
η	sediment surface elevation relative to a fixed datum	$\mathbf{L}$
θ	deep-water carbonate production parameter	$\mathbf{T}^{-1}$
ν	transport coefficient	$\mathbf{L}^2\mathbf{T}^{-1}$
ν'	modified transport coefficient. νC_0	$\mathbf{L}^2\mathbf{T}^{-1}$
ρ	fluid density	$\mathbf{ML}^{-3}$
σ	subsidence rate	$\mathbf{LT}^{-1}$
τ_0	boundary shear stress	$\mathbf{ML}^{-1}\mathbf{T}^{-2}$
τ_*	dimensionless boundary shear stress $\tau_0/[\rho g(s–1)D]$	
ϕ	logarithmic grain size measure, $-\log_2(D)$, D in mm	
ω	shallow-water carbonate production parameter	$\mathbf{T}^{-1}$
Modifiers		
0	value supplied at edge of depositional basin; bed value	
$\langle\rangle$	value averaged per unit basin width	