Assembling the stratigraphic record: depositional patterns and time-scales in an experimental alluvial basin

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ABSTRACT

Our understanding of sedimentation in alluvial basins is best for very short and very long time-scales (those of bedforms to bars and basinwide deposition, respectively). Between these end members, the intermediate time-scales of stratigraphic assembly are especially hard to constrain with field data. We address these ‘mesoscale’ fluvial dynamics with data from an experimental alluvial system in a basin with a subsiding floor. Observations of experimental deposition over a range of time-scales illustrate two important properties of alluvial systems. First, ephemeral flows are disproportionately important in basin filling. Lack of correlation between flow occupation and sedimentation indicates that channelized flows serve mainly as conduits for sediment, while most deposition occurs via short-lived unchannelized flow events. Second, there is a characteristic time required for individual depositional events to average to basin-scale stratigraphic patterns. This time can be scaled in terms of the time required for a single channel-depth of aggradation, and in this form is constant through a four-fold variation of experimental subsidence rate.

INTRODUCTION

The piecemeal construction of the rock record poses an enormous problem to stratigraphers. On the one hand, modern process studies allow us to observe sedimentation on time-scales from minutes to tens of years. On the other hand, rock sequences that span millions of years, of which only a small, unknown fraction are actually represented in the deposits, form the bulk of the stratigraphic record. The imbalance, by many orders of magnitude, in rates of deposition for these two end members indicates that, of the innumerable events (floods, storms, tidal cycles) that have occurred in any given location, only a few have left traces in the rocks (Sadler, 1981).

Rarely, if ever, does the requisite chronostratigraphic control exist in field studies to address this problem. Depositional dynamics over time-scales longer than those associated with individual features such as dune and bar cross-strata (typically of the order of 10^3 years or less) but below the resolution of the best dating techniques (typically of the order of 10^6 years or more) are especially difficult to constrain in the field. In many cases, these intermediate time-scales (10^3–10^5 years) are the domain of large-scale stochastic behaviour associated with avulsion and reorganization of the fluvial system. Loosely following usage in atmospheric sciences, we refer to fluvial dynamics in this time-scale range as ‘mesoscale’ dynamics. A heuristic definition of the mesoscale time range is that it lies between a ‘short’ time-scale on which individual channels or channel segments behave coherently and deterministically—the time-scale of most engineering models—and a ‘long’ time-scale on which autocyclic variability sums to produce the average behaviour represented in large-scale stratigraphic models (Paola, 2000).

Here we use experiments on a fluvial system in a subsiding-floor experimental basin to examine two basic issues of mesoscale fluvial dynamics. The first begins with the well-known idea that the stratigraphic record is a highly biased recorder. Certain events in a fluvial system are much more likely to be preserved in the sediments than others. Indeed, there may be little relation between a particular short-term flow configuration and the sedimentary package that is ultimately left behind. Interpreting the assembly of the stratigraphic record is like putting together a jigsaw puzzle in which the few dozen pieces that form the picture are mixed with thousands of apparently similar, but extraneous, pieces. As in natural basins, only a small fraction of the morphological events that occurred in the experiment are preserved in the

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stratigraphy. However, unlike natural basins, experiments allow us to observe deposition through a range of time-scales in addition to controlling the relevant boundary conditions (Peakall et al., 1996; Ashworth et al., 1999; Paola, 2000; Van Heijst, 2000; Paola et al., 2001).

The second issue pertains to depositional patterns across time-scales. Intuitively, the geometry of an alluvial depositional package produced over a short-time interval should reflect the morphology of the fluvial delivery system. However, sedimentation patterns over long time-scales are generally recognized to be a function of the interplay between sediment supply and accommodation space. Evidently, as time-scale increases, the depositional pattern shifts from reflecting the short-term flow pattern to reflecting long-term basinial accommodation. Individual depositional events are averaged to produce large-scale stratigraphic patterns. What is an appropriate scale for this averaging time? Borrowing a term from turbulence theory, we can think of this averaging time-scale as a kind of ‘integral scale’ for deposition. We will term it the stratigraphic integral scale. The detailed record that experiments provide is ideally suited to investigation of the stratigraphic integral scale, and overcome some of the limitations of field-based studies.

The Experimental EarthScape (XES) basin

The XES basin is a unique experimental facility located at the Saint Anthony Falls Laboratory (SAFL), University of Minnesota, that allows experimental formation of physical stratigraphy in a system with a flexible subsiding floor. The facility has been described in Paola (2000) and Paola et al. (2001), so only a brief summary will be presented here. Various groups have been performing experiments that preserve physical stratigraphy (e.g. Wood et al., 1993; Ashworth et al., 1994; Koss et al., 1994; Wood et al., 1994; Ashworth et al., 1999; Van Heijst, 2000; Moreton, this volume). The advantage of the XES basin is that its honeycomb arrangement of 432 subsidence cells allows a nearly unlimited range of spatial and temporal subsidence (or relative uplift) patterns, coupled to independent base-level control. In addition to subsidence rate and pattern, the basin is designed such that sediment and water discharges can be controlled independently.

At the beginning of an experiment, the basin is filled with well-sorted pea gravel that serves as the ‘basement’. Subsidence is produced by gradual removal of this layer of gravel from the bases of the hexagonal subsidence cells (Paola et al., 2001). A computer-activated pulse of pressurized water in any cell removes a small amount of gravel and produces approximately 120 µm of subsidence at the top of the gravel over the centre of that cell. The honeycomb cell pattern is not imprinted on the gravel surface, so the subsidence is spatially continuous. The pea-gravel basement layer is overlain by a thin latex membrane, onto which experimental sediment and water can be delivered from any point along the perimeter of the basin. Base level is controlled independently with a siphon attached to a movable constant head tank.

EXPERIMENTAL DESIGN

The experiment reported here (henceforth Run 99-1) was designed to investigate the influence of various proposed controls on the character of alluvial stratigraphy. Much of the design presented below was motivated by a seminal paper by Leeder (1978) which introduced a simple model for ‘alluvial architecture’—the stacking arrangement of fluvial sand bodies (channel belts) in avulsion-dominated alluvial basins. This theoretical model and its descendants (e.g. Allen, 1978; Bridge & Leeder, 1979; Alexander & Leeder, 1987; Leeder & Alexander, 1987; Mackey & Bridge, 1995; Peakall, 1995) are referred to as the ‘Leeder–Allen–Bridge’ (LAB) models. In particular, Run 99-1 focused on the roles of differential cross-stream subsidence, overall subsidence rate and the ratio of sediment discharge to water discharge. In order to isolate each of these signals, the experiment was divided into four stages, during each of which a particular tectonic or climatic scenario was imposed while all other experimental characteristics were held constant.

Run 99-1 used only one quarter of the full basin (108 cells; Fig. 1). Sediment and water were mixed outside of the tank and fed through four feed points spaced 37.5 cm apart along one of the short sides of the basin while a constant base level was maintained at the opposite end. The sediment was a mixture comprising 60% 120 µm quartz sand and 40% bimodal (460 and 190 µm) crushed anthracite coal. The coal has a specific gravity of 1.3 and thus is substantially lighter and more mobile than the quartz. The coal mixture was chosen both for its demonstrated grain-size segregation properties, and for its high optical contrast with the sand.

The sediment flux was matched to the accommodation rate throughout the experiment with the aim of maintaining a consistent mean surface elevation and a constant shoreline position. Under these conditions, the fluvial portion of the experiment averaged 4.1 m in length (source to shoreline) with an average slope of 0.050, varying somewhat with the stage of the experiment (Table 1). Water discharge was constant during each stage of the experiment.

In the first two stages of the experiment, contrasting subsidence patterns were imposed on the system. The 40-h-long first stage imposed a lateral (cross stream) subsidence variation on the fluvial system in the form of a streamwise trough running the length of the system, as well as a steep streamwise subsidence gradient located approximately half the distance from the source to the shoreline (Fig. 2). By contrast, the 30-h-long second stage imposed a simple downstream-tilting, rigid-beam subsidence pattern with no lateral subsidence gradient and a smooth subsidence gradient from source to shoreline (Fig. 2). The ratio of water to sediment discharge (qW/qS) was 44 throughout stage 1, whereas this ratio for stage 2

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was 39; the ratios differ slightly (10%) because of mechanical problems associated with a sediment feeder. This reduction produced no significant effect in the morphology of the fluvial system or in the resultant stratigraphy.

In stage 3, which lasted 99 h, the same rigid-beam style subsidence as in stage 2 was imposed on the system (Fig. 2), but the subsidence rate, accommodation rate, sediment flux and water discharge were reduced to 25% of their stage 2 values (Table 1). The purpose was to isolate the effect of a strong (four-fold) reduction in subsidence and discharge rates in the deposited stratigraphy. It is worth noting that had the subsidence rate been reduced while holding all other variables constant, the stratigraphic signature would have been dominated by the effect of progradation resulting from the imbalance between sediment supply and accommodation.

Finally, between stages 3 and 4 the subsidence pattern and rate were held constant while discharge conditions changed (Fig. 2). The water discharge was doubled while maintaining the sediment discharge at its stage 3 value. This stage contrasts with the preceding stages in that we imposed a potentially climate-related change rather than a tectonic change. Stage 4 lasted 49 h.

Through the course of the experiment, we periodically measured the topography of the fluvial surface by using a laser-based scanner (Rice et al., 1988; Wilson & Rice, 1990). These measurements were taken every 4 h during stages 1 and 2, and every 8 h during stages 3 and 4. The topographic scans had a resolution of 1 mm laterally, 60 mm longitudinally and 1 mm vertically, and covered an area 2.2 m wide by 4.88 m long co-centric with on the basin. In addition to the topography, we obtained photographs of the surface at regular intervals (1–10 min) in order to record more precisely the timing and nature of morphological events. Three video cameras—a continuous 8 mm camera and two time-lapsed VHS cameras—also documented the entire experiment, each with a slightly different perspective of the basin.

After completing the 218 h experiment, the deposit was drained, tried and then sectioned in order to observe and record the stratigraphy. The XES basin is equipped with a straightedge blade 0.2 m long suspended from a frame that constrains it to swing across a precise vertical plane. The sediment pile, even when damp, was unstable and prone to failure along normal faults parallel to the deposit face, but was stabilized by injecting a hot solution of commercial agar. Once stabilized, the sediment was sectioned vertically, perpendicular to mean flow direction, at 20 mm intervals. Owing to the stability problems, some portions were cut at slightly greater interval spacing and some slices are missing altogether. In all, 170 cross sections of the stratigraphy were cut, digitally imaged and reconstituted into 3D voxels in order to create a three-dimensional image cube from which dip-sections could be visualized.

**Table 1.** Experimental conditions by stage.

<table>
<thead>
<tr>
<th>Stage</th>
<th>Length (h)</th>
<th>$Q_s$ (L s$^{-1}$)</th>
<th>$Q_w$ (L s$^{-1}$)</th>
<th>$Q_s : Q_w$</th>
<th>Avg. slope</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>40</td>
<td>0.43</td>
<td>0.0097</td>
<td>44 : 1</td>
<td>0.048</td>
<td>Rapid, laterally asymmetric subsidence, cross-stream fault</td>
</tr>
<tr>
<td>2</td>
<td>30</td>
<td>0.53</td>
<td>0.0136</td>
<td>39 : 1</td>
<td>0.047</td>
<td>Rapid, laterally uniform, rigid beam subsidence</td>
</tr>
<tr>
<td>3</td>
<td>99</td>
<td>0.14</td>
<td>0.0031</td>
<td>45 : 1</td>
<td>0.060</td>
<td>Slow rigid beam subsidence</td>
</tr>
<tr>
<td>4</td>
<td>49</td>
<td>0.30</td>
<td>0.0032</td>
<td>94 : 1</td>
<td>0.060–0.050</td>
<td>Slow rigid beam subsidence, 2X water discharge</td>
</tr>
</tbody>
</table>
Fig. 2. Run 99–I subsidence patterns (left) and stage isopach maps (right). Flow is from right to left. Note that the system was allowed to adjust to an equilibrium slope before subsidence was imposed (initial condition).
In addition, we made 1 m by 2.5 m, plastic-backed peels of the sediment face at approximately 0.3 m intervals, thereby preserving the actual deposit. These ‘outcrops’ serve as important ground truth references in the analysis of the image data. We also obtained numerous detail images of specific deposit features.

We stress that this experiment was not intended as a ‘test’ of the LAB predictions in the strictest sense because many of the conditions assumed in the models (e.g. meander belts with clearly delineated floodplains) cannot yet be replicated at laboratory scales. Rather, the goal of the experiment was to determine whether any of the behaviour proposed in the LAB models could be detected in a generic physical system where all relevant boundary conditions are well known. We also stress that experiments such as this one are not intended as exact scale models of a particular natural prototype (Paola et al., 2001). Rather, following Hooke (1968) we view them simply as small depositional systems that have many of the same transport and depositional mechanisms as their larger counterparts. The applicability of observations in such small systems to the field usually cannot be unambiguously established through formal scaling analysis. Apart from comparing field and experimental observations where possible, the best way to evaluate the extent to which experimental observations do or do not pertain at field scales is to understand them mechanistically.

**EXPERIMENTAL RESULTS**

The unique data set produced in Run 99-1 will form the basis for several different lines of investigation, including analysis of how the stratigraphic record of the experiment was assembled; analysis of the dynamics of flow shifting and distribution (Cazanacchi et al., 2002); relation of individual deposit features to specific geomorphic events; comparison of the experimental results with the LAB model results; and development and testing of new theoretical models. The focus of this paper will be the first of these: how was the stratigraphy shown in a panel such as Fig. 3 assembled across the range of time-scales in the experiment—from fractions of a second to a few hundred hours? However, in order to provide a context for the discussion, it is worth briefly summarizing some main results of the experiment.

The sediment and water feed point configuration produced a distributive fluvial system analogous to a braided fan-delta. The position and lateral mobility of the channels were constrained by the location of the feed points in the most proximal portions of the system (<1.5 m from the source). In this region, the dominant mode of channel shifting was by gradual lateral migration. Beyond 1.5 m, however, the lateral flow distribution was not affected by the infed geometry, and channels tended to migrate by spilling over the banks creating a new channel, and abandoning the old channel. In general, the flow covered a relatively large fraction of the surface (30–40% in stages 1 and 2, 20–30% in stages 3 and 4). Channelized flow depths and widths ranged from 5 to 50 mm and 30–200 mm, respectively. Overflow from the braided channels produced sheet flows with depths <10 mm and widths ranging from 200 to 400 mm. The Run 99-1 fluvial system is discussed in more detail in Cazanacchi et al. (2002).

**The experimental deposit**

The deposit of Run 99-1 comprises two of the major elements of natural fluvial deposits: both channel-fill structures with scoured bases and depositionaly based, laterally extensive sheets. The channel-fill structures range in composition from all sand to all coal—that is, fines—and bear marked morphological similarity to field examples of individual channel fills (Fig. 4). Aspect ratios (width/thickness) of channel-fill structures cluster around 5 (similar to those of the fluvial channels themselves), in contrast to sheet deposits that have aspect ratios of 15 and greater. The sheet-like deposits are nearly always <10 mm in thickness, and are not the product of in-channel deposition.
Fig. 4. Comparison of Run 99-1 deposit features to field examples. (a) Coarse-grained (gravel) channel-fill structure from Canterbury Plains, New Zealand. (b) Coarse-grained (sand) channel-fill structure from XES basin. (c) Fine-grained (sand) channel structure from Canterbury Plains, New Zealand. Fieldbook is 20 cm tall. (d) Fine-grained (coal-rich) channel structure from XES basin.
The horizons separating the four experimental stages are recognizable in the stratigraphy at any distance from the source (Fig. 3). The boundary between stages 1 and 2 is evidenced by a transition from the laterally tilted strata of the former, to the flat-lying horizons of the latter, as would be expected from the contrasting subsidence patterns. In most of the deposit, the transition from stage 2 into stage 3 is a sharp boundary between relatively sandy and relatively coal-rich units. The irregular, but distinct, surface separating stages 3 and 4 is a transition from the relatively coal-rich strata of the former back into sandy strata of the latter. Further, owing to the two-fold increase in the ratio of water to sediment in the supply, the base of stage 4 is marked by erosive throughout most of the basin.

The first step in the analysis of the experimental deposit was to map the stratigraphic locations of channel-fill structures in the basin from the sediment peels. Using a digital photograph taken of the stratigraphy as a base map, each channel structure was outlined and documented. For each channel, the vertical and horizontal position, minimum brightness value (0–255), maximum brightness value, mean brightness value, cross-sectional area, width and interpreted experimental stage of deposition were recorded. The cross-sectional areas of the channel fills are log-normally distributed, considered both by experimental stage and through the entire section. The channel-fill statistics show no trend with subsidence rate or sediment supply, but the mean channel-fill cross-sectional area varies by stage as the square root of the water discharge (Fig. 5).

The channel-fill structures are identified on the basis of their erosive base and considerable thickness relative to the sheet deposits. In the proximal portion of the basin, where the fluvial channels were constrained by the incised

Fig. 5. Plots of cross-sectional area channel statistics. (a) Probability plots of cross-sectional area showing log-normal distribution of areas at 2.40 m and 3.40 m downstream from the source. (b) Plots showing the mean cross-sectional area by stage at the same downstream locations. Stage means are given as large dots. The mean area for all stages together is indicated by the horizontal line. Note that cross-sectional area varies with distance from source, as well as by stage.
geometry, channel-fill structures reflect the gradual lateral migration of the fluvial channels (Fig. 6). In more distal portions of the deposit, however, where the fluvial channels were increasingly free to change course abruptly, the channel-fill structures tend to be discrete bodies with dimensions comparable to those of the fluvial channels themselves.

**RELATING SURFACE PROCESSES TO STRATIGRAPHY**

The sequential overhead still photographs were used in concert with the continuous video record of the experiment to analyse the dynamics of the surface flow over timescales ranging from 1 s to 100 h. The result most relevant to the analysis presented here is that, though the flow eventually visits the entire surface if one waits long enough, the time required for flow occupation to approach uniformity is surprisingly long—a substantial fraction of the stage duration (Cazaclaci et al., 2002). We will return to this point below.

One of the great advantages of experimental work is the ability to relate active surface processes to their resultant deposits. In particular, the deposition of individual channel-fill and sheet-like bodies observed in the deposit can be related to the continuous overhead video. Such analysis is complicated by the dynamic nature of the fluvial system throughout the experiment—unambiguous chronostatigraphic horizons are rare. Indeed, study of experimental fluvial strata highlights how elusive ‘time surfaces’ really are in highly dynamic sedimentary systems. The experimental deposit is just an accumulation of fragmentary surfaces that cannot be precisely correlated over distances of more than (at best) a few channel widths. However, there are several easily identifiable events recorded in the deposit that can be used for tying the stratigraphy to the video.

As discussed above, the transition from stage 2 to stage 3 was everywhere associated with a transition from more proximal to more distal facies. This, together with the decrease in channel activity associated with the reduction in water and sediment supply, means that the topography of the fluvial surface at the end of stage 2 is well preserved in the deposit. From observations around this horizon as well as others, a detailed picture of surface processes responsible for specific channel-fill structures can be assembled.

Four fluvial processes dominate the depositional picture and the deposit is created through the interaction of these processes. The first process is erosional and involves the upstream migration of confluence scours. While this process is shown instantaneously in Fig. 7(a), it is best seen in the video record of the experiment, available at our website (http://www.geo.umn.edu/orgs/seds). These scours commonly form near a fluvial slope break at the sand to coal transition and move upstream through the system until they dissipate in the focused flow near the infed points, 0.5–1.0 m from the upstream end of the basin. These scours produce the concave upward, erosional channel-fill bases.

The second process is depositional and most important in the proximal portion of the basin where the infed geometry focuses the flow and restricts the movement of the fluvial channels. In this region, scoured channel bases are filled by gradual (~3 mm h⁻¹) lateral migration and lateral accretion in the channels. This process produces laterally extensive channel bodies that are encased in sheet-like deposits (Fig. 6).

The third of these processes, also depositional, is flow expansion (Fig. 7a). This occurs just downstream from one of the upstream migrating confluence scour points and the expanded flow is shallow (<10 mm) and wide (>200 mm). Such expansions often fill the trace of the scour itself, creating channel-fill structures that are composed of a relatively homogeneous mix of sand and coal. The specific sand content is related to composition of the underlying bed through which the scour erodes. In this case, the transport distance is so short (50–200 mm) that there is no sorting of the material. Such channel-fill structures, where entirely preserved, have similar dimensions to the fluvial channel responsible for their deposition (Fig. 7b). Whether flow expansions create channel-fill deposits or sheet-like deposits depends largely on the topography they encounter. Given relatively flat topography, the expanded flow deposits sediment as sheet-like deposits (Fig. 7d). However, should the flow encounter an antecedent channel, it often occupies this channel and continues downstream, where the scour/expansion sequence may begin again.
The fourth of the processes is associated with upstream avulsions and failed avulsions. The bankfull channelized flows in the upstream portion of the basin (1.0–2.5 m) tend to avulse by gradually overspilling their banks at a bend in the channel (cf. Leddy et al., 1993). The success of an avulsion—its tendency to form a new active channel with a course different from the pre-existing channel—depends largely on the topography it encounters. Avulsions are most successful when the overspilling flow finds a nearby abandoned channel, an effect well known from field studies (Aslan & Blum, 1999; Morozova & Smith, 2000). From a depositional standpoint, successful avulsions of this type actually do very little. Rather, they revert to the scour/expansion processes discussed above. Failed avulsions are much more effective at depositing sediment. Short-lived overspills that wane rapidly deposit whatever sediment they carried onto the topography they encountered. Thus they are analogous to crevasse-splay events in natural rivers, and in this experiment usually result in sheet-like deposits (Fig. 7d). However, where such failed avulsions spill into a preexisting channel scours, a wide range of channel-fill compositions—from pure sand to pure coal and nonhomogeneous mixtures (Fig. 7c)—are possible. The composition of such channel fills depends on their relative downstream position, and therefore sand content. Further, the degree of sorting in such channel-fill structures depends on how far from the over-spill point the deposition occurred.

Well-established channelized flow (< 200 mm wide, and > 30 mm deep) was relatively inefficient as a depositor of sediment in the more distal portions of this experiment. The flow often occupied a particular topographic channel many times before finally filling the space with sediment. Indeed well-channelized flow was observed to occupy the same channel as many as 10 times in 2 h with little apparent deposition. The final filling of channel topography is generally accomplished by the depositional processes, as described above, rather than by the gradual aggradation of bedload sediment.

The observation that channelized flows were inefficient depositors of sediment is of particular interest. In order to quantify this effect, we analysed the relationship between...
flow occupation, as calculated from the overhead still photographs of the experiment, and aggradation rate, as calculated from the laser-topographic scans through time. Using sets of overhead photographs taken 10 min apart during the first three stages, we assembled a statistical picture of flow occupation over 8 h intervals, the shortest interval for which there was a statistically significant number of photographs. Each of 48 photographs taken during a particular interval was binned into ‘wet’ and ‘nonwet’ pixels. If, for example, a particular place in the basin was ‘wet’ on 24 of the 48 pictures, that location was assigned an occupation value of 0.5, or 4 h, and so on (Fig. 8). While this analysis was not designed to distinguish between sheet-like flow and channelized flow on a particular photograph, the integration of data from photographs taken over 8 h identifies areas of the basin that were occupied by flow for a significant fraction of the time interval. In much of the experimental system, the dominant mode of channel switching was by the abandonment of an old channel and establishment of a new channel, rather than by lateral migration. As such, areas of high flow occupation tend to be areas of persistent channelized flow.

At each time during the experiment at which surface topographic data were collected, total sediment thickness was calculated by subtracting from the surface topography the basement topography that was recorded by the subsidence control computer. Further, by subtracting the total sediment thickness from the sediment thickness 8 h earlier, aggradation was calculated.

Maps of flow occupation and aggradation show little relationship with each other when compared visually (Fig. 9). Thus the amount of sediment deposited in any time interval seems to have little to do with the amount of time the point in question was wet over the same time interval. In order to investigate the relationship between the two data sets, we define the effective aggradation rate as the amount of deposition per hour wet. This was calculated by dividing the aggradation (vertical distance) at a particular location over the 8 h interval by the number of hours that location was occupied by flow over the same time interval. Mapping the effective aggradation highlights the relationship between flow occupation and aggradation.

By way of illustration, consider a situation where aggradation is directly proportional to flow occupation. In such a system, a persistent flow is the most effective process for depositing sediment and effective aggradation would then tend to be constant. Conversely, if aggradation were approximately independent of flow occupation, the effective aggradation would be inversely proportional to occupation. While it is clear that there is a limit to which an effect like this could persist—aggradation must eventually cease as flow occupation tends to zero—the relation between effective aggradation and flow occupation statistics serves as a measure of the influence of ephemeral flows in constructing the experimental stratigraphy.

It is clear from Fig. 10 that the effective aggradation calculated every 8 h through the experiment is inversely proportional to flow occupation; that is, high flow occupation corresponds to low effective aggradation. The relationship between aggradation and flow occupation is not simply inversely proportional because there must be a finite amount of flow at a particular place in the basin in order to produce any aggradation. As such, the inverse relationship breaks down for lower values of flow occupation. However, the inverse relation between occupation and effective aggradation for larger occupation values implies that relatively short-lived unchannelized flows are more efficient at depositing sediment than are persistent channelized flows.

Fig. 8. Flow occupation calculation method. If these three photos were a sample set, the upper box would be assigned an occupation value of 0.66 (‘wet’ in 2/3 photographs). The lower box would be assigned a value of 0.33.
Assembling the stratigraphic record

Fig. 9. Plan view maps of flow occupation and total aggradation during stage 1 between 8 h and 16 h runtime. Flow is from top to bottom. Note that there is no obvious correlation between occupation and aggradation.

Fig. 10. Plot of effective aggradation vs. flow occupation for 8 h intervals from stages 1, 2 and 3.

The clearest indication of this is the relatively high values of effective aggradation in areas of infrequent flow flanking the main channels (Fig. 11). This conforms with observations made from the continuous video coverage that well-channelized flow may occupy a particular route in the basin several times without filling it and that a short-lived failed avulsion or other flow-expansion event is necessary to actually fill in the topography.

ASSEMBLY OF THE STRATIGRAPHIC RECORD

One of the most basic observations about sedimentary basins is the general persistence of facies types over great vertical distances. For example, looking beyond short-term facies variability, passive-margin sequences often comprise several kilometres of strata all deposited within about 200 m of sea level. Observations like this imply that in many sedimentary basins, the variability that stratigraphers naturally focus on is played out within an arena in which, to first order, subsidence and sedimentation are nearly in balance. Yet, on short time-scales, subsidence cannot be a factor; rather, local flow processes shape the deposit.

Over a sufficiently long time interval, the fluvial system distributing sediment around the surface of a basin has time to visit every spot many times. If subsidence
Stage 1; 0–8 h runtime

Flow occupation

Effective aggradation

Stage 2; 52–60 h runtime

Flow occupation

Effective aggradation

Stage 3; 86–102 h runtime

Flow occupation

Effective aggradation

Fig. 11. Plan view maps of flow occupation (left) and effective aggradation (right) during stages 1–3. Flow is from left to right. Note that high values of flow occupation correlate with low values of effective aggradation, and that effective aggradation is particularly high in regions flanking areas of high flow occupation.

and sedimentation are in balance, the ratio of the mean aggradation to the mean accommodation at any point in the basin should approach unity. This balance develops over the 'long' time-scale referred to in the Introduction.

However, on short time intervals, the fluvial system distributes sediment according to its flow configuration during that time interval. Over most of the basin, the ratio of aggradation to accommodation will be greater.
than unity (relatively over filled) or less than unity (under filled), depending on the configuration of the fluvial system (Fig. 12). Thus the variance of the aggradation/accommodation ratio serves as a measure of the extent of subsidence control on sedimentation. In particular, the variance should approach zero on time intervals for which subsidence controls deposition. The time over which this occurs is proportional to the 'stratigraphic integral scale' referred to in the Introduction.

The ratio of aggradation to subsidence was calculated across the experimental basin on time intervals ranging from 4 h (the shortest interval between topographic scans) to 99 h (the length of stage 3, the longest period of constant experimental parameters). Each of the first three stages of the experiment was considered individually. As shown in Fig. 13, the variance decayed with increasing time interval. In the 40 h of stage 1, the 4 h standard deviation was approximately 0.3 and decayed down to approximately 0.05 through the stage. In the 30 h of stage 2, the 4 h value was closer to 0.2 and decayed to 0.05, a history similar to stage 1. Stage 3 started at 0.25 and decayed to approximately 0.075. While the rate at which the standard deviation decays is similar between stages 1 and 2, the rate of decay associated with stage 3 is much lower. In order to quantify this rate, an exponential function was fitted to the data from each stage, giving a characteristic time (the 'σ-folding' time—the time required for the standard deviation to decay to 1/e of its original value, analogous to a half-life). Not surprisingly, stages 1 and 2 have similar decay times: 20 h and 17 h, respectively. The decay time associated with stage 3, however, was nearly four times as long, approximately 61 h, corresponding to the four-fold reduction in overall subsidence rate imposed at the beginning of the stage (Table 2).

These decay times are of little meaning outside the context of this particular experiment. In order to relate such numbers to natural environments, we must nondimensionalize them by using physical parameters relevant to the processes that underlie the transition from flow-controlled to subsidence-controlled deposition. Consider a simple way of framing the problem: the fundamental requirement for deposition to respond to subsidence is that the flow must be able to free itself from a particular configuration and fill space produced by the externally imposed subsidence geometry. A simple means of liberating the flow from any particular channel is to fill the channel with sediment (e.g. Mohrig et al., 2000). If sedimentation in channels is the main impetus for flow redistribution, then a reasonable natural scale with which aggradation can be measured would be the scour depth. Comparing the total vertical accumulations given above for each stage with the mean scour depth from each stage.

Fig. 12. Schematic illustration of the development of an alluvial section (left) and a representation of the ratio of aggradation to subsidence (right). As the fluvial system has time to move across the basin repeatedly, the value of this ratio approaches unity.

Fig. 13. Plot showing the variance decay of the aggradation to subsidence ratio with increasing time interval. While all three stages have similar short-term variance values, the function decays much more slowly during stage 3 than in stages 1 or 2. The 'σ-folding' time associated with each curve is given in Table 2.

Table 2. Nondimensionalization of decay times.

<table>
<thead>
<tr>
<th>Stage</th>
<th>Decay time (h)</th>
<th>Avg. aggradation (mm)</th>
<th>Mean scour depth (mm)</th>
<th>Aggradation: scour depth</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>20</td>
<td>127</td>
<td>18</td>
<td>7.0</td>
</tr>
<tr>
<td>2</td>
<td>17</td>
<td>120</td>
<td>19</td>
<td>6.3</td>
</tr>
<tr>
<td>3</td>
<td>61</td>
<td>130</td>
<td>15</td>
<td>8.7</td>
</tr>
</tbody>
</table>

(measured directly from the sections), we find that the decay time for stage 1 corresponds to a mean accumulation of 7.0 scour depths; the decay time for stage 2 corresponds to a mean accumulation of 6.3 scour depths; and the decay time for stage 3 corresponds to a mean accumulation of 8.7 scour depths. (Table 2). These three dimensionless numbers are much closer to one another than the raw decay times are, suggesting that this form of nondimensionalization has captured at least a good part of the relevant mechanics. That the numbers still differ suggests that there may be more going on than is included in this simplified analysis.

DISCUSSION

Despite scale differences, many of the basic mechanisms that fill sedimentary basins are present in experimental and natural environments. One advantage of experimental work is that it allows us to study depositional mechanics systematically over various time-scales, which is difficult to do in natural environments. Owing to the scale differences, however, experimental records must be used judiciously. For example, the variance decay times given above make sense only in the context of the experimental system. If, as argued above, they are nondimensionalized with an appropriate length scale, the decay times can be compared with field cases.

The analysis presented here illustrates two features of mesoscale fluvial dynamics. The first point pertains to the role of ephemeral flows in basin filling. The effective aggradation statistic is a useful measure of this role. The mesoscale deposition and flow-occupation data from the experiment show that persistent channelized flow is, on average, relatively inefficient at depositing sediment. This is further supported by the short-term observation that a particular channel may be occupied many times before it is finally filled with sediment and thus preserved in the stratigraphy. The bulk of the deposition is accomplished by short-lived flows. Several factors contribute to the ultimate geometry and composition of the deposits left by such short-lived flows: where the flow splits from a main channel, the composition of the bed, transport distance, and antecedent topography.

Second, the analysis provides a means of estimating the stratigraphic integral time-scale, the upper limit of mesoscale dynamics. On short time-scales the fluvial system does not ‘feel’ the subsidence. Averaging of deposition over an interval proportional to the stratigraphic integral scale is required for the sedimentary package to resemble the subsidence pattern. In the experiment, this integral time-scale corresponds to the time needed to deposit several (5–10) channel depths worth of sediment at the average aggradation rate. Consider a field scenario: using a long-term aggradation rate of 1 mm year$^{-1}$ for a system whose average channel depth is 3 m, the depositional transition from flow control to subsidence control would occur on a time-scale of the order of 15,000–30,000 years. In the experiment, this stratigraphic integral scale stayed constant through a four-fold change in absolute subsidence rate. The measured time-scale is comparable to the time necessary for a fluvial system to occupy each area of a basin several times: over a vertical distance of 5–10 channel depths in the Run 99-1 deposit, several storeys of channel-fill structures are preserved across the width of the basin. As such, the stratigraphic integral scale might also depend on the spacing of the fluvial channels relative to their width in a particular system. In the experiment, the channel spacing is relatively small (roughly four channel widths), whereas in natural systems the spacing is generally much greater. Therefore, the time necessary for the fluvial system to occupy each place in the basin several times, even measured in terms of scour depth, might be relatively longer.

The disproportionate importance of short-lived ‘sediment dumps’ in fluvial deposition has been noted in several recent studies (Smith et al., 1989; Kraus, 1996; Aslan & Blum, 1999; Pérez-Arlucea & Smith, 1999; Morozova & Smith, 2000). Further, Smith et al. (1989) note the relatively high preservation potential of such events. It is striking that these field studies all pertain to large, low-gradient meandering river systems that could hardly be more different from the experimental system. This consistency of behaviour across such diverse fluvial systems suggests that deposition by short-lived flow-expansion events is a fundamental feature of fluvial environments.

CONCLUSIONS

1 Experiments are a useful tool for investigating questions of stratigraphic assembly. One of their major advantages is that they allow study of depositional mechanics across the spectrum of time-scales from individual events to the filling of a basin.

2 In the experiment reported here, the bulk of alluvial deposition is accomplished by short-lived flows, as indicated by poor correlation between flow occupation and net deposition. Established channels act largely as conduits for sediment, while overbank spills, flow expansions and failed avulsions all deposit a disproportionate amount of sediment. This is a phenomenon for which there is field as well as experimental evidence, suggesting that it is a generic feature of channelized flow systems.

3 There is a consistent scale that measures the time required to average individual depositional events into large scale stratigraphic patterns. We term this time-scale the ‘stratigraphic integral scale.’ In the experiment reported here,
the stratigraphic integral scale is equal to the time necessary for the deposition of several (5–10) scour depths worth of sediment at the average aggradation rate.

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