ABSTRACT: In an extended series of papers by M. Leeder, J.R.L. Allen, J.S. Bridge, J. Alexander, and others, the depositional stacking patterns of channel-belt deposits were related to tectonic and other external controls (we refer to these models collectively as LAB models). These models established a basic connection between subsidence, channel avulsion, and the mesoscale (channel-belt) architecture of alluvial deposit. Two fundamental predictions resulted from this work: (1) channel-belt stacking density and hence connectedness is inversely correlated to temporal (vertical) changes in sedimentation rate; and (2) channel-belt stacking density and hence connectedness is directly correlated to lateral changes in sedimentation rate. In this paper we make use of the Experimental EarthScape (XES) Facility at the St. Anthony Falls Laboratory to examine the effects of temporally and laterally variable subsidence rates on mesoscale alluvial architecture. We then compare the experimental results with the main predictions of the early LAB models. Regarding the predictions of the LAB models, under conditions of high sediment supply and a highly active alluvial system, lateral and downstream variation in subsidence geometry and rate have little effect on the details of alluvial architecture. We found that the principal architectural signature of changes in subsidence geometry (i.e., laterally variable subsidence rate) is stratal tilting and that channel belts are not attracted to the subsidence maximum. We hypothesize that the effect of variable subsidence geometry is felt by the fluvial system only if subsidence proceeds faster than the river can adjust to the formation of a topographic low via the deposition of overbank material in the form of splays and sheet sands. Our results lend impetus to the development of more complete three-dimensional numerical models of mesoscale alluvial architecture that couple architecture to broader, allogenic forcing mechanisms.

INTRODUCTION

The most influential line of work on how tectonic and other external controls are recorded in alluvial architecture began with the seminal paper by Leeder (1978). His insight was to relate the depositional stacking of channel-belt deposits to the interplay of avulsion and sedimentation. His analysis was continued and extended in a series of papers by J.R.L. Allen, J.S. Bridge, J. Alexander, and a number of their coworkers (Alexander and Leeder 1987; Allen 1978; Bridge and Leeder 1979; Mackey and Bridge 1995). For convenience, following Bryant et al. (1995) we will refer to the works in this series as the LAB models. The LAB models focus on the influence of external and internal forcing on so-called “alluvial architecture”: the organization of alluvial deposits in the subsurface, with emphasis on channel-belt deposits and their connectivity. If we divide alluvial stratigraphy into microscale (channel scale and below), mesoscale (channel belts), and macroscale (entire basin), then the LAB models focus on mesoscale architecture.

The power of Leeder’s (1978) original paper and the work that built on it was in establishing the fundamental connections among subsidence, avulsion, and mesoscale architecture. In particular, Leeder’s (1978) model remains one of the few stratigraphic models ever proposed that makes clear, testable predictions about the formation of mesoscale architecture. The two most important predictions of the early LAB models were (1) that channel-belt stacking density and hence connectedness is inversely correlated to temporal (vertical) changes in sedimentation rate (Leeder 1978); and (2) that channel-belt stacking density and hence connectedness is directly correlated to lateral changes in sedimentation rate (Alexander and Leeder 1987; Bridge and Leeder 1979). Reduction in subsidence rate with time increases the stacking density by, in effect, allowing channel belts more time to remove floodplain fines. Lateral increases in subsidence rate tend to draw channels laterally to the subsidence maximum and so cause them to stack more densely there. We will refer to these two predictions as P1 and P2, respectively.

The LAB models are physically based in that they rely on avulsion—an abrupt shift of channel-belt position that starts at some focused point on an active alluvial channel—to allow channel belts to migrate and redistribute sediment over the valley floor. In the original, two-dimensional models, avulsion rate was held constant or varied about a mean value. Bryant et al. (1995) found that avulsion rate increased with increasing aggradation rate when avulsions were mapped as nodes in the upstream 10% of the sediment surface of an axisymmetric experimental alluvial fan. These experiments led Heller and Paola (1996) to investigate the effects of differing relationships between avulsion frequency and aggradation rate on two-dimensional alluvial architecture. They found that, if avulsion frequency increased with increasing aggradation rate, they could generate precisely the opposite architecture predicted by P1: an increase in subsidence led to increased connectivity of alluvial channel sand bodies. Ashworth et al. (2004) showed that avulsion frequency was correlated more strongly with channel slope than with aggradation rate. They made a distinction between nodal avulsions in an alluvial fan setting like those examined by Bryant et al. (1995), where slopes are higher and avulsions may be effected by debris flows near the fan apex, and the lower-gradient braided rivers that were the focus of their study. These studies defined and measured avulsions in different ways, so their results are not strictly comparable, but Ashworth et al. (2004) argued that their results supported LAB model results, P1 in particular.

The LAB models were put forward with substantial qualification. Allen...
The Strengths and Limitations of the Experimental Approach

The research we describe here was done in the Experimental Earthscape basin (XES, also known as “Jurassic Tank”) at the St. Anthony Falls Laboratory, University of Minnesota. Experimentation is a natural complement to field and theoretical (numerical or otherwise) research in stratigraphy. The strength of experiments is that boundary conditions can be precisely controlled, surface conditions and topography monitored, and the resulting deposit sectioned to produce a detailed three-dimensional record (Ashworth et al. 1999; Paola 2000; Van Heijst and Postma 2001; Sheets et al. 2002). The primary limitations of stratigraphic experimentation involve (1) physical properties (e.g., grain size and fluid viscosity) that cannot be scaled down in any simple way, and (2) processes with intrinsic, irrevocable time scales, such as settling of fine sediments and many organic processes. Readers interested in further analysis of some of the dimensionless numbers and scaling issues in geomorphic and stratigraphic experiments are referred to the recent papers by Peakall and Warburton (1996) and Peakall et al. (1996).

In our view, there are two basic strategies in geomorphic or stratigraphic modeling: (1) an approach based on classical formal scaling. This approach has the advantage of allowing rigorous justification, but means restricting attention to coarse-grained prototype (field) systems with minor sand and finer fractions in transport. (2) An unscaled or partially scaled approach. In this view, the justification for stratigraphic experiments is not to be found in scaling based on the classical dimensionless numbers traditionally used in civil engineering model studies. Rather, the justification is a more heuristic one based on the ubiquity of scale-independent geometry (e.g., fractality) and dynamics in geomorphic systems (Foufoula-Georgiou and Sapozhnikov 1998, 2001; Rodriguez Iturbe and Rinaldo 1997; Sapozhnikov and Foufoula-Georgiou 1996, 1997) (Scale independence in natural systems is often referred to as “scaling;” this should not be confused with the formal scaling analysis referred to above.) Although the physical meaning of scale independence in nature is still not well understood, it does seem clear that the widespread occurrence of fractal geometry in geomorphic systems implies that there are many aspects of their dynamics that are not sensitive to scale. These aspects of geomorphic systems are susceptible to experimental study, regardless of whether the experiments are “scale models” or “analog models” of some particular prototype or not.

We have a good deal of information about the presence of scale independence in geomorphic systems (Rodriguez Iturbe and Rinaldo 1997), but the ubiquity of scale independence suggests that we would close ourselves off to a rich source of relatively accessible information and insight if we were to insist that geomorphic and stratigraphic experiments must be rigidly interpreted as classical scale models. At the same time, we simply do not understand how scaling works well enough to predict in detail how the dynamics of stratigraphic systems are distorted in scaling them down from field-scale to laboratory space and time scales. Given the present state of understanding, we adopt a pragmatic approach: following Hooke (1968), we view the experimental systems as nothing more than small systems in which we can fully control all the inputs and monitor the results. We try to understand how well they capture the dynamics of full-scale systems by comparing experimental results with field observations and models. In the end, modeling will generally be the key to scaling up experimental results: a physically based model should contain within itself the means for correcting for changes in scale; once tested under fully controlled conditions, the model can be used to evaluate the effects of processes distorted or left out of the experiments.

THE EXPERIMENT

The Experimental EarthScape (XES) Facility

The experiment we report on here was designed to examine the effects of changing subsidence rate and geometry on alluvial architecture. The XES is a small-scale sedimentary basin where sediment and water supplies, base level, and subsidence geometry and rate can be controlled independently. Subsidence is generated by withdrawing small amounts of pea gravel from the floor of the basin through a honeycomb of stainless steel fun-
nels. This gravel serves as a tectonic "basement," and is overlain by a rubber membrane that separates it from the experimental deposit. Subsidence can be controlled with submillimeter precision, and the gravel-extraction system generates extremely smooth subsidence geometries. The subsidence mechanism is described in detail in Paola et al. (2001) and Sheets et al. (2002). Water and sediment supplies are mixed and regulated externally, and can be fed into the basin from any point along its perimeter. Base level is controlled with a siphon mounted along the wall of the basin that controls water discharge into or out of a water body at the downstream end of the system.

Fig. 1.—Isopach maps of the experimental deposit at the end of stages 1, 2, and 3. General configuration of the Jurassic Tank run 99–1 shown on the isopach for the end of stage 1. Contours represent sediment thickness in millimeters and show the location of the cross-stream "fault zone" and the longitudinal trough offset from the basin centerline. The location of the Mylar peels used to characterize cross-stream channel-fill architecture are shown along the upper margin as distances downstream from the upstream edge of the tank (0.60, 0.90, etc.). Note that the edge of the isopach maps does not match the edges of the full tank coordinate system because the topographic scanner could not reach the extreme limits of the tank edges.

Experimental Design

The experimental basin used one quarter of the full basin, an area 2.98 m wide by 5.72 m long underlain by 108 subsidence cells (Fig. 1). Throughout the experiment we fed a sediment mixture comprising 60 percent 120 μm quartz sand and 40 percent bimodal (460 and 190 μm) anthracite coal sand through four feed points spaced 375 mm apart along one of the short walls (Fig. 1). Though the anthracite was coarser, it was transported more easily than the quartz because of a combination of protrusion effects and its lower density. Hence the crushed coal serves as a proxy for...
The experiment began with a period of no subsidence, during which the fluvial system was allowed to build an equilibrium profile. After equilibration, stage 1 began with a spatially complex subsidence pattern. In the streamwise (x) direction subsidence varied so as to simulate a cross-stream fault zone between 1.8 and 2.4 m downstream of the input of sediment and water (Fig. 1). Subsidence downstream of this fault was approximately two times faster than upstream of it, simulating a normal-fault zone with the hanging wall on the downstream side. In the y (cross-stream) direction the subsidence was programmed to produce a flow-parallel asymmetrical basin trough with a longitudinal axis offset 0.5 m to the right (looking downstream) of the centerline of the experimental basin (Fig. 1). The duration of stage 1 was forty hours.

The subsidence geometry throughout stage 2 resembled a hinged rigid plate, i.e., a simplified passive margin. Subsidence rate varied linearly from the upstream to downstream end of the basin. The theoretical hinge point for the stage 2 subsidence geometry was intentionally placed outside of the basin itself, leading to a situation where the subsidence rate at the feed points was approximately 50 percent of the rate at the shoreline.

The ratio of water discharge to sediment discharge ($Q_w/Q_s$) was 44 throughout stage 1, whereas this ratio for stage 2 was 39 (Table 1); the difference in the ratios (10%) was caused by mechanical problems in the sediment feeder. Variation in these ratios within either stage was negligible. The sediment discharge and water discharge were slightly higher during stage 2 because the laterally uniform subsidence geometry of this stage led to a somewhat higher total accommodation rate (Table 1).

The contrasting subsidence geometries of the first two stages were designed to isolate both the effect of laterally variable subsidence and a steep downstream gradient in subsidence rate on alluvial architecture. Neither geometry was intended as a replica of a particular natural environment. Rather, they were simplified scenarios designed to allow us to study the effect of spatial and temporal variation in subsidence on stacking patterns, and specifically to test predictions P1 and P2 given above.

Stage 3 also had a hinged-plate subsidence pattern, but the mean subsidence rate was one-quarter that of stages 1 and 2. Sediment and water supply were reduced accordingly so that the stable shoreline position of stages 1 and 2 could be maintained. Stage 3 was designed to isolate the effect of a four-fold reduction in subsidence rate on the architecture. It is worth noting that had we imposed a four-fold reduction in subsidence rate without concomitant reductions in sediment and water discharges the fluvial system would simply have prograded dramatically. The obvious effects of this progradation would have dominated the stratigraphy and made it difficult to isolate the effects of the reduction in sedimentation rate per se.

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**Table 1.**—Experimental conditions by stage for Jurassic Tank run 99-1.

<table>
<thead>
<tr>
<th>Stage</th>
<th>Duration (h)</th>
<th>$Q_s$ (l s$^{-1}$)</th>
<th>$Q_w$ (l s$^{-1}$)</th>
<th>$Q_w/Q_s$</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</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</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</td>
<td>0.060</td>
<td>Slow rigid beam subsidence</td>
</tr>
<tr>
<td>4</td>
<td>49</td>
<td>0.30</td>
<td>0.00032</td>
<td>94</td>
<td>0.060-0.050</td>
<td>Slow rigid beam subsidence, 2 $\times$ water discharge</td>
</tr>
</tbody>
</table>

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**Fig. 2.**—Changes in sediment surface mean gradient throughout stage 3. Note that within 8 hours after the end of stage 2 the slope of the system had steepened considerably. Following this initial steepening, the system prograded basinward.
Fig. 3.—Results of a preliminary analysis of avulsion frequency and location based on mapping avulsions on segments of overhead video during Jurassic Tank run 99-1. A) Avulsion frequency (gray bars) compared to aggradation rate (black line with symbols) for all stages. B) Avulsion frequency (gray bars) compared to fluvial system slope for all stages. C) Map of avulsion locations during stage 1. D) Map of avulsion locations during stage 2.

The experimental deposit was sectioned in serial vertical slices, oriented normal to the flow direction, spaced at 2 cm intervals. Because of problems with the stability of the deposit, some portions were sliced at slightly greater interval spacing and some slices are missing altogether. In all, 170 cross sections of the stratigraphy were cut and digitally imaged. In addition, we made 1 m by 2.5 m Mylar-backed peels of the sediment face at approximately 30 cm intervals, thereby preserving the actual deposit. These “experimental outcrops” serve as important ground truth references in the analysis of the image data.

Overhead video of the experiment allowed us to analyze surface processes throughout much of the run. We examined the location and frequency of avulsions for one- to two-hour portions of each stage. The initial portions of stage 1 lacked video coverage; as a result, only two video segments from this stage were analyzed. We used four or five, roughly equally spaced video segments to characterize avulsions for each subsequent stage. For this analysis, an avulsion was mapped and counted each time a new channelized flow appeared in the sandy area of the system and persisted for at least 10 seconds of experimental time. This definition is similar, but not identical to, that used by Bryant et al. (1995), yet differs from the methodology of Ashworth et al. (2004). The results of these avulsion analyses must be taken with caution because they represent a preliminary analysis only and were not the main focus of this research, yet they lend some insight into the behavior of the fluvial surface throughout the experimental run.

STRATIGRAPHY OF THE EXPERIMENTAL DEPOSIT

A description of the geomorphology and surface processes associated with run 99-1 have been described elsewhere (Cazanacli 2000; Cazanacli et al. 2002; Sheets et al. 2002). The experimental fluvial system was a very active, laterally unstable braided network. This system debouched into the water body at the downstream end, creating a fan delta. For the most part the delta foresets and the distal topsets were composed entirely of coal. The flow tended to occupy channels 10–15 cm wide or form thin sheet-floods over bar-like surfaces. Channels shifted frequently, mainly through processes of local and large-scale avulsion and gradual lateral migration.

The mean slope of the deposit surface remained constant throughout stages 1 and 2 (Table 1). The large reduction in sediment and water supply associated with the onset of stage 3, however, was accompanied by a 30% increase in average surface slope (Sheets et al. 2002) (Fig. 2). This added slope was built by upstream deposition, which induced sediment starvation near the shoreline, leading to an abrupt transgression at the beginning of stage 3. The stratigraphic effects of the transgression, and a means of accounting for them, are presented in Strong et al. (in press). Mean slope
profiles viewed through time show that the sediment surface prograded basinward throughout stage 3 until the shoreline re-stabilized at a position comparable to that of stages 1 and 2 (Fig. 2).

We found that there was a relatively poor correlation between aggradation rate and avulsion frequency (Fig. 3). Stages one and two had substantially higher aggradation rates than stages three and four, yet avulsion frequencies were comparable across all stages, with perhaps a slight overall decline in the lower aggradation rate stages. However, we found that avulsion rate did appear to be influenced more strongly by surface slope (Fig. 3), a result confirming those of Ashworth et al. (2004). In particular, avulsion rate and surface slope tracked together throughout stages 3 and 4, where we encountered the greatest slope changes.

**Alluvial Architecture in Strike Section**

The alluvial architecture expressed in the experimental deposit shares many features with alluvial outcrop exposures (Miall and Tyler 1991). The deposit comprises well-defined channel belt and unchannelized (sheetflow) facies (Fig. 4). Individual and amalgamated channel fills show concave-up bases and a variety of fill types (Sheets et al. 2002). Unchannelized facies are characterized by sheet-like alternations of sand and coal, forming tabular bodies (Fig. 4).

Laser topographic scans, in conjunction with precise measurements of basement topography, allowed us to superimpose timelines on the stratigraphy (Fig. 5). The transition from stage 1 to stage 2 is coincident with the transition from laterally tilted to horizontal strata (differential lateral subsidence to laterally uniform subsidence; Fig 5). The timelines in stage 1 diverge toward the flow-parallel trough, reflecting the higher rate of subsidence in this zone. The strata diverge in a similar fashion, with sheet and channel-belt deposits inclined toward the subsidence maximum. In all other stages the timelines are subparallel, following the hinge-like subsidence pattern. In many respects, the most distinctive difference between stages 1 and 2 is the change from a tilted to horizontal stratigraphic geometry.

Sheets et al. (2002) analyzed patterns in the architecture of the experimental deposit, exploring the statistics of the channel-belt sand-body dimensions and internal fills at two cross sections, \( x = 2.40 \) and \( x = 3.40 \) m. Upstream of 2.40 m, individual channel fills were difficult to discern because of extensive amalgamation. The cross-sectional areas of the channel fills at each cross section are log-normally distributed. Stages 1 and 2 have comparable channel fill dimensions, whereas stages 3 and 4 are characterized by significantly smaller channel fills. There were no channel fills in stage 3 at 3.40 m, because of the transgression described above. The stage-by-stage mean cross-sectional channel fill area is proportional to the square root of water discharge (Sheets et al. 2002).

**Cross-Stream Variation in Channel-fill Density**

To characterize cross-stream variation in channel fill density at various downstream locations, we selected eight cross-section images for mapping of detailed channel fills (Fig. 1). We chose these sections because they corresponded to locations of Mylar peels, which could be used to check the results of image mapping. Each image was edge-enhanced to emphasize crosscutting channel-fill margins, and to detect the well-stratified overbank areas amid the more chaotic areas of amalgamated channel-fills. Using these images, we produced a binary classification map of channel-belt and non-channel-belt areas (Fig. 6). The percentage of the cross section classified as channel fills was calculated over a 20-cm-wide moving window (about two channel-belt widths), allowing us to analyze the variation in...
channel-fill density across the section on a stage-by-stage basis (Fig. 7). This was accomplished by creating sub-images of each binary map corresponding to each stage of the experiment at that downstream position. These sub-images were then broken down into vertical swaths 200 pixels (20 cm) wide, and the percentage of channel-fill and non-channel-fill area was calculated for each swath.

**Downstream Variation in Alluvial Architecture**

The close (2 cm) spacing of vertical slices allowed us to create a three-dimensional, voxel image of the deposit that could be viewed as downstream-oriented (flow-parallel) slices (Fig. 8). In the immediate vicinity of the sediment and water feed points the architecture was dominated by a zone of extensive channel-fill amalgamation (Fig. 8). Lateral adjacent to the feed points—on the left and right sides of the tank—these zones of amalgamated channel-fills gave way to sheet sands, characterized by diffuse coal layers and abundant sand (Fig. 8, orange zone). A wide zone of amalgamated channel sand bodies occupied the medial portions of the deposit, just downstream from the feed points (Fig. 8, green zone), where channelized flow spread over the entire width of the tank. Downstream of this zone was an area of variable width showing isolated channel sand bodies that appear to have the geometry of distal sand ribbons (Fig. 8, blue zone). The transition from these ribbon sands to pure coal roughly approximates the shoreline position throughout the experiment, which can be seen to shift over a range 10–20 cm wide over stages 1 and 2 (Fig. 8).

As mentioned above, the transition from stage 2 to stage 3 was accompanied by a marked increase in gradient of the fluvial system (Fig. 2), resulting in a transgression at the stage boundary. Figure 8 shows the pronounced stratal effects of this transgression. Throughout stages 1 and 2 the transition between zones of amalgamated channel fill sand bodies and areas of distal sand ribbons remained relatively fixed at a location just downstream from the cross-stream fault (Fig. 8). This transition (and all other “facies” boundaries) migrate rapidly upstream immediately at the onset of the third stage. The transgression is followed by a readjustment of fluvial slopes and progradation of the sand-rich facies, with the shoreline reestablishing itself in a position slightly upstream of that of stages 1 and 2.

Downstream variation in channel-fill density was calculated using a technique analogous to that outlined above to compute the density of channel fills in a cross-stream direction. In this case, however, the channel-fill fraction was calculated for the entire sub-image on a stage-by-stage basis (Fig. 9). For each stage, channel-fill density is low in upstream portions of the deposit, because of focusing of the flow by the geometry of the sediment and water feed. These initially low values are followed by an increase in the medial area (from 1 to 2 m downstream from the upstream tank wall) where channels can freely migrate laterally. Subsequently, these values decline to zero between 2.5 and 3.5 m as the size and frequency of confluence scours decreases, the main process responsible for the formation of channel-shaped sand bodies (Fig. 9). Stages 1 and 2 have almost identical downstream patterns, with stage 2 showing a slightly higher channel-fill fraction (~ 0.7) when compared to stage 1 (~ 0.6). Stage 3 never exceeds a channel-fill fraction of 0.5.

**DISCUSSION**

We begin by summarizing the two predictions of early LAB modeling that we set out to explore: P1 states that temporally decreasing sedimen-
Effects of Changes in Subsidence Rate (Comparisons between Stages 2 and 3)

We begin with P1, for which the relevant comparison is between stages 2 and 3, which used the same subsidence pattern but differed by a factor of four in accommodation rate and supply rates of water and sediment. Comparison of Figures 7 and 9 indicates that, taken at face value, the experimental results seem opposite to those predicted: the channel-fill stacking density is lower for the reduced subsidence rates of stage 3, rather than higher.

What does this result imply? We begin by considering the way in which we imposed the reduced subsidence rate in stage 3. A naïve approach to testing for the effect of subsidence rate on channel-fill stacking geometry would have been to adjust only the subsidence rate and hold sediment and water supply constant. The immediate and dominant effect of this would have been progradation of the fan delta and consequent migration of facies basinward. As a result, vertical changes in the alluvial architecture would have been caused by progradation induced by the imbalance of sediment supply and accommodation, not by the change in subsidence rate per se.

Our experiment was designed to avoid this and determine whether changes in subsidence rate alone could influence the stacking density, maintaining the relative positions of key facies boundaries like the shoreline. To do this, we maintained the same ratio of water discharge to sediment discharge between stages 2 and 3 but reduced the absolute values of these quantities for stage 3. The increase in fluvial slope and associated transgression at the onset of stage 3 implies that we were not wholly successful (Fig. 2). Evidently, the reduction in transport capacity for the reduced water supply $Q_w$ was greater than linear and thus more than offset the reduced sediment supply $Q_s$. A steeper slope was needed, and this was produced by impounding sediment at the upstream end of the system and starving the shoreline. This steepening and transgression caused a major change in architectural style from the large channel fills of stage 2 to mostly fines and sheet sands at distal cross sections in stage 3 (Figs. 8, 9). The vertical succession in stage 3 coarsens upward, a record of the gradual adjustment of the fluvial system to a new, steeper slope. This adjustment is characterized by progradation (Fig. 8). Despite our best effort to avoid superimposing stacking-density changes driven by facies migration, that is precisely what occurred in experimental stages 2 and 3.

We have analyzed the effect of the transgression and associated facies migration in Strong et al. (in press). In that paper we propose to correct for the upstream shift in the depocenter during the transgression by recasting the alluvial architecture in terms of a new dimensionless variable that measures mass extraction down the depositional system. Applied to stages 2 and 3 of this experiment, we find that the architecture is similar at locations in the basin that have equivalent values of this dimensionless mass extraction variable. In other words, the stratal signature is similar where comparable volumes of sediment have been extracted from the system via deposition.

In our view, the most important result of this part of the experiment is not so much an explicit test of P1 as simply highlighting how difficult it is to analyze changes in stacking density in a two-dimensional framework,
FIG. 7.—Cross-stream variation in the percentage of channel fill over a 20-cm-wide moving window. The centerline of the basin lies at 150 cm in all cases. The location of each cross section is shown in the upper left corner of each diagram.
divorcing them from facies migration driven by allogetic factors (sediment supply, base level, subsidence, etc.). Even in a relatively simple experiment, with full control over the input conditions, we were not able to impose a major reduction in subsidence rate without inducing significant facies migration, with its concomitant effects on alluvial architecture up and down the system. It would clearly be far more extraordinary for this to occur in nature. In view of this, it simply does not make sense to expect there to be any consistent relation between subsidence rate and architecture in a
single vertical panel. In the experiment, local architecture was influenced much more strongly by facies migration, which in turn is controlled by variation in the overall sediment mass balance along the system, than by local effects of subsidence rate per se. Had we done the experiment by simply reducing the subsidence rate as described above, the spatial pattern of channel fill stacking density that we described above suggests that we would have seen what P1 predicts (an increase in stacking density)—but because of downstream migration of facies with high channel-fill density rather than a local change in the relative effectiveness of removal of fines by channel migration. This set of observations lends strong support to recent attempts to develop models of alluvial architecture in three dimensions (Mackey and Bridge 1995). It should also lend urgency to the next step, which is to couple architecture modeling with overall fluvial-system modeling including the effects of the standard allogenic drivers on the distribution of deposition within the system (Paola 2000).

Effects of Cross-Stream Variation in Subsidence

Next we turn to P2, which predicts that channel-belt sand bodies within a sedimentary basin subject to lateral variation in subsidence rate will be concentrated over subsidence maxima. Stages 1 and 2 of the experiment described here were designed to highlight the effects of laterally variable subsidence. If channels are attracted to a lateral subsidence maximum, stage 1, with asymmetric subsidence, should show a concentration of channel fills over the longitudinal trough, whereas stage 2 would be expected to have no such strong concentration of channel deposits. The trace of the trough was deliberately offset from the basin centerline in the expectation that there would be a natural tendency for channel fills to concentrate near the middle of the basin. Figure 7 suggests that the longitudinal trough had no effect on the concentration of channel fills across the basin. For each cross section, the distributions of channel-fill fraction for stages 1 and 2 are very similar, suggesting that there is no strong difference between the alluvial architectures in these stages (Fig. 7). For both stages, the highest concentrations of channel fills occur near the center of the basin (Fig. 7), not over the subsidence maximum. This is also seen in the statistics of channel location in the surface-flow data, which do not indicate any tendency for preferred occupation of the lateral subsidence maximum (Cazanacli 2000; Cazanacli et al. 2002). Furthermore, the location of avulsion nodes based on our overhead video mapping show no strong tendency for avulsions to be concentrated over the subsidence maximum of stage 2 (Fig. 3).

We have come to understand this behavior as a result of the relatively high water and sediment discharges, and a relatively low ratio between them (40:1, respectively). The effect of these conditions was highly mobile channels, which tended to visit every point in the basin relatively frequently. In calculations based on the overhead images taken during the experiment, Cazanacli et al. (2002) developed a harmonic function $f_d(t)$ that expresses the fraction of the fluvial surface remaining dry (unvisited by flow) at time, $t$:

$$f_d(t) = \frac{f_{dry} e}{1 + e^{-t/t_{rel}}}$$

where $f_{dry}$ is the initial dry fraction (1-wet fraction) and $t_{rel}$ a characteristic decay time for the remaining dry fraction. After a time equal to $t_{rel}$, the dry fraction was reduced to half, after $2t_{rel}$ a third, and so on. The value of $t_{rel}$ for stage 2 (where conditions were similar to those in stage 1) of the experiment was 376 s, meaning that after 1 hour, 95% of the surface had been occupied by flow, and therefore had been graded to the fluvial slope. This measure provides an analogous, but more complete, view of the “activity” of the fluvial system than does avulsion frequency, the measure applied in all LAB models, particularly given the ambiguities surrounding the definition of an avulsion in these experimental systems (Ashworth et al. 2004).

In order for fluvial channels to be attracted to a lateral subsidence maximum, cross-stream slope associated with tectonics must exceed the fluvial slope. During stage 1 of this experiment, 10 hours of subsidence was necessary to create sufficient cross-stream slope.

Thus we interpret our results as the product of a channel system in which the time scale associated with channel activity is much shorter than that on which differential topography can develop. Had the opposite been true, as would be the case for systems with relatively slow lateral migration rates and small wetted fractions, fluvial channels would have been steered toward developing, underfed topographic lows. More formal description of these timescales is the subject of ongoing research.

Insofar as sedimentation rate is related to the activity of the fluvial system, our result was anticipated by Alexander and Leeder (1987) who noted that “stream[s] can be deflected from [a subsidence maximum] if the sedimentation rate more than compensates for the subsidence.” They also noted that if subsidence rate fell below some sedimentation rate-dependent threshold, channel positions would be determined by local variations in topography at the mesoscale and microscale. The total thickness of deposit, however, would be thickest over the lateral subsidence maximum.

Several field tests have explored the validity of P2, either explicitly or implicitly. All are based on Holocene or late Quaternary systems and show a tendency for channels to be strongly influenced by the formation of structural depressions in full-graben and half-graben settings, thereby lending support to P2. Studies in the Rio Grande Rift (Leeder et al. 1996a) have reasonably good chronostratigraphic control but essentially lack three-dimensional exposure. Analyses of the Rhine–Meuse delta (Törnqvist 1993, 1994; Törnqvist et al. 1993; Berendsen and Stouthamer 2002; Cohen et al. 2002) have excellent temporal control and show lateral and longitudinal variations in architecture in some detail, yet they address only the upper one or two channel stories; these studies also lack three-dimensional control. Finally, studies of the effects of lateral tilting on the Carson River, Nevada (Peakall 1998) have very precise temporal control, yet the effects on alluvial architecture must be completely inferred. Two of these cases, the Rio Grande Rift (Leeder et al. 1996a) and the Carson River (Peakall 1998), lie in arid to semiarid settings that are likely to be limited in terms of sediment supply. As a result, avulsions may be insufficiently frequent and the “activity” of the channel belts may be reduced. This prevents channels from “visiting” the entire floodplain surface between successive
Effects of Downstream Variation in Subsidence

The imposition of a cross-stream fault zone on the stage 1 subsidence geometry had no effect on the location of avulsion nodes (Fig. 3) or on the alluvial channel-fill stacking density (Fig. 9). Curves of channel-fill fraction are almost identical between stages 1 and 2, although no cross-stream fault was present in the latter stage. In particular, the decrease in channel-fill density through the fault zone is comparable between stages, and is due to the overall decrease in channel-fill density distally, as discussed above. To explain this, we make an argument similar to that made above for variation in cross-stream subsidence geometry: sediment supply overwhelmed any effect of the fault, effectively eliminating a tectonic signal in terms of channel-fill fractional area.

Downstream variations in channel-fill stacking geometry occurred mainly as a result of variation in channel position over the sediment surface. Figure 10 shows maps of the proportion of time that the sediment surface was wet during stages 1 and 2; the fan-like geometry of the system is clearly shown. Channels tended to occupy, and therefore wet, the proximal portions of the deposit far more than other areas. In the proximal portions of the sediment surface, channels tended to be concentrated very near the feed points. Laterally away from these feed points the deposit was dominated by sheet sands, leading to a low channel-fill fraction in this region of the deposit (Figs. 8, 9). Downstream of this area, in the medial parts of

Fig. 10.—Maps depicting the proportion of time that the sediment surface was wet averaged over the entire stage for stages 1 and 2. Gray values correspond to the proportion of time that the surface was wet for each stage.
the system, channels shifted freely over the sediment surface, wetting it periodically, but regularly (Fig. 10). As a result, areas of amalgamated channel fills shown as the green zone in Figure 8 developed in this portion of the deposit. As sand was extracted from the system as the flow moved downstream, and as channels became narrower and less competent, the surface was occupied less frequently by large channels (Fig. 10). Channelized sand bodies diminished in size in the distal portions of the system, leading to the deposition of distal ribbon and sheet sand bodies (Fig. 8).

CONCLUSIONS

In our experiment, we found that the principal architectural signature of changes in subsidence geometry is stratigraphic (Fig. 5). Regarding the prediction that subsidence rate is inversely proportional to channel-fill stacking density (P1), our results suggest that the two-dimensional variation in alluvial architecture is controlled very strongly by externally forced facies migrations such as changes in sediment supply, base level, or subsidence. Any change in subsidence rate, in our experiment, leads to upstream or downstream facies migration and thus becomes the dominant control on the two-dimensional architecture. Local changes in subsidence rate constitute a second-order control on channel-fill stacking patterns.

With respect to P2, under conditions of high sediment supply, lateral and downstream variation in subsidence geometry has little effect on the details of the architecture of alluvial channel fills. We hypothesize that the effects of variable subsidence geometry will be felt by the fluvial system only if subsidence precedes faster than the river can adjust to the formation of a topographic low via the deposition of overbank material in the form of spilays and sand sheets.

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