Chapter 3

Fluvial response to sea-level changes: a quantitative, analogue experimental approach

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Abstract

Quantitative relationships of fluvial response to allocyclic change are crucial for further progress in understanding the stratigraphic record in terms of processes that have dominant control on landscape evolution. For instance, without quantitative insight into the time lag that is known to exist between a fall in relative sea level and the fluvial response, there is no way to relate fluvial stratigraphy to sea level. It is difficult to put firm constraints on these time-lag relationships on the basis of empirical studies. Therefore we have quantified time-averaged erosion and deposition in the fluvial and offshore realms in response to sea-level change by means of analogue modelling in 4 x 8 m flume-tank model. Sea level was the only independent variable, while other conditions like sediment supply, discharge, and initial geometry were kept constant over 18 experiments.

The experimental results support the idea that neither fall nor rise in sea level does affect the upstream fluvial system instantaneously. An important cause for the delayed fluvial response is that a certain amount of time is required to connect initial incisions on the just emerged shelf (shelf canyons) with the fluvial valley. Base-profile lowering in the fluvial system starts only after the connection of an active shelf canyon with the fluvial valley; until that moment the profile remains steady. We quantified the process of connection through introducing the quantity “connection rate”. The connection rate has a strong bearing on fluvial and shelfal stratigraphy, since it controls: 1) the amount of fluvial aggradation during the sea-level fall; 2) the total sediment volume that bypasses the shelf edge; 3) the percentage of fluvial relative to shelf sediment in the lowstand delta; 4) the volume of the transgressive systems tract and 5) the amount of diachronicity along the sequence boundary. The experiments demonstrate that the sequence-stratigraphic concept is difficult to apply to continental successions, even when these successions have been deposited within the reach of the influence of sea level.

(Submitted for publication in Basin Research)
Introduction

The original sequence-stratigraphic concept that relates basin-margin architecture to eustatic sea-level changes (Vail et al., 1977) was primarily tested for shelf-slope settings and did not include the fluvial system. However, an increasing number of high-resolution shelf studies (e.g. Suter & Berryhill, 1985; Bartek et al., 1990; Coleman & Roberts, 1990) have led to the understanding that rivers do play an important role in sediment delivery to the shelf, especially during lowstands. Posamentier & Vail (1988) included some basic constraints on fluvial response to eustatic fluctuations within their sequence-stratigraphic framework. The recognition that the fluvial response to sea-level changes would have significant implications for sediment delivery and depositional geometries on the shelf and slope (Butcher, 1990; Wescott, 1993) explains the increasing interest into the timing of shelfal and adjacent fluvial deposition in relation to the sea-level curve (i.e., attributing systems tracts to alluvial deposits, e.g. Shanley & McCabe, 1991). The special issue on incised valley systems (Zaitlin et al., 1994) expressed a strong demand for a consistent sequence-stratigraphic concept for the fluvial domain. Several years later, Ethridge et al. (1998) make a strong case not to define systems tracts for the eustatically unaffected, and tectonically and climatically controlled upstream fluvial reach. The problem to which point in the fluvial system the sequence-stratigraphic concepts can still be applied or not is a complex one, so that the question remains how much of the concept of sequence stratigraphy can be applied to continental strata that are beyond the direct influence of sea level (Shanley & McCabe, 1994). The heart of the problem probably lies in our limited understanding of the processes that control base-profile adjustments. The base profile is usually defined as the ideal graded profile at a specific moment relative to a chrono-stratigraphic datum (Quirk, 1996). A graded river (e.g. Mackin, 1948) represents morphological stability (Ethridge et al., 1998), where the sediment delivery to the coast equals the supply from the drainage basin. The base profile grades towards base level, which is equal to sea level in coastal regions, although we recognise that rivers can locally erode below sea level (Salter, 1993; Schumm, 1993; Best & Ashworth, 1997). In any case, the position of the base profile is very dynamic: while the river responds to a sea-level change, the base profile changes continuously (Quirk, 1996).

Base-profile adjustment forced by a fall in sea level will start at the coastline and move progressively landward by headward erosion resulting in knickpoint migration (Salter, 1993; Leeder & Stewart, 1996; Quirk, 1996). The adjustment of the valley profile’s gradient proceeds by migration of one single knickpoint or by an array of local knickpoints (Gardner, 1983). The point of intersection of the old and new base profile, i.e., the most landward knickpoint, would put an upward limit on the stratigraphic effect of the sea-level fall (cf. Posamentier & Allen, 1993, their fig. 6). Hence, a sea-level fall will not immediately rejuvenate the entire river profile (Leopold & Bull, 1979; Schumm & Ethridge, 1994). Also, slow rates of sea-level fall can be accommodated by a change of channel sinuosity without causing an instantaneous change of the valley-floor gradient (Schumm, 1993).
How far upstream will a sea-level fall be noticed? Paola (1991) regards the fluvial system as a low pass filter for sea-level fluctuations and defines propagation distance of base-level-fall-induced erosion as being proportional to the square root of the period of variation (e.g. a 100 kyr base-level cycle will affect approximately 100 km of fluvial profile). It depends on several factors, how far upstream a river will be rejuvenated. Among these factors, the magnitude and rate of sea-level fall, the river gradient and the supply rate from the catchment area are the most important (Schumm, 1993). The effect of a sea-level fall on fluvial stratigraphy fades upstream in favour of climate and tectonic influence, as well as autocyclic changes such as sediment flux variations and changes in fluvial discharge in the upstream direction (Posamentier & James, 1993). A drop in sea level is felt only several kilometres upstream for small, high-gradient rivers, whereas large, low-gradient rivers with larger drainage basins seem to adjust their profiles 100’s of kilometres upstream. For instance, in response to the last glaciation the small Obitsu River incised about 15 km upstream (Saito, 1995), the Colorado River nearly 100 km (Blum, 1993), and the Mississippi 300 km (Saucier, 1996) and possibly even up to 1000 km (Fisk, 1944); see Table 3.1.

The morphologic concept of river-profile adjustment by headward erosion thus implies a time lag between the onset of a sea-level fall and the upstream adjustment of the river’s base profile (Butcher, 1990). Our understanding of the time lag is poor (Shanley & McCabe, 1994; Quirk, 1996; Dalrymple et al., 1998). The time lag may cause that erosional and depositional cycles in the coastal zone are out-of-phase with the sea-level cycles (Ethrige et al., 1998), which illustrates a major difficulty in the application of sequence-stratigraphic concepts to fluvial strata. Application of the concept is also much hindered by the problem of convergence (Schumm, 1991): different allocyclic (climate, tectonics, eustasy) causes and different processes can produce similar results (stratigraphy). For example, any fall in base profile will cause an erosive surface, irrespective whether this bounding surface relates to a fall in sea-level, tectonism or climate change affecting the ratio of discharge over sediment load in the river (Shanley & McCabe, 1994; Quirk, 1996). Reviews on hydrodynamics (Thorne, 1994) and field studies (e.g. Blum & Price, 1998; Törnqvist, 1998) refer to this complex response of the fluvial system. So, even if the coastal and inland fluvial systems seem to be in-phase, it may well be because of different controls involved.

<table>
<thead>
<tr>
<th>River</th>
<th>Drainage basin (km²)</th>
<th>river gradient</th>
<th>shelf gradient</th>
<th>Lowstand river extension (km)</th>
<th>Upstream limit of influence from last glacial sea-level lowstand with respect to the present shoreline (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Obitsu River</td>
<td>274</td>
<td>0.002</td>
<td>0.006</td>
<td>30</td>
<td>15 (Saito, 1995)</td>
</tr>
<tr>
<td>Hawkesbury River</td>
<td>22000</td>
<td>0.0005</td>
<td>0.06</td>
<td>40</td>
<td>140 (Nichol et al., 1997)</td>
</tr>
<tr>
<td>Colorado (TX)</td>
<td>110000</td>
<td>0.0004</td>
<td>0.0008</td>
<td>100</td>
<td>90 (Blum &amp; Valastro, 1994)</td>
</tr>
<tr>
<td>Brazos</td>
<td>118000</td>
<td>0.0002</td>
<td>0.0003</td>
<td>100</td>
<td>70 (Anderson et al., 1996)</td>
</tr>
<tr>
<td>Mississippi</td>
<td>3344000</td>
<td>0.00002</td>
<td>0.00025</td>
<td>150</td>
<td>300–400 (Saucier, 1996) 1000 (Fisk, 1944)</td>
</tr>
</tbody>
</table>
(Ethridge et al., 1998). Hence, the complexity of the fluvial system does not allow an easy quantification of stratigraphy-controlling parameters through the examination of real-world examples. It probably explains the relative underdevelopment of sequence-stratigraphic models for the alluvial domain with respect to their marine counterparts (Shanley & McCabe, 1998). Only through numerical and physical experimental studies we may be able to resolve some of fluvial complexity by carefully testing the impact of each parameter on stratigraphy (Ethridge et al., 1998; Marriott, 1999; Blum & Törnqvist, 2000).

We use an analogue flume model to investigate the response of both the fluvial and the shelfal domain to various rates of sea-level change. The advantage of the analogue approach is that, in contrast to theoretical models (e.g. Burgess & Allen, 1996), the process of knickpoint migration is intrinsically embedded in the flume experiments. Sea-level change is the isolated variable in our study, while initial topography, discharge, sediment supply and tectonic subsidence were held constant. The methodology has been inspired by the analogue experiments of Wood et al. (1993) and Koss et al. (1994), who produced mainly qualitative results. By making use of high-resolution surface mapping techniques we generate quantitative data on rates of erosion, deposition and knickpoint migration for a wide range of experimental sea-level curves.

### Table 3.2. Facts on the set-up and the experimental method.

<table>
<thead>
<tr>
<th>Experimental set-up</th>
<th>Properties</th>
</tr>
</thead>
</table>
| Dimensions          | Main tank: 4 x 4 x 1 m, table with shelf-slope configuration 3x3.4 m  
                      | Fluvial valley, Duct: 4 x 0.5 x 0.11 m |
| Co-ordinate system  | x, y and z axes with values in mm (Fig. 3.1) |
| Measurements        | Main tank: automated bed profiler, accuracy of x, y and z data within 0.4 mm.  
                      | Applied data point spacing 20 mm.  
                      | Fluvial valley: manual stream profile measurement with rulers spaced 100 mm  
                      | apart (accuracy 2 mm). |
| Discharge           | 400 dm$^3$/h |
| Sediment supply     | 1 dm$^3$/h (~1.85 kg of dry sediment per hour) |
| Sediment properties | Unimodal medium sand used as uniform substrate (bed material) and as supply  
                      | for the fluvial valley.  
                      | D$_{50}$: median grain diameter = 250 µm  
                      | D$_{90}$: ninety percentile grain diameter = 700 µm  
                      | And 40 µm < D < 1000 µm to avoid cohesion problems with clays and to  
                      | exclude partitional sorting effects of large grains. |
| Hydraulic conditions| h = 6-10 mm  (channel depth)  
                      | U = 0.18 m/s  (average flow velocity in the fluvial valley)  
                      | Fr = 0.77  (Froude number in the fluvial valley)  
                      | Re = 886 (Reynolds number in the fluvial valley) |
| Sea level (single variable) | Highstand: z=420 mm  
                      | Lowstand: z=340 mm or z=260 mm (depending on amplitude) |
| Length of stream profile (from head to shoreline) | Highstand: 4.7 ± 0.2 m (variation induced by highstand delta size)  
                      | Lowstand: 6.0 ± 0.2 m (variation induced by amplitude and lowstand delta size) |
Fig. 3.1—(a) Experimental set-up consisting of a water and sediment filled basin margin (main tank) and a fluvial valley (rectangular duct with the sediment feeder at its up-slope end). An automatic bed profiler (laser) measures the topography of the sedimentary basin. (b) Schematic plan view and (c) cross-section of the experimental set-up showing the x, y and z-axis of the bed profiler that is used as the co-ordinate system.
Methodology

Experiment facility

The set-up consists of an experimental tank of 4 x 4 x 1 m that is connected with a rectangular duct (the fluvial valley) of 4 x 0.11 x 0.5 m (Fig. 3.1, Table 3.2). The tank contains a sediment table with sidewalls that support a sand sheet, which forms the coastal plain, shelf, slope and basin configuration (Table 3.3). A water tap with flow meter provides discharge. A sediment feeder with adjustable conveyor-belt speed controls the sediment supply rate. Both are located at the upstream end of the fluvial valley and act as a surrogate for the drainage basin. The applied sediment is uniform, medium sand (Table 3.2) that is supplied by the feeder and is used as substrate. An adjustable level of overflow controls the water level (sea level) in the main tank. An automatic positioning system, with x and y-axes is attached to the ceiling above the main tank. It carries a Dynavision SPR-2 laser sensor to collect altitude data (z-axis) of the coastal plain-shelf-slope-basin topography. The data are measured according to a 20 x 20-mm grid and has an accuracy of 0.4 mm for all three dimensions. Changes of the fluvial valley’s stream profile are measured by means of rulers attached to the valleys glass wall at 10-cm spacing.

Table 3.3. Initial morphology of experiments after the 15 hour preparation run.

<table>
<thead>
<tr>
<th>Initial model morphology</th>
<th>Inclination ((\Delta y/\Delta x))</th>
<th>y-co-ordinates ((\text{mm}))</th>
<th>z-co-ordinates ((\text{mm}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basin</td>
<td>0</td>
<td>0-300</td>
<td>0-10</td>
</tr>
<tr>
<td>Slope &amp; shelf edge</td>
<td>0.42</td>
<td>300-1200</td>
<td>10-385</td>
</tr>
<tr>
<td>Shelf</td>
<td>0.03</td>
<td>1200-2400</td>
<td>385-420</td>
</tr>
<tr>
<td>Coastal plain</td>
<td>0.02</td>
<td>2400-3400</td>
<td>420-440</td>
</tr>
<tr>
<td>Fluvial Valley</td>
<td>0.025</td>
<td>3400-7000</td>
<td>440-530</td>
</tr>
</tbody>
</table>

Scaling real-world to experimental time-space dimensions

The relative dimensions of fluvial-valley length versus shelf width and vertical exaggeration is scaled following Hook’s (1968) similarity of process approach. This is the only possible way to model landscape evolution, since prototype dimensions are too large to keep up with conventional scale models (Bruun, 1966) and Froude scaling (e.g. Ashworth et al., 1994). We added to the qualitative aspects of analogue modelling by quantifying the time-averaged sediment flux \((Q_s)\) from river to basin over sufficiently long time spans (i.e., graded time of Schumm & Lichty, 1965). However, such quantitative treatment of analogue model results demands an alternative scaling strategy as proposed in Chapter 2. Here we briefly summarise the relevant scaling aspects for analogue experimental studies of fluvial response to sea-level change.

The dimensions in our flume model are designed to represent a common conceptual Quaternary passive margin setting. We have chosen a shelf gradient that is just steeper than the equilibrium profile of the downstream reach of the fluvial valley (see Table 3.3), since most recent natural fluvial-shelf systems do so (e.g. Miall, 1991; Nummedal et al., 1993, see Table 3.1). All gradients of the model are scaled...
proportionally to the gradient of the stable equilibrium profile of the fluvial valley, which in turn depended on the bed-load transport of the applied sediment and discharge (Table 3.2). The equilibrium profile of the model’s fluvial valley (duct) is 0.025, which is 10-100 times steeper than that of natural rivers (Table 3.1). It is important to note that the coastal plain is chosen less steep than the fluvial and shelf profile which is common for rivers in a passive margin setting (Butcher, 1990; Nummedal et al., 1993). The water-level variations are designed to model glacio-eustatic cycles composed of a slow fall that forces the shoreline below the shelfbreak (cf. Talling, 1998) followed by a rapid rise.

Scaling by similarity of process means that hydraulic scaling conditions are relaxed. However, we kept realistic Froude numbers in the fluvial valley (lower flow regime, see Table 3.2) to avoid bedform formation and, yet, to ensure a constant bed-load transport rate. The fraction smaller than 40 $\mu$m was removed to avoid unwanted effects caused by cohesion. All grains larger than 1000 $\mu$m were sieved out to avoid any large ratio of particle size over water depth that can lead to partial sorting and bed armouring.

**Equilibrium time**

In order to investigate the time-lag relationships between sea-level change and fluvial response systematically, the response time of the sedimentary system must be taken into account. The response time compares to the equilibrium time $T_{eq}$ that was defined by Paola et al. (1992). They stressed the importance of the ratio between the period of change of a variable and the system’s equilibrium time and we agree! For a proper time scaling, we maintain similar values for the ratio of the equilibrium time, $T_{eq}$ over the duration of one cycle of sea-level change, $T$ in both model and prototype by defining a Basin Response factor ($Br$):

$$Br = \frac{T_{eq}(w)}{T_{eq}(rw)} = \frac{T_{exp}}{T_{eq}(exp)}$$

(3.1)

To apply above scaling condition we need to establish the equilibrium time of both model and real-world river-shelf systems. The observed equilibrium time is defined as the time that is needed for the fluvial system to regain its equilibrium base profile from the moment that it is disturbed by sea-level or discharge changes. The establishment of an initial equilibrium profile at the beginning of each experiment took 6-10 hours keeping discharge and supply at default values (Table 3.2) and with sea level at highstand.

A few empirical formulae exist to estimate the equilibrium time ($T_{eq}$) for sedimentary systems. We applied three examples to verify our model’s $T_{eq}$. The predicted values are displayed in Table 3.4. The significance of a response time for a graded river to a downstream change in base level was already recognised by De Vries (1975) who proposed:

$$T_{eq50} = \frac{3 \cdot b \cdot S \cdot (\frac{1}{2} L)^2}{\alpha \cdot Q_s} [T] \quad \text{if } L > 3h/S (= \text{length of backwater curve})$$

(3.2)
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$T_{eq50}$ represents the time required to accommodate 50% of the base-level change and $L$ is the length of the river affected by the change. $S$ is the bed slope; $b$ and $h$ are river width and depth respectively. $Q_s$ is the volumetric bulk sediment transport for the river segment under consideration. We applied the sediment transport formula of Engelmund & Hansen (1967) and found a best fit for $\alpha = 4.6$ for a data set of calibration experiments of the 4 m fluvial valley (see Appendix). Substitution of $\alpha$ in the De Vries (1975) equation yields equilibrium times between 7 and 16 hours, so a little longer than the observed values.

In a review on equilibrium time scales in geomorphology, Howard (1982) proposed:

$$T_{eq} = \frac{L^2 \cdot S}{4 \cdot q_s}$$  \hspace{1cm} \text{[T]} \hspace{1cm} (3.3)

Where $q_s$ is the sediment transport rate per unit width. This formula estimates long equilibrium times (10-20 hours) relative to our observations.

Paola et al. (1992) determines the equilibrium time for basins to reach equilibrium by:

$$T_{eq} = \frac{L^2}{k}$$  \hspace{1cm} \text{[T]} \hspace{1cm} (3.4)

The diffusivity constant $k$ was derived by Paola et al. (1992) from first physical principles, using the bed-load transport formula of Meyer-Peter & Müller (1948) to describe the volumetric sediment transport rate:

$$k = \frac{8q_w \cdot A \cdot \sqrt{C_f}}{C_o (s - 1)}$$  \hspace{1cm} \text{[L$^2$/T]} \hspace{1cm} (3.5)

Where $q_w$ is the discharge per unit width ($m^2/s$) and $A$ is a constant for riverbank stability. $C_f$ is the dimensionless drag coefficient, $C_o$ the sediment concentration of the bed and $s$ the specific density ($\rho_b/\rho_w$). According to the equation, diffusivity depends mainly on discharge and the constant $A$ for riverbank stability. Parker (1978) found a riverbank stability constant $A = 1$ for the meandering case and $A = 0.15$ for the braided case. Similarly we can assume $A = 1$ for our fluvial valley that is contained between glass walls. The average value of $A$ must be lower than unity downstream the confined fluvial valley, because the stream diverges owing to decrease in bank stability. Therefore, we applied an average value of $A=0.8$, which yields values for $T_{eq}$ that compare well with the observations (Table 3.4). Hence, we find a diffusivity of $1.17q_w$ for the experiments. Based on empirical data Paola et al. (1992) used $0.1q_w$, which furthermore suggest that the real-time values for sediment transport rates in our flume diverge about a factor ten from these real-world values.

All formulae predict $T_{eq}$ within the range of observations for the 4 m fluvial valley length (Table 3.4). However, for lowstand conditions ($L \sim 6m$) the estimates lie further apart. During the experiments we observed that the gradient and the actual
sediment-supply rate increased as an experiment progressed towards lowstand. This is not accounted for in the calculations. Therefore, the calculated values for $T_{eq}$ might be overestimated for $L > 5\text{m}$ conditions (Table 3.4).

Based on observations and above calculations we assume the model’s $T_{eq} \approx 10$ hours and have set out a suite of experiments that explores the fluvial response to sea-level changes systematically for $0.4 < Br < 4$ (Fig. 3.2). The $Br > 1$ condition (i.e., $T > T_{eq}$) that we have chosen for the majority of our experiments (Table 3.5) compares nicely with moderately-sized Quaternary river systems issuing on 100 km wide shelf margins during the Quaternary.

### Table 3.4. Estimates of the equilibrium time, $T_{eq}$ of the model. Note that $L$ depends on the shoreline position and progradation.

<table>
<thead>
<tr>
<th>$L$, Length (m)</th>
<th>$T_{eq}$ (h)</th>
<th>$T_{eq}$ (h)</th>
<th>$T_{eq}$ (h)</th>
<th>$T_{eq}$ (h)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(De Vries, 1975)</td>
<td>(Howard, 1982)</td>
<td>(Paola et al., 1992)</td>
<td>Observed</td>
</tr>
<tr>
<td>4 (~fluvial valley)</td>
<td>7</td>
<td>11</td>
<td>5</td>
<td>6-10</td>
</tr>
<tr>
<td>5</td>
<td>11</td>
<td>17</td>
<td>7</td>
<td>~10</td>
</tr>
<tr>
<td>6 (~shelf exposed)</td>
<td>16</td>
<td>24</td>
<td>10</td>
<td>~12</td>
</tr>
</tbody>
</table>

*Applied values: channel width: $b=0.11\text{ m}$, $h=0.0056\text{ m}$, $S=0.025$, $Q_w=0.4 \text{ m}^3/\text{h}$, $Q_s=0.001 \text{ m}^3/\text{h}$, $C_f=0.027$, $C_o=0.68$, and $\rho=1000 \text{ kg/m}^3$ and $\rho_s=2650 \text{ kg/m}^3$.*

### Experimental procedure

All experiments started with an identical coastal plain-shelf-slope topography (Table 3.3). The sediment in the fluvial valley was levelled according to an inclined marker line on the valley wall prior to each experiment. The sidewalls of the table in the main tank were used as standard levels to assure similar sediment-surface height before the start of each experiment. The sand bed was submerged two times to improve its packing. The water level was raised to highstand level at the beginning of each experiment. A 15-hour preparation run with sea level held at highstand level and sediment supply and discharge at their default values preceded each experiment in order to establish a stable fluvial profile as uniform starting condition. Thus, we started each experiment with a graded stream where the sediment supply rate at the valley outlet is constant and equals the constant supply rate of the feeder at its upstream end. Discharge and rate of sediment supply were monitored throughout the experiments. Table 3.5 gives an overview of the 18 experiments. Sea-level variations consisted of 160 or 80 mm amplitude sinusoidal curves (Fig. 3.2). Four sea-level curves have been repeated two or three times to test the reproducibility of the analogue-model results.

The subsequent morphological changes were monitored, photographed and recorded on time-lapse video. The fluvial stream profile was measured every hour and twice hourly for short experiments. The topography of the entire sand bed in the main tank was scanned every 5 hours, and more frequently for very short experiments (Table 3.5). Laser scans were done subaerially. Therefore, prior to each scan the tank was drained slowly to avoid disturbances in the grain fabric. Water level, discharge and sediment supply were checked before the experiment was resumed.
Fig. 3.2—Applied sinusoidal sea-level curves with 80 mm and 160 mm amplitude. The labels indicate experiment numbers corresponding to Table 3.5. The majority of the experimental sea-level cycles took two to four times longer than the model’s equilibrium time of approximately ten hours.
Table 3.5. Overview of the experiments.

<table>
<thead>
<tr>
<th>Experiment number (Fig. 3.2)</th>
<th>Sea-level amplitude (mm)</th>
<th>Duration of sea-level fall (h)</th>
<th>Rate of sea-level fall (mm/h)</th>
<th>Duration of sea-level rise (h)</th>
<th>Rate of sea-level rise (mm/h)</th>
<th>Time between scans (h)</th>
</tr>
</thead>
<tbody>
<tr>
<td>90</td>
<td>160</td>
<td>0.1</td>
<td>1000</td>
<td>0</td>
<td>-</td>
<td>5</td>
</tr>
<tr>
<td>70</td>
<td>160</td>
<td>2</td>
<td>80</td>
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<td>140*</td>
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<td>30</td>
<td>5.33</td>
<td>10</td>
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<tr>
<td>170</td>
<td>160</td>
<td>20</td>
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<td>10</td>
<td>16</td>
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<td>180</td>
<td>80</td>
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<td>4</td>
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</tr>
<tr>
<td>190</td>
<td>80</td>
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<tr>
<td>200</td>
<td>80</td>
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<td>210</td>
<td>80</td>
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<tr>
<td>220</td>
<td>80</td>
<td>30</td>
<td>2.67</td>
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* Fluvial valley width is 220 mm instead of the 110 mm during other experiments

Results

In each experiment, a stable fluvial profile was established between the 6th and 10th hour of the 15 hour preparation run. The equilibrium slope of the fluvial valley varied between 0.024 and 0.026. Meanwhile, a highstand delta developed on the coastal plain. Although the highstand delta deposits are very thin (~ 1 cm), they caused the coastal plain to be less steep than the fluvial valley and the shelf (Table 3.3). This section focuses on the results of two extremes of the 80 mm amplitude experiments; a fast rate of sea-level fall and a very slow rate of sea-level fall relative to the model’s equilibrium time. The comparison is based on two experiments for each extreme, both showing good mutual reproducibility. After the comparison, the quantitative data of the full range of experiments are presented.

>>> The following pages show the results of a comparison between an experiment with a fast rate of sea-level fall and one with a slow rate of sea-level fall.

Figures 3.3 and 3.4 show block diagrams that depict the evolution of the coastal plain and shelf. Figure 3.5 shows a medial cross-section of experiment 210 with a fast rate of sea-level fall. The stages a to d correspond to Fig. 3.3. The successive diagrams illustrate the balance between the deposition of newly introduced sediment from the feeder and redeposited sediment from cannibalisation of the shelf and fluvial valley. The successive profiles show the delay in upstream propagation of the headward erosion in the fluvial domain. The knickpoint reaches the fluvial valley at lowstand (10 h) and the sea-level-fall-induced erosion continues until late rise.

Figure 3.6 shows a medial cross-section of experiment 250 with a slow sea-level fall. The stages a to d correspond to Fig. 3.4. The experiment shows significant fluvial aggradation during stages a and b. Erosion on the shelf and within the fluvial valley resulting from a slow sea-level fall is much more in phase than for the fast-fall experiment (cf. Fig. 3.5). The knickpoint reaches the fluvial valley at early lowstand (18 h) and fluvial erosion ceases at early rise.
Fig. 3.3—Block diagrams illustrating coastal plain and shelf evolution during experiment 210 with a fast sea-level fall. The subsequent scans depict current topography as well as changes (contours) with respect to the previous topography for 5 hour time steps. The first block diagram shows the topography at t=5 hours and the volume changes that occurred between t=0 and t=5 hours. Erosion (red) and deposition (green) have been plotted with 5-mm contour intervals.

>>> next page: Fig. 3.4—Similar block diagrams illustrating coastal plain and shelf evolution during experiment 250 with a slow sea-level fall.
Experiment 250

**Slow fall**

- **i.** 0-5 hours
  - shore erosion
  - slow contour interval

- **ii.** 5-10 hours
  - shelf edge delta progradation
  - shallow distributaries deposit small lobes

- **iii.** 10-15 hours
  - shelf valley not connected with the fluvial valley yet
  - small valley ceases to exist

- **iv.** 15-20 hours
  - knickpoint of largest valley retreats the fastest
  - lowstand delta progradation

- **v.** 20-25 hours
  - knickpoint enters fluvial valley at early lowstand
  - shelf erosion and bypass of fluvial sediment
  - abandoned valleys

- **vi.** 25-30 hours
  - upstream erosion in fluvial valley
  - bypass of fluvial load and valley widening

- **vii.** 30-35 hours
  - transgressive backstepping lobes fill the shelf valley (TST)
  - aggradational deposits (HST)

- **viii.** 35-40 hours
  - aggradational deposits (HST)
  - highstand shoreline
Fig. 3.5—Medial cross-sections (x=1500 mm) illustrating experiment 210 with a fast rate of sea-level fall. The sea-level curve is divided into stages a to d that correspond to Fig. 3.3.
Fig. 3.6—Medial cross-sections illustrating experiment 250 with a slow rate of sea-level fall. The stages a to d correspond to Fig. 3.4
Shelfal response to sea-level changes

Figures 3.3 and 3.4 show the topographic changes on the coastal plain-shelf-slope setting for Experiment 210 (fast fall) and 250 (slow fall). Both experiments show many similarities in shelf evolution in response to one full cycle of sea-level fall and rise. As the shoreline drops over the outer shelf and shelf edge, small canyons develop along that part of the lower shelf that receives the discharge. Multiple, small, V-shaped canyons simultaneously incise the outer shelf and deposit small shelf-edge deltas (Figs 3.3a and 3.4a). The shelf canyons compete for the available discharge from the fluvial valley that is distributed through stream avulsion on the apex of the delta. Canyon cutting proceeds by knickpoints migrating up the shelf. The valley that captures most of the discharge has the most proximal knickpoint i.e., the upstream limb of the red coloured central shelf valley in Fig. 3.3b. By cutting off smaller valleys, this valley starts monopolising all available discharge as its knickpoint reaches the middle shelf. The dominating valley progressively deposits the largest lowstand delta while the smaller systems become starved of discharge and sediment. Finally, the dominant valley connects with the fluvial valley and becomes a single, shelf bypass valley that feeds one large, lowstand delta (Figs 3.3c and 3.4v). During the subsequent sea-level rise, which is of similar duration for both groups of experiments, lowstand delta progradation shifts to aggradation. As the rate of rise increases towards the rise inflection point, the main shelf-bypass valley becomes flooded and filled by small backstepping lobes (Figs 3.3c and 3.4c). During late transgression, the upper part of the incised valley is backfilled.

Fluvial response to sea-level changes

Although the drainage development on the shelf is similar for both the slow and the fast fall in sea level, the timing at which the main valley erodes into fluvial strata differs significantly. The differences in timing of deposition and erosion on the coastal plain and fluvial domain for both cases are illustrated by their medial cross sections in Figs 3.5 and 3.6 for experiments 210 and 250, respectively. Both experiments start with a graded fluvial profile in equilibrium with a highstand delta on the coastal plain. Near the fall inflection point, and earlier with increasing rate of fall, the stream incises into the highstand delta. The incision is forming the apex for a new delta lobe that aggrades and progrades at a lower level on the shelf. Simultaneously, aggradation continues in the fluvial system (cf. Figs 3.5a and 3.6a). The volume of fluvial aggradation, however, is largest for the slow fall (Fig. 3.6a) because of the longer time span available. The knickpoint positions on the exposed shelf are still fairly similar for both cases (cf. Figs 3.5a and 3.6a). At lowstand, the dominating shelf canyon has not connected with the fluvial valley, as yet, in case of the fast sea-level fall. In contrast, connection occurs before lowstand during the slow-fall experiment (cf. Figs 3.3b and 4v). Remarkably, the early transgression of the shelf is still accompanied by upstream erosion in the fluvial valley resulting in extra sediment delivery to the shelf edge in both experiments. Erosion in the fluvial domain continues until the rise inflection point during the slow sea-level fall experiment (Fig. 3.6d), and until late rise during the fast sea-level fall experiment (Fig. 3.5d). Finally, the fluvial valley re-establishes its initial equilibrium slope towards the new highstand shoreline before highstand in both examples.
Effect of sea-level amplitude

The above results show that the connection time, the time at which the dominant shelf valley connects with the fluvial valley, differs for the fast-fall and slow-fall experiment. A star below the sea-level curves in Fig. 3.7a indicates the connection time (i.e., the first appearance of the knickpoint in the fluvial valley). The knickpoint paths over the shelf are shown in Fig. 3.7b and illustrate that the knickpoint of the dominant shelf canyon migrated much faster upstream during the fast fall than the slow-fall experiment. Figure 3.7c shows the rate of deposition on the shelf (deposition rate of the green patches in Figs 3.3 and 3.4).

Fig. 3.7—Quantitative comparison of two fast and two slow sea-level fall experiments with 80 mm amplitude. (a) Sea-level curve. (b) Knickpoint migration paths show that shelf valleys connect at or just after lowstand for the fast-fall experiments while the slow-fall experiments show connection already at early lowstand as indicated by the stars below the sea-level cycle. (c) The measured mean rate of deposition downstream of the fluvial valley shows the effect of a sea-level change on the sediment flux to the coastal plain, shelf and lowstand delta. (Fluxes were calculated from the green depositional volumes on the scans).
Similar trends in knickpoint migration and the rate of deposition on the shelf were found for experiments with large amplitude (Fig. 3.8). Obviously, the 160 mm amplitude experiments connected sooner than the 80 mm amplitude experiments (cf. Figs 3.7 and 3.8). It must be noted at this point that during the high, 160 mm, amplitude experiments the sea level dropped further below the shelf break and resulted in a steeper lowstand stream gradient. The results show a difference in timing of connection relative to the sea-level cycle for the 80 mm and 160 mm amplitude experiments (Fig. 3.9). Generally, experiments of both amplitudes show that the sequences on the shelf and the fluvial system become progressively more out-of-phase as the rate of sea-level fall increases.

Fig. 3.8—Quantitative comparison of a fast and a slow sea-level fall experiment with 160 mm amplitude. (a) Sea-level cycle. (b) Knickpoint migration paths show that the main shelf valley connects at early lowstand in both experiments (indicated by the star below the sea-level curve). The 160 mm amplitude experiments show much faster knickpoint migration rates on the shelf and consequently higher fluxes were measured downstream of the fluvial valley (c) than for the 80 mm amplitude experiments (cf. Fig. 3.7).
Fig. 3.9—The timing of connection, the onset of fluvial erosion, relative to the sea-level cycle. Experiments with a fast rate of sea-level fall develop a shelf valley that connects closer to lowstand than experiments with a slow rate of sea-level fall. High-amplitude and low-amplitude experiments are marked by open and closed symbols, respectively.

**Connection Rate**

All experiments show a delay in response between a fall in sea level and the first features of erosion in the fluvial domain. This delay relates to the time required to propagate the headward erosion process up the shelf towards the fluvial valley. It can be more generally quantified as connection rate, $R_c$:

$$R_c = \frac{L_s}{T_c - T_i} \quad [\text{L/T}]$$

(3.6)

Where $L_s$ is the shelf width under lowstand conditions, $T_c - T_i$ represents the connection delay defined as the period between the moment that the shelf break becomes exposed and the first shelf canyons develop ($T_i$) and the connection time ($T_c$), when a shelf canyon connects with the fluvial valley. The experimental results show that connection time depends on the rate of sea-level fall (Fig. 3.10a). High rates of sea-level fall relate to minor connection delays. However, even for an instantaneous sea-level fall (Experiment 90), it took at least an hour before a shelf canyon connected with the fluvial valley and up to 12 hours before a lowstand equilibrium profile was established. The connection rate correlates well with the average rate of sea-level fall (Fig. 3.10b). For the homogenous substrate in our set-up, the connection rate compares also with the average rate of knickpoint migration on the shelf associated with the headward cutting shelf valleys. Consequently, both the rate of shelf erosion (Fig. 3.11)
and the connection rate (Fig. 3.10b) show a similar strong correlation with the rate of sea-level fall. The observed rates of erosion and deposition on the shelf are entirely reproducible and show a consistent trend over the full range of experiments (cf. Figs 3.7, 3.8 and 3.11).

Fig. 3.10—(a) The connection delay is defined as the time span between the first signs of headward erosion on the shelf edge (triangles) and connection time, the moment that the dominating shelf valley connects with the fluvial valley (stars). (b) The connection rate (Eq. 3.6) correlates well with the rate of sea-level fall. High-amplitude and low-amplitude experiments are marked by open and closed symbols, respectively.
Fig. 3.11—The rate of erosion on the shelf correlates well with the rate of sea-level fall. High-amplitude and low-amplitude experiments are marked by open and closed symbols, respectively.
Chapter 3

The effect of connection time on shelf and fluvial stratigraphy

Figure 3.12 compares the composite stratigraphic sections for the fast-fall and slow-fall experiment (Figs 3.5 and 3.6). It illustrates the degree of diachroneity of the lowstand bounding unconformity (i.e., sequence boundary) on the shelf and fluvial domain. The formation of the sequence boundary on the outer shelf is completed at lowstand, irrespective of the rate of sea-level fall. The surface is thus synchronous with the lowstand. However, further upstream the unconformity was still being formed by continued fluvial erosion during the sea-level rise, which illustrates a clear diachroneity. The amount of diachroneity along the unconformity is much higher for the fast-fall than for the slow-fall experiment, which is reflected in the fluvial and shelf stratigraphy. Thus, the age of the sequence boundary becomes younger upsection and correlates with the rise-inflection point for the slow fall and with the highstand for the fast-fall experiment (Fig. 3.12).

The timing of the connection of one of the shelf canyons with the fluvial valley has a strong bearing on the final volume of the slope fan and lowstand delta, because it determines the change of aggradation to degradation of the fluvial valley. Low rates of sea-level fall produce large lowstand deltas and high rates of sea-level fall produce small ones (Fig. 3.13). The volume of the lowstand delta relates to the time period available for lowstand deposition. The fluvial erosion starts after the peak in deposition of the lowstand delta on the shelf in case of the fast fall (Fig. 3.7c, left). This illustrates that the timing of connection relative to the sea-level cycle not only controls the lowstand-delta volume, but also affects its composition, i.e., the percentage of fluvial over redeposited shelf sediment. As a result, lowstand deltas that are formed during high rates of sea-level fall consist largely of cannibalised shelf material and for only 20-30% of fluvial sediment (Fig. 3.14). In contrast, the lowstand deltas of the slow-fall experiments (shelf valley connected during early fall) contain up to 50% fluvial sediment. Obviously, the percentage of fluvial relative to shelf sediment in the lowstand delta drastically increases after connection.

The rate of sea-level fall has also implications for the rate of deposition on the shelf and in the fluvial valley during the subsequent rise. Experiments with fast rates of sea-level fall show connection close to or just after lowstand, while connection proceeds much earlier for slower rates of sea-level fall (Fig. 3.9). Connection accompanies fluvial degradation and points to the moment that the fluvial valley starts to release previously aggraded sediments. Consequently, the transgressive valley-fill sequence in the fast-fall experiments shows a larger volume than in the slow-fall experiments, as is evidenced by the difference in geometry of the stage-c deposits in Figs 3.12 and 3.13. The full range of experiments shows that the total volume of the transgressive systems tract increases with increasing rates of the preceding sea-level fall (Fig. 3.15).
Fig. 3.12—Composite stratigraphic sections compiled from Experiment 210 and 250. The stages a to d marked in the sea-level curves correspond with the sediments and with the stages of Figs 3.3 to 3.6. The stratigraphic columns on the shelf, coastal plain and river show that the diachronity of the sequence boundary is largest for the fast-fall experiment. With increasing diachronity of the sequence boundary it becomes less straightforward to apply systems tract terminology to upstream alluvial strata. The fluvial deposits overlying the sequence boundary lag half a sea-level cycle behind the sediments overlying the (lowstand) unconformity on the shelf. The comparison between the two experiments is a conceptual test that reveals the problems related to attributing systems tract terminology to alluvial strata (vertical exaggeration 2.25 times).
Fig. 3.13—Lacquer peels of experiments 210 and 250 illustrating the lowstand delta on the shelf break of Fig. 3.12. The stages a to d are delineated on the experimental stratigraphy and show that the fast fall resulted in a smaller lowstand delta (b) but in a thicker transgressive systems tract (c) compared to the slow fall.
Modelling fluvial response to sea-level changes

Fig. 3.14—Composition of the lowstand systems tract (lowstand delta) related to the rate of sea-level fall. Low rates of sea-level fall relate to a high content of fluvial sediment in the lowstand delta. The content of fluvial sediments in the lowstand delta depends on the rate at which sediments are released by the fluvial valley and bypasses over the shelf. The rest of the lowstand-delta volume is cannibalised and redeposited shelf sediment.

Fig. 3.15—The rate of deposition of the transgressive-valley-fill sequence related to the rate of the preceding sea-level fall. The faster the preceding fall, the higher the rate of deposition of the valley fill during the subsequent transgression. The connection delay causes fluvial degradation to occur during early rise in case of a fast sea-level fall with as consequence higher fluxes to the coastal plain and shelf as compared to the case of a slow rate of fall.
Chapter 3

Discussion

Analogue models according to Hooke’s (1968) similarity of process approach, are small dynamic sedimentary systems that, perhaps better than existing numerical models, incorporate simplified versions (analogues) of the fundamental processes (Paola, 2000). Analogue models, in comparison with numerical models, have at least one strong point in favour: They allow only a limited control by the experimentalist and are, therefore, less susceptible to purposeful matching. Our experiments show to be reproducible and to give deterministic results for a range of rates in sea-level fall (see Figs 3.7, 3.10 and 3.11). Hence, the approach is suitable for sensitivity analysis of isolated variables. However, we regard on the one hand a model as heuristic (see Oreskes et al., 1994); it can support a hypothesis, but being only a model it is unable to prove it. On the other hand, it is unlikely that the complex feedback between allocyclic changes can be unravelled from ancient stratigraphy alone (e.g. Blum & Price, 1998).

The first part of the discussion concerns the question which aspects of the model can be applied to real-world river and shelf settings and how the observed rates of knickpoint migration and connection rates in our flume model relate to values for Quaternary shelf evolution. The second part discusses the analogue experimental results in the light of existing stratigraphic concepts for both the fluvial and shelfal realms exemplified by empirical studies. Finally, we discuss what parts of the sequence-stratigraphic concepts need modification in the light of our experimental results to apply for fluvial Successions.

Scalability of headward erosion process

A base profile responds to a drop in base level by the process of upstream knickpoint migration. Here we point to the possibility to use the process of knickpoint migration and the average rate of knickpoint retreat as an independent scaling tool. The knickpoints in our flume tank migrated according to the inclination model of Gardner (1983), which is a realistic process for the applied uniform, non-resistant bed material. The average rate of knickpoint migration on the shelf rapidly declines with increasing time from the onset of sea-level fall (Fig. 3.16a). According to Begin (1988), the distance over which the knickpoint travels is proportional to the square root of the time since the beginning of the sea-level fall. Consequently, the rate of knickpoint migration must be inversely proportional to the square root of the time since base level was lowered (see also Quirk, 1996). In our 18 experiments, knickpoint migration rates for the shelf fit close to this relation as shown in Fig. 3.16b. The question is how do the flume values relate to knickpoint migration rates on an emerged Quaternary alluvial plain and shelf?

The magnitude of knickpoint migration rates in various flume studies and the real world on different time scales is given in Fig. 3.17. Substrate properties, slope and bankfull discharge are not included in the comparison because it is not intended to give a functional relationship for knickpoint migration rates here, but rather to display the range of variation. Knickpoint migration rates in small-scale systems like flume models are presumably higher than on real-world shelves owing to differences in slope and substrate. Sediment transport is proportional to the slope of the channel bed (following any diffusive approach e.g. Begin et al., 1981; Salter, 1993; Leeder & Stewart, 1996) and real-world alluvial plain and shelves have a typically non-uniform
morphology and geology. All experimental studies, except for the one that had very resistant bed material, show values for knickpoint retreat rates within a close range, their variation being related mainly to the duration of an experiment. The lower-right hand side of Fig. 3.17 shows retreat rates for bedrock rivers that vary between 0.001 to 0.1 m/yr., which is in close agreement with model estimates of Whipple & Tucker (1999). Rivers that incise sediments show high knickpoint migration rates for flashflood dominated streams observed over decades. Based on the data in Fig. 3.17 we infer that knickpoint migration rates over sandy and muddy Quaternary shelves and alluvial plains range between 1 and 70 m/yr. However, it must be noted that the data set will include the effect of the measured time interval; a $10^4$ to $10^5$ year change may include a one order of magnitude decrease in erosion rate (Gardner et al., 1987). Thus, the expected values for knickpoint migration on Quaternary shelves may actually be more concentrated in the high part of the range, between 10 and 70 m/yr. Diffusion models support similar numbers for the Mississippi River in response to the 100 m sea-level fall during the Wisconsin glaciation (Salter, 1993; Leeder & Stewart, 1996). Both models assume that sea-level-fall-induced incision occurred as far as 350 km inland of the Mississippi outlet (shelf edge), thus up to Baton Rouge (Saucier, 1996). Salter (1993) calculates that more than 6600 years are required for the knickpoint to reach that position. This would implicate a maximum average rate of knickpoint migration of 50 m/yr (A in Fig. 3.17). In contrast, the model of Leeder & Stewart (1996, their fig. 6) implies a lower value, around 3 m/yr for knickpoint migration up to Baton Rouge (B in Fig. 3.17). Although it is clear that estimates for knickpoint migration rates show a large spreading, field observations and diffusion models support the existence of several kyr response time between 4th order glacio-eustatic sea-level fall and the related fluvial response.

![Graphs](image)

Fig. 3.16—(a) Average knickpoint migration rate shows a rapid decline when plotted against the time span over which sea level was lowered. (b) The same data show a linear correlation with the reciprocal of the square root of the time since base level was lowered.
Fig. 3.17—Average knickpoint migration rates plotted against the time scale of their occurrence (observation). The graph does not intend to show a numerical relation, but wants to illustrate the order of magnitude of knickpoint migration rates in flumes and various rivers. In the graph we plotted average retreat rates, although it is known (Gardner, 1983; Leeder & Stewart, 1996) and also observed from our flume experiments that knickpoint migration rates decrease from an initially high value. Bedrock rivers typically adjust their profiles with knickpoint migration rates between 0.001-0.1 m/yr while short term river adjustment indicates values between 100-1000 m/yr. Knickpoint migration rates on alluvial plain and shelves that accommodate 100 kyr glacio-eustatic sea-level changes are inferred to range from 1-70 m/kyr. The symbols represent data of flume studies (Brush & Wolman, 1960; Holland & Pickup, 1976; Begin et al., 1981; Gardner, 1983; Bryan, 1990; Lee & Hwang, 1994). A and B are model estimates for knickpoint migration during the Wisconsin glaciation of the Mississippi River from Salter (1993) and Leeder & Stewart (1996) respectively. The numbers represent river data from: 1 West Tennessee channels (Simon, 1991); 2 Homochitto River (Yodis & Kesel, 1993); 3 Dry Creek (Begin, 1988); 4 Deep Creek (Schumm et al., 1996); 5 St. Catharine Creek (Yodis & Kesel, 1993); 6 Homochitto tributaries (Yodis & Kesel, 1993); 7 Pechahalee Creek (Begin, 1988); 8 Crawfords Creek (Begin et al., 1981); 9 St. Catharine Creek tributaries (Yodis & Kesel, 1993); 10 Harding Bayou (Yodis & Kesel, 1993); 11 Spanish Bayou (Yodis & Kesel, 1993); 12 Saikawa River (Begin, 1988); 13 Pleasant Valley, Nevada (Begin, 1988); 14 Oaklimiter Creek (Begin, 1988); 15 Indus (Leland et al., 1998); 16 Wolf Creek (Eaton, 1991); 17 Mattole River (Pazzaglia et al., 1998); 18 Blue Hills channels (Dick et al., 1997); 19 Gully near Imlay, Nevada (Begin, 1988); 20 Niagara Falls (Wohl, 1998); 21 South River (Bank & Harbor, 1998); 22 Rio Jemez (Pazzaglia et al., 1998); 23 Rappahannock River (Howard et al., 1994); 24 Susquehanna (Pazzaglia et al., 1998); 25 Swede (Stock & Montgomery, 1999); 26 French (Stock & Montgomery, 1999); 27 Amargosa River (Butler, 1984); 28 Tumut River (Young & McDougall, 1993); 29 Cowlet (Stock & Montgomery, 1999); 30 Tumbarumba Creek (Young & McDougall, 1993); 31 Wheeo Creek (Stock & Montgomery, 1999); 32 Shoalhaven River (Nott et al., 1996); 33 Paddys River (Young & McDougall, 1993); 34 Dabang River (Yang & Li, 1988); 35 Maclean River (Weisel & Seidl, 1997); 36 Hawaiian channels (Seidl et al., 1994); 37 Boggy Creek (Young & McDougall, 1993).
Fluvial and shelf response to sea-level fall

Posamentier & Vail (1988) launched the sequence-stratigraphic concept that widespread fluvial deposition would occur during early stages of eustatic fall, followed by progressive valley incision and sediment bypass during late stages of fall. According to this view the first significant river incision commences when the bayline moves basinward from the equilibrium point, i.e., when the shoreline is forced basinward as a result of a relative sea-level fall (forced regression e.g. Posamentier et al., 1992).

In our experiments we observed a quite different sequence of events. The fluvial valley that was aggrading during the highstand progradation continued to aggrade during the fall in sea level until one of the shelf canyons was connected with the fluvial valley. It confirms the ideas of Quirk, (1996), that alluviation during a sea-level fall must be placed in the context of different processes that can occur simultaneously downstream and upstream of the migrating knickpoint. During the fall we observed in every experiment (Figs 3.3 to 3.6): 1) continued aggradation in the river valley and on the highstand-delta plain; 2) continued avulsion at the apex of the original highstand-delta plain; 3) incision by headward erosion at local negative changes in gradient (e.g. highstand delta front); 4) avulsion- controlled change in discharge and activity of middle shelf distributaries and development of inner-shelf down-stepping delta lobes; 5) canyon formation by headward erosion on the outer shelf. Observations 1-4 are similar to the findings of Törnqvist et al. (in press) who described both deposition and erosion in the coastal prism of the Rhine-Meuse system (see also below) during the last glacio-eustatic sea-level fall.

Until connection, the experimental fluvial profile continued to grade to local base level: i.e., the extended highstand-delta front and not to the actual shoreline at the shelf edge. This explains why all experiments showed continuing aggradation in the fluvial valley on the coastal plain and inner shelf during the early phase of sea-level fall. As the sea-level fall proceeded, the aggradation was observed to diminish in the river valley and on the shelf as the knickpoint of the dominant shelf canyon approached the fluvial valley. Aggradation stopped completely after connection. At this point in time, migration of the knickpoint accelerated once it was in the confinement of the valley (e.g. Fig. 3.7b). Erosion in the valley continued until a new equilibrium profile was established (Fig. 3.5c and 6d). The rate at which the equilibrium was established depended on the amount of erosion during the preceding sea-level fall, which in our experiments was related to the rate of fall. The results are conform Schumm’s (1993) assumptions that high rates of sea-level fall will cause significant channel erosion upstream, even after sea-level has returned to its original position.

The observations above support Nummedal’s (1993) concept of alluviation during early sea-level fall for the common river-shelf setting, where the gradient of the fluvial valley is greater than that of the coastal plain. Such deposits are equivalent to the falling stage systems tract (Hunt & Tucker, 1992; Plint & Nummedal, 2000). We add to this concept by pointing out the importance of connection time, which marks the onset of degradation in the fluvial realm. The connection rate is an intrinsic variable of the fluvial system that relates connection time to the sea-level curve (Eq. 3.6). The connection rate illustrates that the rate of sea-level fall not only has
implications for the rate and volume of erosion on the shelf, but also for the volume of aggradational deposits in the fluvial domain, as shown by our experimental results. Consequently, a slow sea-level fall produces thicker falling-stage deposits in the fluvial valley than a fast sea-level fall (Fig. 3.12).

The connection rate depends on the geology of the substrate; the topography (including shelf width), the amplitude and period of sea-level change, and the sediment supply rate by the fluvial system. The connection rate, however, will have a lower value than locally observed knickpoint migration rates, because avulsions of distributaries determine the amount of supply and discharge to the multiple canyon heads. Therefore, in our experiments with a uniform, non-cohesive substrate, the connection rate was observed basically similar to the average knickpoint migration rate of the connecting shelf canyon. In contrast with our experiments that started with identical smooth shelf topography, real-world shelf evolution occurs less predictable, because of the effect of antecedent relief (e.g. Talling, 1998; Ricketts & Evenchick, 1999), a heterogeneous substrate, etc. (e.g. Woolfe et al., 1998).

**Fluvial and shelfal response to sea-level rise**

The experimental valley-fill sediments deposited from lowstand to early rise are composed of small, backstepping delta lobes that onlap on the main unconformity. The backstepping geometry of the transgressive valley fill and its coastal onlap are in high agreement with the general sequence stratigraphic concepts (Posamentier & Vail, 1988). It was successfully modelled in previous analogue experiments (Koss et al., 1994). The backstepping results from coastal retreat that forces the new base profile to intersect its former profile progressively closer to the hinterland (Quirk, 1996). In our experiments, backstepping of the shoreline and subsequent progradation of small delta lobes appears to be an autonomous process, forced by distributary shift and local supply changes within the flooded river valley and not by higher order changes in the sea-level curve. Thus, the stacked delta lobes can be designated as parasequences.

The slow-fall experiments show significant aggradation in the middle part of the fluvial valley during late rise (Fig. 3.6d). This contrasts with the model of Nummedal et al. (1993) that allows aggradation in the lower flooded estuary of the river close to the shoreline only. However, in our view some effects of a sea-level rise further upstream seem likely, since backfilling during a transgression of a formerly incised lower reach of a river involves re-grading of at least a part of its upstream profile (Schumm, 1993). Overall, the experimental results support the notion that the average depth of incision and the length over which the valley floor rejuvenates due to base-level fall largely exceed the average thickness and longitudinal extent of the deposits formed during the subsequent rise (Ethridge et al., 1998).
Modelling fluvial response to sea-level changes

Examples from the Quaternary

How do we attribute river-valley stratigraphy to features like connection rate, and can this concept improve our understanding of fluvial valley stratigraphy in relation to the sea-level curve? A recent review by Blum & Törnqvist (2000) illustrates how the Quaternary alluvial plains of the Colorado and Rhine-Meuse Rivers were controlled by a complex interaction of climate and sea-level changes. Both examples are governed predominantly by sea level and climate controls, because subsidence rates on the alluvial plains were low, about 12 cm/kyr for the Holocene Rhine-Meuse (Törnqvist, 1998) and 3-4 cm/kyr for the Colorado (Blum, 1993). Very analogous to our experimental results, the Colorado and Rhine-Meuse system demonstrate that aggradation occurred in the river valley during the falling stage of the last glacial lowstand, and that incision of fluvial strata occurred during the sea-level rise (Blum & Törnqvist, 2000).

The upper Colorado alluvial plain shows high gradient fluvial terraces, deposited during the falling stage, lowstand and early rise (20-14 ka), that are intersected by less steep, younger (<11 ka) terraces 80 kilometres upslope from the present shoreline (Blum, 1993). Deposition during falling stage and lowstand has been attributed to high sediment yield enabling the river to form fluvial terraces during multiple episodes of aggradation, degradation and abandonment of flood plains (Blum & Valastro, 1994; Blum & Price, 1998). A comparable reconstruction of two intersecting fluvial terraces on the Rhine-Meuse alluvial plain revealed that a high-yield, braided system resulted in predominant aggradation during the last glacial lowstand until the early rise (Törnqvist, 1998). Similar deposits are found to be preserved in the Texas Gulf Coast River systems and in the Po coastal plain (Törnqvist et al., in press, and references therein). Although Blum & Törnqvist (2000) point to climate control as being primarily responsible for alluviation during relative sea-level fall, our model results suggest that forced regression can produce down-stepping fluvial terraces (alluviation) on the emerged inner shelf in conjunction with aggradation on the highstand delta plain.

A phase of aggradation on the Colorado floodplain during the sea-level rise from 20-14 ka was followed by incision, while the sea-level rise continued. Blum et al. (1994) suggested that the incision occurred due to diminished river loads with respect to stream capacity during early to mid-rise from 14-11 ka. The Colorado coastal plain sediments are underlain by a composite basal unconformity that corresponds in part to the lowstand systems tract and in part to the transgressive systems tract (Blum & Valastro, 1994). The same coastal prism is truncated by an erosive surface (14-11 ka) that merges with the basal unconformity upstream, which shows that the unconformity is strongly time transgressive (Blum & Price, 1998). The diachroneity along the sequence boundary complicates the correlation of fluvial strata within the influence of sea-level changes with the relative sea-level curve (Blum, 1993). Note that the diachroneity in the sequence boundary of the model stratigraphy causes similar ambiguities (Fig. 3.12). The model results thus support the notion of Blum & Price (1998) that falling stage and lowstand does not instantly result in complete bypass of the floodplain as proposed in earlier sequence models (e.g. Posamentier & Vail, 1988). However, Blum & Price (1998) regard the time lag between sea-level fall and its related incision upstream the alluvial plain as insignificant compared to the rate and
duration of the base-level fall itself. Our modelling shows how fluvial incision continues during early to mid sea-level rise (e.g. Fig. 3.5). From this perspective we suggest that incision of the Colorado alluvial plain during rise (14-12 ka) might have been promoted both by delayed headward erosion originating from the sea-level fall in combination with an increase of the river load relative to stream capacity related to climate change.

**What do the analogue experiments tell us?**

On a common level, the experimental results support the hypothesis that a sea-level fall does not instantly lower the downstream reach of the fluvial system (Butcher, 1990; Blum & Price, 1998; Dalrymple *et al*., 1998; Ethridge *et al*., 1998). The model clearly shows a delay between exposure of the shelf and the moment that sea-level fall induced erosion advanced to the fluvial domain. Connection time in relation to the sea-level curve is most important for understanding fluvial and basin stratigraphy in terms of genetically related sequences (cf. Butcher, 1990; Quirk, 1996).

The strong impact of the Basin response factor, Br (Eq. 3.1) on the analogue model results support the notion of Paola *et al.* (1992) that the ratio of the period of change (T) over the model’s equilibrium time (T_eq) is extremely important. A few equations exist to estimate equilibrium times from real-world rivers. Application of present-day values for large river systems yields equilibrium times of the order of 1 to 10 kyr, but it is not straightforward to derive input values from ancient stratigraphy. However, we feel that the Basin response factor needs further exploration and refinement for the coupling among basin models themselves and with their prototypes. The strong bearing of the Basin response factor on the final analogue model stratigraphy supports the notion that comparison of Quaternary 4th order glacio-eustatic sequences with longer term eustatic sequences must be done with caution (cf. Boyd *et al*., 1989; Poag, 1992; Talling, 1998, with Blum, 1993 page 277).

A remaining question is how the sequence-stratigraphic concept applies to the fluvial valley sediments within the knickpoint reach. Delineation of our experimental stratigraphy allows a precise reconstruction of the timing of unconformity formation relative to the imposed sea-level cycle. The exercise shows that it is not straightforward to put systems tract terminology on the fluvial and even shelfal deposits that were only affected by a single sea-level cycle (Figs 3.12 and 3.13). Apparently, there are some obvious pitfalls as misinterpretation of aggradational strata that formed in the fluvial valley during the early sea-level fall. It becomes preserved as a fluvial valley-fill that seems out of phase with the sea-level change. The experiments make clear that with increasing diachronity of the sequence boundary it is increasingly difficult to attribute systems tracts to fluvial strata. Under such conditions correlation between sediment body and position in the sea-level cycle remains speculative, despite their very suggestive names. The degree of complexity of our experimental stratigraphy demonstrates that there cannot be a simple and common rule that can correct for these out-of-phase relationships. Only absolute time-constrained stratigraphy, which is implicit in analogue model studies can elucidate on time-lag relationship in stratigraphic successions.
Conclusions

The analogue experiments with sea-level change as the only independent variable proved to be reproducible and produced statistically significant quantitative data for various rates of sea-level change. The results support the notion that neither a fall nor rise in sea level does instantly affect the upstream fluvial reaches. We have quantified this lag in system’s development by the term connection delay that represents the time required to connect incipient shelf canyons on the just emerged shelf with the fluvial valley by the process of headward erosion. In order to study such delays more generally, we introduced the quantity connection rate: the ratio between shelf width and the connection delay. The Connection rate is a function of the rate of headward erosion induced by the sea-level fall. It showed a strong bearing on fluvial and shelfal stratigraphy by controlling:

1) the amount and duration of initial fluvial aggradation during sea-level fall;
2) the percentage of fluvial sediment versus eroded shelf material in the lowstand delta;
3) the volume of the lowstand delta;
4) the volume of the transgressive systems tract;
5) the amount of diachroneity along the sequence boundary.

The results support the idea that designating systems tract terminology to fluvial strata is appropriate up to the upstream limit of sea-level-fall-induced erosion (i.e., knickpoint) for small connection delays. Only absolute time-constrained stratigraphy can elucidate on time-lag relationships in stratigraphic successions.

Acknowledgements

Shell International Exploration and Production, Rijswijk, The Netherlands funded the research. We thank R. G. de Jongh and G.W.M. de Ruiter for initiating the scientific collaboration with Shell. We acknowledge the permission of Shell to publish this paper. P.L. de Boer, J. Cleveringa, W.P. van Kesteren, X.D. Meijer, T.E. Törnqvist and W. Schlager are thanked for their critical reading of an earlier version of the manuscript. At Utrecht University we are grateful to A.C. van der Gon Netcher, J.H. Bliek, P. Anten and M. Reith for their technical support and M. de Kleine for his assistance during the first series of experiments.