ABSTRACT: Growth faults and deltaic coastlines are common features along basin margins with a high flux of sediment delivery. It has been observed in previous studies that wave-dominated deltas are the characteristic and long-lived coastline type in such growth-faulted, extensional basins. During deltaic progradation the structural and depositional systems migrate and adjust as sedimentation rates keep pace with fault displacement rates. A numerical model indicates that increased subsidence along growth faults slows the progradation rate of deltas as new accommodation is being created. The decrease in forward delta growth results in a longer period of wave reworking per unit of sediment due to the retention of large fetch and water depth. Since subsidence is maintained through time, sediments tend to accumulate on delta topsets, creating thick wave-dominated delta-front successions on the downthrown sides of faults. The proposed model may guide subsurface interpretations where it is difficult to assess the shoreline type without cores. It is proposed here to use values of expansion index derived from seismic reflection profiles or well-log correlations as a proxy for the likelihood of “wave dominance” on the delta shoreline.

INTRODUCTION

Large, basin-scale growth faults are common at all continental margins and are long-lived structures (million years) that affect stratigraphy. They have been described in the proximity of the shelf edge in both active-margin and passive-margin basins, where sediment supply along a delta-distributary fairway is high (Winker and Edwards 1983; Van Rensbergen and Morley 2000; Cummings and Arnott 2005) (Table 1). Local growth faulting associated with smaller-scale delta progradation and slope instability can also influence stratigraphy (Blatt and Davies 2001; Armstrong et al. 2013), but these faults are not the objective of this study. Large, growth-faulted compartments efficiently trap deltaic sediments, resulting in marked stratal thickening (by up to tens of times) across the fault (Fig. 1) (Winker 1982; Edwards 1995; Rouby et al. 2011). The geometry and mechanism of growth-fault movement have been described using seismic and well-log correlations (Doust and Omatsola 1989; Edwards 1995; Rouby et al. 2011; Khani and Back 2014) and geomechanical and mathematical models (Cranes et al. 1980; McClay et al. 1998), but little is known about depositional environments and sedimentation in the growth compartments. It is argued here that growth faults on the outer shelf or at the shelf edge decrease shoreline progradation and control the amount of sediment delivered to the delta front and the delta process regime. The enhanced wave dominance happens because increased subsidence creates local accommodation which slows down the shoreline progradation and allows enhanced wave reworking. The numerical model documented below suggests a direct correlation between the expansion index (ratio of the downthrown thickness to upthrown thickness, EI) and the likelihood of wave dominance; the higher the EI, the higher the likelihood of a wave-dominated regime. This model helps explain the common occurrence of wave-dominated deposits on growth-faulted margins described in previous studies (Table 1).

WAVE-DOMINATED DELTAS AND DISTRIBUTION OF THE DELTAS ON THE OUTER SHELF AND AT THE SHELF EDGE

The main drivers for shoreline migration and shelf-margin accretion are fall of sea level (Mitchum and van Wagoner 1991; Porębski and Steel 2003) or high sediment supply (Carvajal and Steel 2009). Shelf-margin accretion is attained through multiple regressive-transgressive (R-T) transits of deltas from inner shelf to the shelf edge under allogenic controls (sea level, sediment supply) during which time multiple deltaic depocenters are formed under various process regimes (Galloway 2008; Olariu et al. 2012a; Olariu 2014). It is generally thought that tide-dominated deltas are more common on the inner shelf (Ainsworth et al. 2011) (at highstands, see modern tidal deltas; Allen et al. 1976; Doust and Omatsola 1989; Gastaldo et al. 1995; Reijers 2011), whereas wave-dominated deltas are more common at the shelf edge (Porębski and Steel 2003; Steel et al. 2008; Bowman and Johnson 2014). Coastline morphology, shelf width, bathymetry, and proximity to the shelf edge play a major role in determining whether a coastline is tidally-influenced or wave-modified (Longhitano et al. 2012). Wider shelves (> 75 km) have the potential for decreasing wave energy at the shoreline by increasing frictional energy losses to the sea floor, as the waves move across a shallow wide shelf (Ainsworth et al. 2011). Hence, wide shelves (inner shelves at highstands) have a relatively high potential to produce tidal dominance at the coastline. Porębski and Steel (2003) and Yoshida et al. (2007) argue that deltas at the shelf edge are likely to be wave dominated because wave energy is high as open-ocean waves approach the shelf.

Relative changes in rates of creation of accommodation and sediment supply also control the processes acting at the shoreline (Ainsworth et al. 2011). The presence of fluvial-dominated deltas at the shelf edge is linked with high sediment supply, which overcomes basin reworking processes by waves and tidal currents (Coleman and Wright 1975), and is often
necessary for the delivery of sand to deep water (Dixon et al. 2013). Wave-dominated deltas reworked by longshore drift and tide-dominated deltas reworked by ebb and flood tidal currents are inefficient suppliers of sediment to deep water (Carvajal and Steel 2009; Dixon et al. 2012) unless the head of a canyon indents the shelf edge and connects with the delta (Sylvester et al. 2012) or directly with the river (Savoye et al. 2009).

High ratios of rates of accommodation to sediment supply increase the wave effectiveness unless waves are attenuated by shoreline morphology such as in highly or moderately embayed settings where tidal processes dominate (Longhitano et al. 2012). In unstable settings along high sediment supply margins (Table 1) growth faults control increased local accommodation in their hanging walls due to fault subsidence. The increase subsidence results in expanded thickness of sedimentary units and a longer time of wave reworking per unit of sediment on the outer shelf due to unrestricted fetch and water depth. Wave dominance in the deltas on the outer shelves in the growth compartments implies that building of the upper slope through aggradation due to sustained fault subsidence causes a rising shoreline trajectory and decreases the likelihood of the presence of the deep-water fans unless canyons incise the shelf edge, allowing bypass of sediment to deep water (Olariu et al. 2013). This study presents growth faults as an additional control on wave dominance in the outer-shelf deltas.

GROWTH-FAULT DEVELOPMENT

In unstable and high-supply progradational clastic shelf margins, growth faults can form as a result of differential compaction due to uneven loading of the basin margin with deltaic sediments which deform the underlying ductile, thick, prodelta shales (Fig. 1) (Winker and Edwards 1983; Morley and Guerin 1996; Magbagbeola and Willis 2007). There is a feedback mechanism: the relatively dense delta sediment load is the driver for the fault initiation, but at the same time acts to “heal” the accommodation created by local subsidence. The fault displacement at the surface has only minimal relief (Armstrong et al. 2013) if aggradation rates are high enough to keep pace with fault subsidence. When rates of sediment input are less than subsidence due to fault movement, significant relief can be generated (Cartwright et al. 1988; Edwards 1995; Moscardelli et al. 2012). Basin-scale growth faults can also be associated with basement tectonics, deep salt movement, slumps at the shelf edge (gravity sliding), or a combination of these factors (Crans et al. 1980; Winker and Edwards 1983). Since the faults are active while sediments accumulate, this results in local higher depositional rates in the downthrown compartment and the “expansion” of the stratigraphic interval (Galloway et al. 1982; Winker 1982; Edwards 1995; Rouby et al. 2011) (Fig. 1). Growth-faulted depocenters are wedge shaped and extend for tens of kilometers along depositional dip. The largest faults that bound major depocenters are tens of kilometers long, have seaward-facing concave geometries in plan view, and are oriented subparallel to the shelf edge (Magbagbeola and Willis 2007; Rouby et al. 2011; Olariu et al. 2012b).

Many modern and ancient large deltas are formed on growth-faulted segments of both passive margins such as in the Gulf of Mexico (Galloway et al. 1982; Winker 1982; Edwards 1995), Niger (Magbagbeola and Willis 2007; Rouby et al. 2011) and the Nile delta (Sestini 1989), and in active margins, such as the Orinoco (Wood 2000; Bowman and...
Sedimentation in the growth-fault model continues until the entire
shelf edge, (3) the delta geometry is described by a horizontal topset
and foreset (constant dipping angle, \( \alpha \)) prograding in constant
water depth; (4) sediment supply is constant; and (4) the eustatic sea
level is stationary. These assumptions simplify the model without
affecting the results of the sediment partitioning.

The model (Fig. 2) assumes that river-derived sediments (area A on
stage 1, Fig. 2) covering the area of the delta will be spread over
the topsets and foresets (stage 2, Fig. 2). Conservation of area
(volume in 3-D) is attained when the following equation is valid:

\[ A = A_1 + A_2 + A_3 \]

where \( A_1, A_2, \) and \( A_3 \) are cross-section areas of topset, topset–foreset
transition, and foreset, respectively.

Assuming that a delta has experienced \( n \) steps of growth faulting,
then cross-section areas are

\[ B_{1n} = g \left( d + \sum_{i=1}^{n-1} x_i \right) \]

\[ B_{2n} = x_n g + \frac{1}{2} g \left( \frac{x_n}{\tan \alpha} \right) \]

\[ B_{3n} = \left( x_n + \frac{g}{\tan \alpha} \right) \left[ h_i + g (n - 1) \right] \]

where \( d \) is the landward position of the growth fault with respect to
the shelf edge, \( g \) is the fault growth rate, and \( h_i \) is the height of the delta
front.

Replacing \( B_{1n}, B_{2n}, \) and \( B_{3n} \) in equation 1 and solving for \( x_n \) results in

\[ x_n = \frac{1}{ng} + h_i \left[ nA - \left( \frac{2n-1}{2\tan \alpha} \right) g^2 \right] \left( \frac{h_i}{\tan \alpha} + d + \sum_{j=2}^{n-1} x_j \right) g \right] \]

Equation 2 shows the relationship between \( x_n \) (shoreline progradation
rate after \( n \) steps) and the growth rate \( g \) (stage 3 in Fig. 2).

Using equation 2, the progradation rate through time (Fig. 3A), and
the shoreline trajectory (Fig. 3B) affected by the growth fault can
be described. The progradation rate decreases with time because
the sediment is trapped in the growth-faulted compartment behind
the

### Table 1.—Examples of deltas from growth-faulted shelf margins in passive (upper part) and active (lower part) settings. Process regime for both modern
and ancient deltas within fault bounded depocenters inferred from published studies is based on seismic data (coherence horizon slices), well logs
(sandstone maps), and core and outcrop data.

<table>
<thead>
<tr>
<th>Deltas</th>
<th>Age</th>
<th>No. of Fault-Bounded Depocenters/ Width (km)</th>
<th>Sedimentation Rate (m/Myr)</th>
<th>Clastic Wedge Thickness (km)</th>
<th>Progradation Rate (km/kyr)</th>
<th>Delta Process Regime</th>
<th>Ancient</th>
<th>Modern</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Niger</td>
<td>Eocene to present</td>
<td>(6)/20-50</td>
<td>250-350</td>
<td>12</td>
<td>0.02</td>
<td>wave dominated (seismic, well logs, core)</td>
<td>NA</td>
<td>NA</td>
<td>Rouby et al. 2011; Reijers 2011</td>
</tr>
<tr>
<td>Nile</td>
<td>Eocene to present</td>
<td>20-30</td>
<td>5000</td>
<td>4</td>
<td>10</td>
<td>wave dominated (cores)</td>
<td>NA</td>
<td>NA</td>
<td>Doust and Omatsola 1989</td>
</tr>
<tr>
<td>Norias</td>
<td>Oligocene</td>
<td>(6)/2-5</td>
<td>490</td>
<td>4.5</td>
<td>0.02-0.04</td>
<td>wave dominated (well logs, core)</td>
<td>NA</td>
<td>NA</td>
<td>Sestini 1989</td>
</tr>
<tr>
<td>Mackenzie</td>
<td>Paleogene to present</td>
<td>(3)/20-40</td>
<td>NA</td>
<td>12-16</td>
<td>NA</td>
<td>river dominated, wave dominated (well logs, core)</td>
<td>NA</td>
<td>NA</td>
<td>Costellier and Stanley 1987</td>
</tr>
<tr>
<td>Danube</td>
<td>Cretaceous to present</td>
<td>5-10</td>
<td>NA</td>
<td>2</td>
<td>NA</td>
<td>river dominated (seismic)</td>
<td>NA</td>
<td>NA</td>
<td>Galloway et. al. 1982</td>
</tr>
<tr>
<td>Orinoco</td>
<td>Pliocene to present</td>
<td>(3)/20-30</td>
<td>5000-10000</td>
<td>12</td>
<td>0.016</td>
<td>wave dominated (outcrop, well logs)</td>
<td>Uroza 2012; Sydow et al. 2003; Wood 2000</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Baram</td>
<td>Pliocene to present</td>
<td>NA</td>
<td>1200</td>
<td>6</td>
<td>0.011</td>
<td>tide dominated, wave influenced (well logs, seismic)</td>
<td>Saller and Blake 2012; Van Rensbergen and Morley 2000</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mekong</td>
<td>Pliocene to present</td>
<td>NA</td>
<td>4700</td>
<td>8-20</td>
<td>0.004</td>
<td>tide-dominated (core)</td>
<td>Nguyen et al. 2000</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Champion</td>
<td>Miocene to present</td>
<td>10-30</td>
<td>1000</td>
<td>6.75</td>
<td>0.004</td>
<td>wave dominated (well logs, seismic) wave dominated</td>
<td>Saller and Blake 2012; Van Rensbergen and Morley 2000</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Johnson 2014); Baram and Champion deltas (Saller and Blake 2012) (see
Table 1). Growth faults are typically mapped on seismic reflection
profiles with sparse well and core control (Winker and Edwards 1983;
Maghabeela and Willis 2007; Back et al. 2006). As a result, seismic
profiles are useful for predicting the fault geometry and structural
details, but not the lithology or depositional environments (Winker and
Edwards 1983). Studies based on well-log correlations commonly interpret
wave-dominated shorelines or deltas in growth compartments (Galloway et al.
1982; Doust and Omatsola 1989; Edwards 1995; Brown et al. 2004; Saller
and Blake 2012) based on interpretation of log patterns and sandstone
isopach maps. There are relatively few detailed studies based on core
and outcrop interpretations (Bhattacharya and Davies 2001; Sydow et al.
2003; Olariu et al. 2013; Bowman and Johnson 2014).
shoreline (Figs. 2B, 3A, B). The decrease in the rate of the delta shoreline progradation (Fig. 3A) is accompanied by aggradation and a rising shoreline trajectory (Fig. 3B). It is considered that the progradation rate reflects the amount of sediment delivered to the coast. Under the same wave regime at the shoreline, a decrease in the progradation rate (due to lower sediment flux) is reflected in stronger wave reworking of the sediments and results in a wave-dominated delta with characteristic facies and morphology (Bhattacharya and Giosan 2003; Hampson and Storms 2003). To depict an example of river- to wave-dominated delta-front transition, a progradation rate threshold of 500 m/time interval (one step for the time interval can be considered 50 years or less) was chosen as a tipping point to switch to a wave-dominated regime. Obviously, the physical characteristics of each system, sediment flux or wave energy (Wright and Coleman 1972), will control the threshold from river- or tide-dominated regime to a wave-dominated regime. Growth rates along the fault vary through time and also spatially. The model, run with different growth rates, shows that the higher the rate of subsidence induced by growth faults, the slower the shoreline progradation (Fig. 3C), increasing the chance that the shoreline will turn into a wave-dominated system. The numerical model shows that at a high rate (1.1 m/time unit, green line in Fig. 3) the delta becomes wave dominated relatively quickly, whereas at a slower rate (0.1 m/time unit, blue line on Fig. 3) the delta maintains a fluvial character for a longer time.

**EXPANSION INDEX AS A PREDICTOR OF WAVE-DOMINATED DELTAS IN THE GROWTH COMPARTMENTS**

The modeled decrease of the shoreline progradation rate and the presence of an aggradational shoreline trajectory is similar to models of shoreline autotreat (Muto and Steel 1992), a key process in autostratigraphy (Muto et al. 2007). In the case of autotreat, the shoreline trajectory is controlled by the regional steady rise of relative sea level, whereas in this study the shoreline trajectory is controlled by the locally variable rate of growth-fault movement (which induces relative sea-level rise due to sustained subsidence). However, the rate of growth-fault movement used for geometrical models is difficult to measure or estimate in ancient deposits. A common parameter used to describe the behavior of growth faults in subsurface data is the expansion index

\[
EI = \frac{T g_r + h_i}{h_i}
\]

where \(h_i\) is the height of initial delta, \(T\) is time, and \(g_r\) is the rate of fault growth (Thorsen 1963). The expansion-index method is useful in determining times of most significant growth (Edwards 1995). If each growth compartment was active for an equal amount of time, then EI is proportional to the growth rate. Therefore, a higher EI (Fig. 4D) would be associated with a higher growth rate and, according to the model, a higher probability of wave-dominated shoreline deposits in the growth strata. The EI values vary during the evolution of a fault (Fig. 4B, C, D), but usually increase and then decrease (Winker and Edwards 1983; Edwards 1995). According to the model presented here, the deposits are more likely to be wave-dominated during the maximum EI (commonly during the middle of the growth-fault evolution). The threshold value of the EI for wave dominance depends on the characteristics of each system, and more studies are needed to link the EI with sediment flux and wave regime. For modern deltas the discharge effectiveness index (ratio of the average river discharge per unit channel width to the average nearshore wave power per unit crest width) is an indicator for fluvial to wave morphology and dominance (Coleman 1976). If the river (sediment) discharge to the shoreline decreases relative to wave energy (e.g., due to...
up-dip storage of sediment in localized accommodation created by a growth fault, the discharge effectiveness index will decrease and create a wave-dominated deltaic morphology (Coleman 1976). However, for ancient systems imaged in the subsurface, EI is a useful proxy for local relative accommodation increase and as a consequence for the dominant shoreline process because it can be estimated from seismic or well-log data in the absence of core data.

WAVE-DOMINATED DELTAS IN GROWTH-FAULTED COMPARTMENTS

The two findings of this study are that (1) increased subsidence in growth-faulted compartments decreases the rate of shoreline progradation and increases the likelihood of shoreline wave dominance, and (2) the expansion index (EI) can be used as a proxy for wave dominance because it controls the shoreline progradation rate and hence the ability of waves to rework river-derived sediments.

The numerical model (Fig. 2) indicates that deltaic coastlines affected by growth faults have to accommodate the new space continually created on the shelf behind the shoreline, leading to a decreased deltaic progradation rate (Fig. 3A). The modeling supports the ubiquity of wave deposits in the growth compartments, as have been commonly interpreted in many previous studies (Table 1).

Multiple growth faults on the northern Gulf of Mexico margin have controlled the stratigraphy since the Cretaceous (Fig. 4A). The Wilcox Group in Texas records the first major Cenozoic influx of clastic sediment into the west and central Gulf Coast basin. Basin-margin subsidence was controlled largely by differential sediment loading on thick unstable shale (Winker 1982; Edwards 1995). The sedimentary sequence is extensively affected by synsedimentary and postsedimentary normal faults (Fig. 4D) which separate narrow elongate subbasins oriented parallel to the coastline (Winker and Edwards 1983; Ewing 1986). Overall, there is a transition of depositional environments from tide-influenced and wave-modified to wave-dominated deltas as the shoreline approached the shelf edge during deltaic progradation (Figs. 4D, 5B). A thick sedimentary succession characterized by repeated stacking of shoreface sequences, characteristic of wave-dominated delta fronts, was deposited in the downthrown depocenters (Fig. 4).

The Oligocene Frio Formation shorelines are affected by growth faults which define six subbasins along the Texas shelf and are the main mechanism for providing space to thicken the section basinward (Galloway et al. 1982; Ewing 1986; Brown et al. 2004; Bonnaffé et al. 2008). Correlation within each subbasin (Olariu et al. 2012b) shows growth faults affecting multiple shoreline transits (Fig. 4B). The sedimentary succession in the downthrown basins is characterized by repeated vertical stacking of shoreface sequences and an increase in sediment thickness towards the fault. The preserved deltaic strata display an aggradational architecture (Fig. 4B) with little landward or seaward migration of the shoreline (Olariu et al. 2013). Strong aggradation is due

![Fig. 3.—Selected outputs from numerical model. A) Decrease of the shoreline progradation rate with time (for parameter definitions see Fig. 2). B) Shoreline trajectory in the growth compartment (V.E. = 500; parameters have the same values as in Fig. 3A). C) Shoreline progradation rate. The decrease in shoreline progradation rate is controlled by the rate of movement of the growth fault.](image)
UBIQUITY OF WAVE-DOMINATED Deltas IN OUTER-SHELF GROWTH-FAULTED COMPARTMENTS

A

B

C

D

J S R
to a combination of high sediment supply and high subsidence rates enhanced by growth faulting. Core samples taken from various subbasins and at various time intervals (Fig. 5A) show the shoreline deposits to have a predominantly storm-wave-dominated character.

In wave-dominated deltas, the sediment delivered to the distributary-mouth bar is carried away by longshore transport (Coleman and Wright 1975; Coleman 1976; Bhattacharya and Giosan 2003). Thus, compared to river-dominated deltas, the progradation rate of wave-dominated deltas is retarded; hence, deltas stay at the shelf edge and are reworked by waves for longer periods of time. Consequently, the primary reason for enhanced wave dominance is the slowing down of progradation (less sediment supplied to the same shorezone width per unit time), or alternatively the partitioning of sediments over large areas (increasing the shorezone width along strike), giving less sediment supplied in the shorezone per unit time. Growth faults act as an additional control on the outer-shelf deltas, increasing the probability of wave-dominated sedimentation in the growth-faulted depocenters.

**MODEL APPLICATION TO WILCOX GROWTH FAULTS**

To model ancient deposits, parameters describing sediment supply, the geometry (angle) of the delta front, or the fault growth rate are needed. The delta (parasequence) thickness and EI can be estimated from well logs (Fig. 4 and table in Fig. 6). The relationship between EI and fault growth rate is given by

\[ g_r = \frac{h_i (EI - 1)}{T} \]  

where \( g_r \) is the growth rate, \( h_i \) the height of the delta front, \( EI \) is the expanding index, and \( T \) is time.

The most important parameter, and the most difficult to estimate, is the duration of deposition in the growth compartment. Since the entire lower

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**Fig. 5**—Core samples of A) Frio Formation and B) Wilcox Group (each core sample is 3 inches (7.6 cm) wide). Alternation of highly bioturbated sandstones representing fair-weather wave deposition with fine-grained, massive sandstone beds representing storm events indicates deposition on wave-dominated shorelines. The trace fossil diversity in fair-weather deposits is high, with assemblages reflecting *Cruziana* ichnofacies, whereas storm-event beds contain only *Ophiomorpha*.

**Fig. 6**—Geometrical model applied to Wilcox growth faults. A) The two models for calculating the delta progradation in growth-fault compartments. Case 1: No Sediment Spill—the faults are considered separately. Case 2: Sediment Spill—the delta units are affected by successive growth faults. B) Modeling results for nine Wilcox Group units from growth-fault compartment 2. C) Modeling results for nine Wilcox units from growth-fault compartment 3. D) Modeling results for nine Wilcox units that are affected by successive growth faults in compartments 2 and 3. The table shows the parasequence thickness \( h_i \) and the expansion index \( EI \) used in the model, and obtained from Figure 4D.
Table with units thickness and expansion index used in the modeling

<table>
<thead>
<tr>
<th>Unit name</th>
<th>Thickness fault 2 (m)</th>
<th>EI for fault 2</th>
<th>Thickness fault 3 (m)</th>
<th>EI for fault 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wx-A</td>
<td>122</td>
<td>2.7</td>
<td>293</td>
<td>1.3</td>
</tr>
<tr>
<td>Wx-B</td>
<td>125</td>
<td>3.9</td>
<td>313</td>
<td>1.7</td>
</tr>
<tr>
<td>Wx-C</td>
<td>54</td>
<td>2</td>
<td>76</td>
<td>1.2</td>
</tr>
<tr>
<td>Wx-D</td>
<td>46</td>
<td>1.6</td>
<td>45</td>
<td>1.2</td>
</tr>
<tr>
<td>Wx-E</td>
<td>38</td>
<td>1.9</td>
<td>35</td>
<td>1.5</td>
</tr>
<tr>
<td>Wx-F</td>
<td>51</td>
<td>1.3</td>
<td>50</td>
<td>1.2</td>
</tr>
<tr>
<td>Wx-G</td>
<td>43</td>
<td>1.4</td>
<td>51</td>
<td>1.1</td>
</tr>
<tr>
<td>Wx-H</td>
<td>76</td>
<td>1.3</td>
<td>59</td>
<td>1.1</td>
</tr>
<tr>
<td>Wx-TopA</td>
<td>68</td>
<td>1.6</td>
<td>81</td>
<td>1.1</td>
</tr>
</tbody>
</table>

Legend:
- Fluval delta
- Wave delta
- Transgression
Wilcox Group lasted about 1.1 Myr and is interpreted to contain 10 regressive-transgressive cycles or parasequences, this results in about 100 kyr for each parasequence. The problem is that depositional time at the scale of 100 kyr represents the entire regressive time from inner to outer shelf and the succeeding transgression, but the growth fault was active only for that part of the time interval when the delta prograded (about 5 km) over the growth fault (Fig. 4D). The question that arises is: for how long was the delta actually prograding? If a minimum progradation rate of 1 m/yr (100 km in 100 kyr) is considered, this will result in a progradation time of about 5 kyr. Alternatively and most likely, if the delta prograded at a rate of 10 m/yr (half of the rate for modern Mississippi Delta; Roberts 1997) then the time for delta progradation over the growth fault is about 500 years.

For the Wilcox Group calculations we used the $h_i$ and EI values obtained (Table in Fig. 6) from well-log correlations (Fig. 4D), a delta front foreset angle of $2^\circ$, and an initial progradation rate of 20 m/yr. The distance of the growth fault back from the shoreline was considered to be 0 m, which means that the growth fault starts moving as soon as the delta progrades to the fault location.

Modeling results for deltas in nine regressive-transgressive units (parasequences) mapped in the Wilcox Group (Fig. 4D) show a variable response (Fig. 6). Modeling was done for individual growth faults ("No Sediment Spill" model) or combined the growth of successive faults ("Sediment Spill" model) (Fig. 6A). Growth-fault compartment 2 (Fig. 4D) was modeled for all deltas with an initial progradation rate of 20 m/yr and shows that two of the deltas cease progradation and undergo transgression (negative progradation), five deltas develop a wave-dominated character (progradation decreased to less than 10 m/yr), and two deltas maintain their fluvial dominance with progradation rates higher than 10 m/yr (Fig. 6B). Similar modeling (with an initial 20 m/yr progradation rate) of growth-fault compartment 3 (Fig. 4D) shows that most of the shorelines maintain high (over 10 m/yr) progradation rates, suggesting fluvial dominance, and only two shorelines become wave dominated or are transgressed (Fig. 6C). However, when the effects of the two growth faults acting on the same deltaic shoreline are combined, all the shorelines become wave dominated (progradation less than 10 m/yr) or transgressive (Fig. 6D). Although simple, this model shows the importance of considering multiple (more than two) growth faults that are active along the basin margin.

**DELTA PROCESS REGIME CHANGES AND GROWTH FAULTS**

Possible process regime changes during deltaic progradation across a shelf in both stable and unstable settings are illustrated in Figure 7. If a river discharges into a coast with strong waves and longshore currents, the deltas will be wave dominated throughout their progradation (Coleman and Wright 1975; Dominguez and Steel 1996; Rao et al. 2005; Vella et al. 2005), from the inner shelf to the edge of the shelf (Fig. 7A). Deltas that are tide- or river-dominated on the inner shelf can turn into wave-dominated deltas on the outer shelf (Porbiski and Steel 2003; Yoshida et al. 2007; Ainsworth et al. 2011; Olariu 2014), as they prograde into deeper water with a stronger wave regime while fluvial and tidal energy remain the same or decrease (Fig. 7B). If the delta topset aggrades during deltaic progradation, the sediment flux at the shoreline will decrease (Fig. 7C) even at a steady sea-level stand (Fig. 7D). In wave-modified deltas, much of the sediment delivered to the distributary-mouth bar is carried away by longshore transport. Thus, the progradation rate of wave-influenced deltas is retarded allowing rivers to maintain a greater slope that inhibits avulsion (Bhattacharya and Tye 2004). Autocyclic processes related to shifting distributaries result in delta lobe switching and produce multi-lobate deltas of either river-dominated (Roberts 1997; Roberts et al. 2004) or wave-dominated character (Sornosa et al. 1998) that thicken toward the shelf edge. Delta-lobe switching occurs mostly on the inner shelf as distributaries change their direction of progradation through avulsion, since accommodation on the inner shelf is low. Increased accommodation due to fault-driven subsidence in the vicinity of the shelf edge reduces deltaic progradation and inhibits avulsion, favoring delta aggradation. This will result in deposition of stacked and thickened wave-dominated delta lobes. However, when growth faults start a significant distance back on the inner shelf and the growth rate is relatively small compared to the river sediment discharge, the deltas are either tide- or river-dominated (Bhattacharya and Davies 2001; Armstrong et al. 2013, see also blue line in Fig. 3A). Such inner-shelf growth faults are small-scale features (tens of meters of displacement) associated with delta progradation and local slope instability. Thinner underlying muds and smaller gradients (< 0.1°) on the continental shelf allow less accommodation for growth-faulted structures (Bhattacharya and Davies 2001).

Despite the fact that wave dominance is emphasized in this study, we do recognize that there are several examples of growth-faulted river-dominated (Lagniappe Delta, Kindiger 1989; Roberts et al. 2004) or tide-influenced shelf-edge deltas (Sable Delta, Cummings and Arnott 2005). Wave processes should be expected where accommodation is higher than river sediment supply, unless waves are attenuated by shoreline morphology (Coleman 1976). The evolution of the river-dominated Lagniappe shelf-edge delta (Wisconsinan of Gulf of Mexico margin) was peculiar in that it was controlled by salt diapirs uplifted along the shelf break (Kindiger 1989). The diapirs acted as a barrier to basinward progradation of the delta, and directed sediment fairways between and around the uplifts. In addition, rapid sea-level fall (> 30 m in less than 3000 years) during the late Pleistocene glacial maximum forced rapid progradation of the delta to the outer shelf, but diapirc uplifts entrenched sediment delivery networks landward of the shelf edge (Roberts et al. 2004), thus reducing wave influence.

Core data show tidal deposition in the more proximal parts of the Sable Delta (Jurassic of Nova Scotia margin), whereas on the outer shelf many of the sandstone beds show characteristics of storm deposition (Cummings and Arnott 2005). The presence of thick, tide-influenced outer-shelf delta deposits lead to the interpretation (Cummings and Arnott 2005) that the first delta that arrived at the shelf edge prograded into a bathymetric depression (possibly the head of a slope canyon) that sheltered storm-wave energy and amplified tidal currents. Storm-wave-dominated (deltaic) deposition at the shelf edge turned into strongly tide-influenced (estuarine) sedimentation during subsequent transgression (Cummings and Arnott 2005).

The growth-fault model presented here will have only minor applicability to depositional systems in which growth-fault-related subsidence has a minor contribution compared to regional relative sea-level changes, or in which other features such as diapirs strongly affect local basin configuration (Kindiger 1989; Cummings and Arnott 2005).

**GROWTH FAULTS AND REGRESSIVE-TRANSGLSSIVE CYCLES**

This study describes growth-fault influence on 2-D regressive delta deposits, but natural regressive-transgressive (R-T) systems are influenced by additional complicating factors than those considered in our simple numerical model. Growth-fault-related stratal geometries have been previously described (Douit and Omatsola 1989; Edwards 1995; Rouby et al. 2011), but little is known about other effects and interactions of growth faults with depositional systems over longer time scales.
Autocyclic changes within the deltaic system (Roberts 1997; Correggiari et al. 2005) lead to considerable spatial and temporal facies variability at short time intervals in many modern and ancient deltas (Ainsworth et al. 2011; Olariu et al. 2012a; Olariu 2014). There is a long-term coupling between fault activity and sediment loading (Cartwright et al. 1998), resulting in active growth faults during peak sedimentation rates and periods of fault inactivity when sediment accumulation is modest. Periodicity in cycles of active fault growth of 40–50 kyr has been linked to cyclicity in sedimentation history (regressive-transgressive cycles) (Cartwright et al. 1998). Therefore growth faults are thought to...
be active mostly during regression. It is thought that during transgression, growth-fault movement is minimal (Bruce 1972; Cartwright et al. 1998), because the sediment load that activates it is displaced landward. If growth-fault subsidence is efficient enough to slow delta progradation and to induce transgression, a local barrier island-lagoon system is formed that will migrate landward. Minor fault movement during transgression will accentuate the irregularity of the coast and favor tidal processes, as has been documented on the Jurassic Nova Scotia margin (Cummings and Arnott 2005). However, during the subsequent highstand, renewed progradation will push more sediment basinward and thus drive loading of the margin and reactivation of growth faults. Significant growth-fault scarps (meters to tens of meters of relief) filled with mud might develop on the outer shelf during highstand when a weak shelf layer (deposited during transgression of the shoreline) is subjected to destabilizing gravitational stress (Edwards 2000).

Over a longer time interval, the subsidence pattern of large growth faults (northern Gulf of Mexico, Niger Delta) might interfere with relative sea level, thus affecting shoreline processes. If the fault-related subsidence is high, comparable with eustatic and regional tectonic movements, then local changes in shoreline processes (as suggested by the present model) or R-T shoreline trends will be observed. When growth faults are active (or reactivated) during regression, wave dominance of the delta would be expected during multiple R-T cycles, as was observed in the Oligocene Frio Formation (Olaru et al. 2013). However, at times when the growth-fault subsidence is negligible relative to eustatic and regional tectonic movements the R-T shorelines and delta processes will not be particularly affected.

Future studies may address additional complications that arise as a result of growth faults affecting deltas, such as the effects of: (1) 3-D variability (along strike) and geometry of the growth faults, (2) multiple growth faults affecting the same delta complex, and (3) growth faults that control the avulsion of fluvial and delta distributary channels.

CONCLUSIONS

The common occurrence of wave-dominated deltas in growth-faulted depocenters is explained with reference to a simple numerical model by (1) increased local accommodation due to subsidence in the downthrown compartments, which is also enhanced by (2) a stronger wave regime on the outer shelf due to unrestricted fetch and water depth. The expansion index (EI) is proposed as a proxy for the degree of wave dominance. As EI is increased, more sediment is trapped into growth-faulted depocenters and less sediment is delivered to the shoreline. However, additional quantitative studies are needed to establish threshold values of EI that are associated with changes in depositional process regime. The dominance of wave processes in delta deposits of growth-faulted depocenters is counterintuitive because of the high sediment supply needed to load the mobile substrate and to initiate growth-fault movement. Most likely delta landward of the fault are river or tide-dominated whereas the deltas in the downthrown compartments might be initially tide or river-dominated, but will more likely become wave-dominated as the deltaic shoreline progrades.

Wave dominance in the deltas in outer-shelf growth compartments implies that building of the upper slope through aggradation due to sustained fault subsidence causes a rising shoreline trajectory and decreases the likelihood of the presence of the deep-water fans unless canyons incise the shelf edge and allow bypass of sediment to deep water.

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