Greenhouse shoreline migration: Wilcox deltas

Jinyu Zhang, Ronald Steel, and William Ambrose

ABSTRACT

In contrast to the high-frequency and high-amplitude sea level changes of icehouse times, eustatic sea level changes in greenhouse times are now generally accepted as significantly lower frequency and amplitude. As a corollary of this, frequency and extent of cross-shelf shoreline transits in greenhouse times are also likely to have been modest by comparison, and it has been suggested that greenhouse deltas may have been docked at the shelf edge for long periods, thus delivering sediment to deep-water areas more frequently. A revisit of upper Paleocene–lower Eocene Wilcox data across south Texas shows repeated regressive–transgressive shoreline migrations longer than 50 km (31 mi) at a time scale of some 300 k.y. This style of repeated shoreline transits is documented from well logs and is supported by the repeated presence of transgressive estuarine deposits with strong tidal evidence as interpreted from core. We argue, therefore, that Wilcox paleogeography was more varied than commonly portrayed and that the greenhouse shoreline transits were caused by greenhouse sea level change but severely modulated by variable sediment discharge caused by Paleogene hyperthermals. Periodic climate warming during Wilcox deposition and Laramide relief generation in the drainage areas were also responsible for unusually high sediment flux into the Gulf of Mexico. The factor of sediment supply in shoreline growth and retreat has been understated in the literature, partly because of an overemphasis on accommodation as the main driver of stratigraphic sequences.

INTRODUCTION

Controls on Shelf Building and Cross-Shelf Shoreline Migration

Conceptually, shoreline migration rate was controlled by accommodation to sediment supply ratio (Curtis, 1970; Swift and Thorne, 1991), modified by autogenic responses (Muto et al.,...
For any point on the preexisting shelf surface, the dimensionless equilibrium index \( E(d) \) is a function of relative sea-level change \( R \), sediment input \( Q \), and fluid power \( P_g(h) \) (equation 1) (Swift and Thorne, 1991).

\[
E(d) = \frac{(Q_g - kR)}{P_g(h)}
\]

where \( Q_g \) is sediment input of grain size \( g \), \( k \) is a dimensional constant of conversion relating length to mass, and \( P_g(h) \) is fluid power expended to disperse sediment of that grain size at the depth \( h \).

Relative sea-level change was influenced by global sea-level change, local and regional subsidence, and compaction. Sediment input was supplied by the rivers, but significant volumes were redistributed by longshore drift because storm waves were a common component of the regressive half cycles (Hovius, 1998; Carvajal and Steel, 2009). Fluid power was controlled by depositional process regime (river, wave, tidal currents, and gravity flow) and water depth. A change in any of these parameters would have influenced the shoreline behavior on the shelf. For example, increased sediment supply while keeping all other factors constant would have resulted in a shallower and more rapidly regressive shoreline. The autogenic responses, which are generated by self-organization within the deltaic system itself while supply and accommodation are steady, modify the shoreline migration at a smaller time scale (hundreds to tens of thousands of years). Autoretreat predicts that any prograding shoreline with constant sediment supply will inevitably retreat landward as the finite sediment supply fails to fill increased accommodation (Muto and Steel, 1997). Therefore, deposition and erosion on the preexisting shelf and regression or transgression of the shorelines transiting that shelf were mainly controlled by unsteady and steady rates of global sea-level change, local and regional subsidence, and sediment supply, resulting in allogenic and autogenic stratigraphic responses, respectively (Swift and Thorne, 1991; Muto and Steel, 2002; Kim et al., 2006; Sømme et al., 2009a).

**Purpose of this Study**

Shoreline migration is a well-studied phenomenon, examined mostly from observations of modern and Holocene systems as well as from numerical and physical experiments. Quaternary icehouse shoreline migration and its linkage of sediment supply and accommodation change over time scales of \( 10^5 \)–\( 10^6 \) yr are well studied, benefiting from available high-resolution data (e.g., eustatic, chronostratigraphic, and seismic data sets). In the case of the northern Gulf of Mexico, late Quaternary shoreline migration, including long-distance regressive and transgressive delta lobe cross-shelf transits, has been well reconstructed and thought to have been driven mainly by glacioeustatic sea level oscillations (Anderson et al., 2004). Analogous flume experiments using...
constant sediment supply and a linear hinge-type subsidence profile with constant subsidence rate created deltaic shoreline migration where the primary control was base level change, followed by sediment supply to the delta lobe foresets, geometry of the foresets, and the average subsidence rate over the foresets, in order of importance (Kim et al., 2006). Another experimental case, using a similar approach, except employing variable water discharge at higher frequency than that of base level change, strongly suggested that sea level remains the dominant control on large-scale stratigraphic architecture of deltas (Bijkerk et al., 2014). However, the internal geometry of this large-scale stratal configuration was influenced by water discharge because of its effect on the timing of sediment transport onto the delta (Bijkerk et al., 2014).

Unsolved problems remain regarding shoreline migration. First, our understanding of how different amplitudes of sea-level change (e.g., icehouse vs. greenhouse) influence shoreline migration is still limited. Blum and Hattier-Womack (2009) suggested that the long-distance cross-shelf shoreline migration observed in icehouse stratigraphy would likely be severely limited in greenhouse cases because of the lower amplitudes of sea-level change. However, although it is obvious that greenhouse shelf edge deltas are likely to reside longer at the shelf edge than icehouse ones (greenhouse cycle frequency is longer; Miller et al., 2005a, b), there are only limited data to support the suggestion that greenhouse deltas would have been locked at the shelf edge in greenhouse periods, and there are some studies that are inconsistent with this idea (e.g., Carvajal and Steel, 2006; Gomez-Veroiza and Steel, 2010). The Wilcox case discussed herein also suggests otherwise. Second, the factor of sediment supply during high-frequency sea level change may be underestimated (Somme et al., 2009a). Most studies assume constant sediment supply, but sediment flux to the basin is well known to change over a range of time scales. Although tectonic activity affects catchment area and relief generation, both of which affect sediment supply (e.g., Vakarelov et al., 2006; Carvajal and Steel, 2012), sediment flux is also known to be climate driven, and some of the best-known cases show significant change with even modest temperature and humidity variation, over time scales shorter than 5000 yr (Goodbred and Kuehl, 2000a, b).

We have chosen here to take the upper Paleocene–lower Eocene Wilcox Group of onshore Texas to document greenhouse shoreline behavior. One of the reasons to pick this system as an example is that the shelf of the northern Gulf of Mexico was relatively wide (>100 km [62 mi]) already by late Paleocene (Galloway et al., 2011) so that Cenozoic cross-shelf, shoreline transits should be observable if they existed. The purpose for the study is threefold:

1. Provide some clarity on the Wilcox cross-shelf shoreline migrations in view of its greenhouse setting, not only on the million-year time scale of the thick, key shale horizons that conventionally subdivide the Wilcox stratigraphy but also on a higher-frequency time scale consistent with cross-shelf progradation time scales for delta systems. Thirty-seven widespread shelf transit sequences have now been identified in the studied Wilcox succession.

2. Document a broader range of depositional process regimes for the Wilcox delivery system on the Texas shelf than has previously been emphasized, particularly the importance of tidal systems in the frequently developed transgressive estuaries. This also contributes to a better understanding of sand delivery across the Wilcox shelf edge to the deep-water area.

3. Argue that climate warming on the time scale of well-documented Paleocene–Eocene hyperthermals is also likely to have been a very strong influence on Wilcox regressive–transgressive sequences, modulating the eustatic sea level changes. The general temperature increase during the combined period of hyperthermals is also likely to have driven the very large sediment discharge volumes from the Texas shelf to the deep-water Gulf of Mexico.

**Wilcox System**

The late Paleocene–early Eocene Wilcox dispersal system of the Gulf of Mexico was remarkably sand-rich both on its coastal plain and shelf reaches as well as on its deep-water slope and basin floor (Galloway et al., 2000, 2011; Zarra, 2007). The deep-water sandy fans, located up to 600 km (373 mi) beyond their paleo-shelf edge were supplied by fluvial and wave-dominated deltaic and tidal estuarine delivery systems with switches of depocenter between Rio Grande, Houston, and Mississippi embayments (Figure 1) (Galloway et al., 2011; Fulthorpe et al., 2014).
Deep-water Wilcox deposits in the Gulf of Mexico are reported to be not only thick (up to 1829 m [6000 ft]) but also really extensive with greater than 104,000 km$^2$ (40,155 mi$^2$) (Meyer et al., 2005; Zarra, 2007). The huge sand volumes indicate that the system was unusually well supplied, especially initially (Zarra, 2007). The sediment volume delivery rate decreased from 150,909 to 72,708 and to 25,529 km$^3$/m.y. (36,205 to 17,444 and to 6125 mi$^3$/m.y.) from lower Wilcox to middle Wilcox and to upper Wilcox, respectively (Galloway et al., 2011), as also confirmed by decreased shelf margin progradation rates: 20–30 km/m.y. (12–19 mi/m.y.) in lower Wilcox and 4–8 km/m.y. (2–5 mi/m.y.) in middle and upper Wilcox (Carvajal et al., 2009). The large Wilcox sediment supply is partly caused by a dramatic reorganization of provenance and catchment systems from Late Cretaceous to Paleogene (Blum and Pecha, 2014). Wilcox drainage area has been calculated to greater than 1,000,000 km$^2$ (386,102 mi$^2$) with supply river channels greater than 2000 km (1243 mi) in length (Sweet and Blum, 2011) based on detrital zircon studies.

**Previous Studies on Wilcox Shoreline Migrations**

Numerous studies conducted on the upper Paleocene–lower Eocene Wilcox system provide information about the distribution of lithofacies, depositional environment, and stratigraphic framework (Fisher and McGowen, 1967; Fisher, 1969; Edwards, 1980, 1981; Bebout et al., 1982; Hamlin, 1988; Miller, 1989; Xue and Galloway, 1995; Breyer, 1997; Xue, 1997; Crabaugh and Elsa, 2000; Galloway et al., 2000; Crabaugh, 2001; Galloway, 2001; Hargis, 2009; Yancey et al., 2010). Most of these studies focused on third-order stratigraphic sequences but paid little attention to the fundamental shelf transit time scale (a few hundred thousand years) for most shelf sediment delivery systems (Burgess and Hovius, 1998; Muto and Steel, 2002). There have also been some local studies that have provided data suggesting that long-distance transgressive shelf transits have been important. For example, the Wilcox outcrop trend (>100 km [62 mi] inland from the shelf margin) was originally interpreted as dominated by fluvial deposits, especially for lower Wilcox equivalent. However, some newly discovered outcrops and reinterpretation of old outcrops (e.g., Indio and Carrizo Formations at Bee Bluff in Zavala County documented by Breyer [1997], Calvert Bluff and Carrizo Formations at Big Brown mines in Freestone County documented by Sturdy [2006], and Calvert Bluff and Carrizo Formations at Red Bluff in Bastrop County documented by Yancey et al. [2010]) show there were clear marine (wave and tide) shorelines.
in Indio, Calvert Bluff, and Carrizo Formations (middle and upper Wilcox equivalents). These outcrop observations indicate that marine transgressions extending far back inland across the shelf were common.

Recently, the debate between Rosenfield and Pindall (2003) and Sweet and Blum (2011) provides two quite different models for the Wilcox delivery system: Paleocene–Eocene isolation of the Gulf of Mexico and lowstand drawdown of the delivery system, and greenhouse shelf edge docking because of limited eustatic sea level amplitudes. These two models have dramatically different shoreline migration scenarios.

Rosenfield and Pindall (2003) proposed that sea level was drawn down when the Gulf of Mexico became isolated from the world’s ocean, similar to the Messinian isolation and drawdown of the Mediterranean. As a result of this isolation, they related the transgressive–regressive cycles on the Gulf shelf to potential drawdown–flood cycles including lower Wilcox, middle Wilcox, and upper Wilcox cycles. They argued that large amounts of sediment bypassed the paleoshelf and created deltaic deposition on the paleoslope during the drawdown period (third-order lowstand) (Rosenfield and Pindall, 2003, p. 100, point 6). Therefore, they implied that the shoreline may have stayed below the paleo–shelf edge for most of the third-order sequence.

Sweet and Blum (2011) argued against the drawdown model and proposed that the deep-water Wilcox deposits were simply supplied by a large river system in a greenhouse world. Based on recent zircon-based provenance studies (Mackey et al., 2012; also see Blum and Pecha, 2014), they argued that the Wilcox system was associated with a very large drainage area that created a huge sediment supply. Inputting Wilcox parameters to the source-to-sink morphology relationships of Sømme et al. (2009b), they also suggested that the size of Wilcox Yoakum Canyon in Gonzales and Lavaca Counties was consistent with the large volume of Wilcox deep-water deposits. More importantly, Sweet and Blum (2011) proposed that the great thickness and high net to gross character of Wilcox deep-water sediments was caused by only modest shelf transiting of the deltas and that the Wilcox river mouth supply may have remained docked at the shelf margin or in canyon heads, thus constantly supplying sediment to deep-water fans.

Therefore, three framework possibilities exist for how Wilcox shorelines behaved:

1. Shorelines stayed mainly below the paleo–shelf edge during lowstand drawdown periods in the Gulf of Mexico with landward transiting of the delivery system mainly during floodback intervals represented by the thick, third-order time scale (1–5 m.y.) shale intervals.
2. Shorelines remained docked close to the shelf margin during most of Wilcox shelf growth, and the back-and-forth, cross-shelf transits of the delivery system were minimal as a result of constant very large sediment supply and minor greenhouse sea-level change.
3. Cross-shelf shoreline sediment delivery transits of 50+ km (31+ mi) were fundamental to the growth of the Wilcox shelf margin prism, on a time scale of a few hundred thousand years; that is, despite the limited greenhouse sea level amplitudes, the Texas coastal plain had a low enough gradient to allow frequent long-distance shoreline transgressions, such as those that happened during the Holocene and as is quite evident from Wilcox transgressive shoreline deposits near Austin, more than 50 km (31 mi) from the coeval shelf edge.

**METHODOLOGY**

A high-frequency sequence stratigraphic framework was created for the Wilcox succession based on 351 well logs (gamma-ray or spontaneous potential logs), supplemented by core descriptions from 10 wells (Figure 1). Flooding surface–bounded regressive-to-transgressive sequences (Galloway, 1989) or clinothems (Rich, 1951) are the key building blocks of Wilcox clastic wedges and are used here to subdivide the Wilcox stratigraphy. Within each sequence, one regressive half-cycle and one overlying transgressive half-cycle are defined on the basis of their log stacking pattern, calibrated by core data, and these tend to show overall coarsening-upward and fining-upward trends, respectively (Figures 2, 3). Well-known key intervals or surfaces (e.g., Big shale, boundary between lower and middle Wilcox, and Yoakum shale, boundary between middle and upper Wilcox) have been correlated to type logs of previous publications (Hargis, 2009; Bebout
Figure 2. (A) Examples of interpretation of facies associations based on spontaneous (SP) and gamma ray (GR) logs, and calibrated by cores (see also Figures 3, 4 for core information). (B) Descriptions and interpretations of regressive and transgressive log motifs for each facies association. MFS = maximum flooding surface; MRS = maximum regressive surface; R = regression; T = transgression.
et al., 1982) to ensure consistent results. The wells are closely spaced (~3 km [2 mi]), also adding to correlation confidence.

First, cores were described and interpreted in terms of regressive or transgressive trends (shallowing and deepening) of depositional systems, then these were calibrated to the well logs to extend the interpretation of lithology and depositional environments to uncored intervals. Log pattern (blocky, serrated, bell, and funnel), contacts (sharp or gradational), scale (thickness), sandstone content, vertical grain size trend (coarsening- or fining-upward), and spatial relationship between each facies association were documented (Figure 2). Based on these log characteristics and core descriptions, seven main facies associations were distinguished for Wilcox strata (Figures 2–4). The transgressive facies associations are dominated by (1) erosively based fluvial and fluvial–tidal distributary channel deposits (commonly inner estuary); (2) channelized, stacked sets of mud-draped cross-strata interpreted as tidal bar deposits of middle to outer estuary areas; and (3) heterolithic, muddy, and bioturbated intertidal to supratidal flat deposits (including coals) interpreted as estuary margin deposits. The regressive facies associations include (1) sandy, hummocky, or swaley delta front or shoreface deposits; (2) sharp-based, unbioturbated, channellized deposits interpreted as delta distributary channels; and (3) prodelta or shelf deposits. Muddy, heterolithic interdistributary bay deposits (where there is little or no active river input) are preserved in both transgressive and regressive half-cycles, though more prominently in the transgressive part. The log motifs, cores, and facies associations are thus helpful not only to define the depositional environments but also to locate the shoreline positions within the stratigraphic framework.

**Figure 3.** (A) Type log through the Wilcox Group in Bee and Karnes Counties illustrating some 37 regressive–transgressive sequences, on average 32 m thick. (B) Details of stratigraphic sequences 28 to 35 based on gamma ray (GR) logs (see interval location in Figure 3A). (C) Cored interval details of stratigraphic sequences 19–25 (see interval locations in Figure 3A). F = fine; M = medium; MFS = maximum flooding surface; MRS = maximum regressive surface; Mu = mud; Si = silt; Vf = very fine.
DESCRIPTION AND INTERPRETATION OF TRANSgressive–REGRESSive CYCLES

Process Variation Seen within Transgressive–Regressive Cycles

Thirty-seven high-frequency sequences in the Wilcox Group are documented in this study (Figure 3). This is a minimum number because the number of muddy sequences at the base of the lower Wilcox is likely to have been underestimated. The duration of the entire Wilcox depositional interval is estimated to be approximately 12 m.y. (Crabaugh, 2001; Galloway et al., 2011) or 8 m.y. (Zarra, 2007); therefore, the average maximum duration of a shelf transit sequence is approximately 220 to 320 k.y. This duration is a maximum estimation because (1) the number of documented sequence may be underestimated as the reason we stated above and (2) the stratigraphy could be incompletely preserved as the presence of large-scale erosion. We refer to the stratigraphy of these as fourth-order stratigraphic sequences, but their real importance is that they represent the fundamental cross-shelf sequences, i.e., the basic building blocks of the Wilcox shelf margin prism.

Commonly, a marked difference in the dominant process regime exists between regressive and transgressive half-cycles of Wilcox sequences. Although most of the systems are mixed energy (river, wave, and tide), it is clear that tidal processes are more important during the transgressive shoreline phases, whereas wave and river processes are dominant during regression (see also Porebski and Steel, 2006; Yoshida et al., 2007). These trends in the Wilcox system result largely from local enhancement of tidal current strength and tidal range as the oceanic tidal wave enters the embayments and estuaries of the transgressing shoreline, in contrast to the frequent dominance of waves on open-ocean regressive coastlines. These regime differences, despite an overall wave–tide–fluvial energy mix, are clearly shown from Wilcox cores, which are also critical in demonstrating the change from shoreface and delta shoreline types during regression, versus estuary and barrier lagoon environments during transgression (Figure 4).

Regressive Half-Cycle

Description—The regressive half-cycles are from 10 to 60 m (33 to 197 ft) thick, changing upward from muddy and thin-beded heterolithic deposits to sand-rich, rippled, and stratified deposits, sometimes capped by channel fill deposits. The basal mudstones and heterolithic interval contain minor sandy laminae with starved ripples (sometimes wave or combined flow ripples but commonly unidirectional) and graded or structureless sandy layers (commonly <5 cm [2 in.] thick) (Figure 5A). Some of the sandy laminae are
Figure 5. Facies of the regressive half-cycles: (A) Mud-dominated heterolithic interval with moderate bioturbation and lower-flow regime sedimentary structures, indicating fair-weather (FW) periods. (B) Alternating FW beds and storm event beds. Note that vertical burrows dominate in event beds, whereas horizontal burrows are more common in FW beds. (C) Clean, very fine-grained sandstone, interpreted as hummocky cross-stratification, overlain by deformed siltstones. (D) Microfaults within siltstone and mudstone. (E) Sandy slumps. (F) Clean, very fine-grained sandstone with low-angle laminations, interpreted as hummocky cross-strata. Ch = Chondrites; HCS = hummocky cross-stratification; Phy = Phycodes; Pl = Planolites; Ro = Rosselia; SSDS = soft-sediment deformation structure; Te = Teichichnus; Th = Thalassinoides.
masked by moderate bioturbation or by soft sediment deformation and microfaults (Figure 5C–E). The upper sand-rich deposits (up to 25 m [82 ft] thick) are upward-coarsening and composed of alternating fine- to very fine-grained hummocky or swaley stratified beds and muddier interbeds of bioturbated siltstones to very fine sandstones with wave or combined flow ripples (Figure 5B, F). Thin (up to 1 m [3 ft]) sandstone-dominated heterolithics with double mud drapes are also observed in these successions. The occasional capping channel fill deposits show basal mud clasts and an upward-finishing pattern from a few to 10 m (33 ft) thick. These sandy units with erosional bases are composed of fine- to medium-grained, plane-parallel, or cross-stratified sandstone. The upper parts of channel fill units are very fine-grained sandstone with stressed bioturbation with diminutive *Paleophycus, Chondrites, Phycodes, Teichichmus*, and *Planolites* (<2 cm [1 in.] long) (Figure 5A, B). Some channel fill successions show multiple channels.

**Interpretation**—These tens of meters thick coarsening-upward successions reflect a progradational and basinward-finishing regressive half-cycle generated by the basinward migration of a delta or delta lobe. Wave and storm dominated conditions on the front of the delta are evidenced by wave ripples and hummocky cross-stratified event beds (Figure 5C, F). The moderate to pervasive bioturbation formed during periods of fair weather (Figure 5A, B). Although storm wave processes dominated in the regressive half-cycle, they were not the only processes present, as in most modern systems (e.g., Danube delta; Olariu, 2014). Evidence for tidal currents (e.g., double mud drapes) is also observed, and sediment was supplied to the system by fluvial currents, especially during periods of flooding. Soft sediment deformation is possibly caused not only by storm events but also by river floods (Figure 5C, E). The distributary channels that supplied sediment to the delta front eventually cut down and became infilled during river flood season.

**Transgressive Half-Cycle**

**Description**—The upward-finishing half-cycles in the upper part of Wilcox sequences range from 1 to 90 m (3 to 295 ft) in thickness. The deposits change upward within the transgressive interval from (1) deeply downcutting (up to 10 m [33 ft]) channels with fine- to medium-grained structureless sandstones showing common mudstone drapes, some rhythmite layering (Figure 6D), abundant mud clasts (>5 cm [2 in.] in length), and much organic matter to (2) stacked sets of ripple-laminated or cross-stratified, fine-to very fine-grained sandstone units up to 20 m (66 ft) thick (Figure 6A, B, E) showing some marine bioturbation, common mudstone drapes, and thicker deformed muddy beds, as well as associated muddier heterolithic units. The sandy succession then passes upward to (3) muddier ripple-laminated (sometimes bidirectional) successions that are heterolithic and become increasingly marine bioturbated upward. These muddy heterolithic units are then commonly capped by (4) faintly laminated to structureless (in distal reaches of the sequence) or coaly mudstones (in proximal reaches), denoting the top of the transgressive unit and the base of the overlying regressive unit. All of the above intervals are associated with abundant organic matter. Diminutive trace fossils, mostly *Planolites, Paleoophyces*, and *Chondrites*, are present in muddy heterolithics or uppermost mudstones. Small-scale soft sediment deformations are also present in muddy heterolithics. Wavy or combined ripple laminations are locally present in the upper half of the interval but are sometimes destroyed by bioturbation.

**Interpretation**—These fining-upward successions are interpreted as landward-stepping estuarine deposits as suggested by the following:

- The basal channels deposits, notably in their erosive behavior and their mud drapes, indicate strong river flux and significant tide influence in the interflood periods.
- In contrast, the uppermost muddy and heterolithic deposits are commonly thoroughly bioturbated, suggesting a more open marine coastal site. It is this vertical trend from a more proximal to an overlying more distal association that is key to interpreting the transgressive character of the upward-finishing half-cycle (Cattaneo and Steel, 2003).
- Transgressive coasts can be either estuaries (with significant river supply from proximal end) or barrier and lagoon depositional systems (without significant
Figure 6. Facies of the transgressive half-cycles and evidence of the dominance of tidal processes in transgressive half-cycles. (A) Bidirectional rippled very fine-grained sandstone, (B) bidirectional rippled very fine-grained sandstone, (C) double-mud-draped fine-grained sandstone, (D) rhythmtes, (E) sandy heterolithics with bidirectional ripples and erosional surface, (F) mud-draped fine-grained sandstone, and (G) tidal bundles.
river inflow), with neighboring open coast tidal flats and marshes along strike (Boyd, 2010). Both of these types of transgressive coast are likely to be present in Wilcox transgressive facies tracts, but we suggest the estuaries were likely more common because of abundant evidence of strong river input along most of the Wilcox coastlines.

- In cases where the transgressive interval is relatively thick and sand-rich and where the sand bodies filling the channels were tidally influenced (mud drapes, rhythmic bedding, stacked cross-stratified sets, and bidirectional paleocurrents [Figure 6]), it is likely that the estuaries were tide dominated with rapidly moving subtidal compound dunes and tidal bars, although the wave indicators near the top of some units suggest mixed tide and wave influence (see also Dalrymple et al., 1992).

- The heterolithic and very bioturbated intervals with capping coaly mudstones near the top of some transgressive facies tracts are likely to be intertidal (bioturbated and rippled heterolithics) to organic-rich (marsh grass) supratidal flats along the margins of the estuary (Dalrymple and Choi, 2007). Despite transgression, rapid accretion and infilling at the estuary margins commonly occur, causing subtidal channels with subaqueous dunes to be overlain by intertidal flats and then by muddy supratidal deposits, as local areas of the estuary become infilled.

- The larger-scale upward fining grain size pattern seen in the transgressive half-cycle (up to 90 m [295 ft] thick) is contained within a very broad (kilometers), relatively shallow valley, infilling with transgressing coastal estuaries. The documented upward trend results from a retrogradational, coastal stacking pattern, so that the mid-outer estuary tidal bars come to overlie the inner estuary fluvial distributary channels, and they themselves become covered by muddy estuary margin, tidal flat deposits, or eventually open marine mudstones as the transgression proceeds and the open marine shelf covers the region. The tidal processes within estuaries tend to be very strong and able to rework and redeposit both river-derived and marine deposits during transgression, and for this reason, there is an abundance of tidal deposits preserved. Therefore, the upward fining grain size pattern is caused directly by neither the upward fining of point bars nor the upward fining of laterally accreting tidal flats, although it may well include these smaller-scale elements.

### Individual Regressive–Transgressive Cycles

The Wilcox high-frequency sequences average 32 m (105 ft) in thickness but reach up to 94 m (308 ft) thick in places (Figures 3, 7, Table 1). This is a normal thickness for shelf transit sequences seen elsewhere (Steel et al., 2008) and represents the undecompressed height of the delta plus overlying estuary or barrier lagoon system. Sequences are bounded by muddy maximum flooding intervals, and the muddy surface with the highest gamma-ray reading inside these intervals is assigned sequence boundary status in the manner outlined by Galloway (1989). In subdividing the succession in this way the sequences take a regressive-to-transgressive character, contrary to the order and succession as prescribed within the conventional sequence stratigraphy. The regressive half-cycle (shelf and prodelta up to delta front deposits) is locally capped by a supply distributary channel, but this channel commonly does not accumulate much sediment during regression, because most sediment is bypassed to the shoreline. If deeply eroded during flooding, this channel then becomes infilled later during transgression as part of the overlying transgressive estuary (Figure 4). The base of these channels or the top of the regressive delta front is commonly the maximum regressive surface of the sequence. The transgressive half-cycle (estuarine channelized bars, intertidal to supratidal estuary margin deposits, interdistributary tidal embayment deposits, and eventually marine shelf deposits) shows an irregular upward fining from distributary channels up through muddy embayment deposits and intertidal or supratidal (coaly) deposits, eventually to open-marine shelf mudstones. This transition from regressive half-cycle to transgressive half-cycle is easily recognized by the grain size trend of well logs and cores (Figure 3B, C). Regressive half-cycles are mostly less than 30 m (98 ft) thick proximally and greater than 60 m (197 ft) thick distally, across the preexisting shelf (Table 1). The regressive unit thickens basinward, as the delta regressed from the shallower waters of the inner shelf (50 m [164 ft]) to the slightly deeper water (100 m+ [328 ft]) of the outer shelf (Figures 4, 7B). The transgressive half-cycle tends to thicken from zero at the shelfward turnaround from regression to transgression and reaches greater than 30 m (98 ft) thick when developed within an estuarine valley (Figures 4, 7B). The
Figure 7. (A) High-frequency sequence stratigraphic framework of the Wilcox Group showing shelf edge and points of maximum shoreline regression and two-dimensional geometry of a single high-frequency sequence. (B) Transgressive and regressive cycles within high-frequency sequence stratigraphic framework. (C) Illustrations of channel sand bodies on the shelf and slope. A = aggradation thickness; GR = gamma ray; MFS = maximum flooding surface; MRS = maximum regressive surface; P = progradation distance; SE = shelf edge; SP = spontaneous; X = sequence number.
Table 1. Transgressive–Regressive Unit Thickness and Sandstone Percentage from Proximal to Distal Locations through 37 Shelf Transit Sequences of the Wilcox Group

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Italicized intervals are outer shelf to slope deposits.
Abbreviations: R = regressive half-cycle; T = transgressive half-cycle.

Greenhouse Shoreline Migration: Wilcox Deltas
thickening landward is caused not only by the estuary containment during transgression but also by the landward rise of the estuary shoreline trajectory and the underlying lower delta plain deposits, particularly if there was continuous river supply during retrogradation (Table 1). As a result of the oppositely directed thickening and thinning of the two parts of the shelf sequence (transgressive half-cycle thickens landward, whereas regressive half-cycle thickens basinward), each combined sequence maintains a fairly sheetlike geometry from distal to proximal before onlapping the steeper hinterland slopes (see also Siggerud et al., 2000; Cattaneo and Steel, 2003). It should be noted that the transgressive half-cycle, despite its capping of muddy marine shelf deposits, is commonly more sand-rich than the muddier regressive half-cycle (Table 1) (see also Chen et al., 2014).

**CROSS-SHELF DELTA–ESTUARY MIGRATION AND SHELF BUILDING OF THE WILCOX GROUP**

**Fourth-Order Sequence Stratigraphic Framework**

A fourth-order sequence stratigraphic framework was built based on the individual transgressive–regressive cycle analysis (see Process Variation Seen within Transgressive–Regressive Cycles section). Thirty-seven cross-shelf sequences are recognized across the study area (Figure 7). Lower, middle, and upper Wilcox lithostratigraphic members were also added to this framework with the picking and recognition of Big shale and Yoakum shale, and these can also be correlated to the type logs of previous publications (Bebout et al., 1982; Hargis, 2009).

The rationale for the importance of fourth-order stratigraphic sequences derives from numerical experiments done by Burgess and Hovius (1998) and by Muto and Steel (2002). These authors selected 29 medium- and large-sized, modern river deltas, and by inputting shelf width and gradient, coastal plain gradient, river discharge, and delta front slope, they computed the time needed for each of these modern deltas to reach their shelf edge under conditions of rising, stable, and falling sea level. This simple modeling provided a minimum estimate of time required for the cross-shelf regressive transit of each delta, under ideal river-driven conditions. The most interesting result of the analysis was that even with moderately rising sea level (2.1 m/k.y. [6.9 ft/k.y.]), two-thirds of the deltas were able to reach their shelf edges (and so to bypass sediment onto the deep-water slope) and that the transit time for these deltas was rarely more than 100 k.y. The conclusion was that the combined regressive and transgressive shelf transit by deltas should take less than 150 k.y. This has some significance for sequence stratigraphy, implying that these high-frequency transits are the fundamental stratigraphic sequences and not third order as originally weighted by Exxon. The fourth-order stratigraphic sequences are therefore believed to be the basic building blocks of shelf margin sedimentary prisms, such as the Wilcox clastic prisms.

**Shelf Edge Building and Shoreline Migration**

By definition, the shelf edge is the rollover area between the very gentle shelf (~0.1°) and the steeper continental slope (1–2°) within continental margin scale clinoforms (hundreds of meters to kilometers). The water depth at the average shelf edge is some 130 m (427 ft), and from there it deepens abruptly (Helland-Hansen et al., 2012). The shelf edge primarily migrates basinward, generally only a slight distance forward during each fourth-order sequence, beyond the preexisting shelf edge. Some rare examples exist of shelf edge retrogradation over long time scales, commonly where the supply system has been diverted elsewhere (see examples and discussions in Helland-Hansen et al., 2012). Relatively persistent shelf edge progradation is accompanied by inevitable bypass of sediment to deep-water slope and basin floor. Although this sediment bypass at any point in time is not uniform along shelf edges, especially because of long storm wave–dominated reaches that hinder direct bypass (Carvajal and Steel, 2009; Dixon et al., 2012) and because of significant strike distance between adjacent shelf edge supply points, time-averaged sediment bypass over geologic time scales generally ensures forward movement of the entire shelf margin clinoform, including the bottomsets (Petter et al., 2013). Although it has been argued that shelf-indented canyons (e.g., the Congo system; Eisma and Kalf, 1984) cause most modern highstand sediment bypass to slope and basin floor (Sømme...
et al., 2009a), this is not likely to be the case at longer time scales. The basin floor (clinoform bottomset) should not be considered a separate geomorphic feature but part of the larger clinoformed margin, or the integrated source-to-sink system.

The shelf edge can be recognized in stratigraphic cross sections by (1) the basinward transition from sandy, less than 100-m (328-ft)—scale R–T cycles with increasing degree of slumped deformation to muddy, larger-scale (hundreds of meters) cycles; (2) a change of stratal dip angle from the gentle shelf (<0.1°) to the steeper deep-water slope (1–2°); and (3) the associated occurrence of growth faults caused by the instability of shelf edge sand on slope mudstone across the shelf margin (Winker and Edwards, 1983).

Based on this understanding of shelf edges, the shelf edge positions of 37 fourth-order sequences of the Wilcox Group in the study area were documented on the depositional dip cross section of Figure 7. Progradation and aggradation dimensions of individual clinoforms, as well as rates of progradation and aggradation, are summarized in Table 2. The latter rates are estimated by using a Wilcox fourth-order sequence duration of some 300 k.y. The results show the shelf edge progradational distance between successive sequences is from 0 to 27.9 km (17.3 mi), with progradational rates from 0 to 86.1 km/m.y. (53.5 mi/m.y.); the aggradational thickness of each fourth-order sequence ranges from 7.6 to 73.9 m (24.9 to 242.5 ft), with aggradation rates ranging from 23.5 to 228.0 m/m.y. (77.1 to 748.0 ft/m.y.). Out of 26 fourth-order sequences, 7 are estimated to have had a progradation rate above 10 km/m.y. (6 mi/m.y.). Most of the shelf edge trajectories are rising, showing aggradational to progradational clinoform style. The exception is from shelf edge 4 to shelf edge 5, where there appears to be a slightly downward regression (Figure 7).

This study highlights maximum regressive and maximum transgressive points in the stratigraphy to describe the shoreline migration (Figure 7). The shoreline moves basinward from maximum transgressive point to maximum regressive point and retreats landward from maximum regressive point to the next maximum transgressive point. Although there are yet higher-order transgressive events within regressive half-cycles and regressive events within transgressive half-cycles, they are mainly of more local extent. Maximum regressive point is located by the most basinward extending regressive half-cycle. From this point, the deltaic deposits changed to become thick, muddy slope deposits. Maximum transgressive point is located by the maximum landward extent of marine muddy deposits and their brackish water landward equivalent. From this point, purely transgressive deposits change to regressive deposits upward.

The shoreline regressive and transgressive distance and rates from Figure 7 are summarized in Table 2. Nearly all of the fourth-order sequences extend out of the study area in the dip direction; therefore, their shoreline migration distance and rate are minimum estimates. Because the first three fourth-order sequences show only small amounts of outer shelf deposits in the study area, they are not taken into consideration. Within fourth-order sequences 4–30, the minimum shoreline regressive or transgressive distance is from 0.3 to 51.4 km (0.2 to 32.0 mi) with average 28.5 km (17.7 mi). Assuming the duration of regression and transgression within each fourth-order sequence is the same, 162 k.y., the minimal shoreline regressive or transgressive rate is from 2.1 to 317.9 km/m.y. (1.3 to 197.5 mi/m.y.) with an average of 176.1 km/m.y. (109.4 mi/m.y.).

**DISCUSSION**

**Evidence of Repeated Long-Distance Cross-Shelf Shoreline Migration on the Wilcox Shelf**

The interpretation and model of the Wilcox system presented above provide clear evidence to support the notion of repeated long-distance cross-shelf shoreline migration in the Wilcox Group despite greenhouse conditions with only a few tens of meters of repeated eustatic sea level change. The implication here is that the topsets of the Wilcox shelf margin prism, which extended at least as far north as the current Wilcox outcrop belt across Texas, do not consist only of delta plain and alluvial plain deposits but also contain significant swathes of regressive and transgressive marine shoreline and shelf deposits, not unlike the Quaternary stratigraphic record of the Gulf of Mexico shelf (Anderson et al., 2004).

Key points in the argument are as follows.
The transgressive–regressive cycles, on a thickness scale of normal shelf deltas (30–80 m [98–262 ft]), are widespread and easily recognized between closely spaced wells in both depositional dip and strike cross sections. This is what provides the tramline images on shelf margin topsets on seismic data (see also Sydow et al., 2003; Johannessen and Steel, 2005; Kertznus and Kneller, 2009; Gong et al., 2015, 2016). Note that the maximum flooding boundaries of these cross-shelf units are much more easily correlated than the depositional elements contained within these boundaries, where there are many autogenic responses in the stratigraphy.

An irregular but detectable tendency exists for the regressive half-cycle to thicken distally because of channel cannibalization and sediment bypass from the more proximal reaches of the shoreline transit to the distal reaches. Likewise, the transgressive half-cycle is thin in the topset distal reaches but tends to thicken irregularly into the proximal reaches.

### Table 2. Parameters of Shelf Edge Trajectory and Transgressive–Regressive Shoreline Trajectory

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<th>High-Frequency Sequence</th>
<th>Distance</th>
<th>Progradation Rate (km [mi]/m.y.)</th>
<th>Aggradation Rate (m [ft]/m.y.)</th>
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<th>Rate (km [mi]/m.y.)</th>
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Shelf edges of sequences 31–37 are out of the study area.

1. The transgressive–regressive cycles, on a thickness scale of normal shelf deltas (30–80 m [98–262 ft]), are widespread and easily recognized between closely spaced wells in both depositional dip and strike cross sections. This is what provides the tramline images on shelf margin topsets on seismic data (see also Sydow et al., 2003; Johannessen and Steel, 2005; Kertznus and Kneller, 2009; Gong et al., 2015, 2016). Note that the maximum flooding boundaries of these cross-shelf units are much more easily correlated than the depositional elements contained within these boundaries, where there are many autogenic responses in the stratigraphy.

2. An irregular but detectable tendency exists for the regressive half-cycle to thicken distally because of channel cannibalization and sediment bypass from the more proximal reaches of the shoreline transit to the distal reaches. Likewise, the transgressive half-cycle is thin in the topset distal reaches but tends to thicken irregularly into the proximal reaches.
because the earlier downcut distributary channels and channel belts then become backfilled with transgressive deposits in the estuaries (see also Steel et al., 2000). The landward extent of the transgressive deposits is commonly much greater than the extent of the previous regressive marine shoreline because transgression also involves a significant landward movement of a belt of brackish water deposits toward the bayline.

3. The tide-dominated estuary or barrier lagoon deposits within the transgressive half-cycle, which have been ignored or underrepresented in previous Wilcox sequence stratigraphy, are now well documented from cores. Abundant evidence exists of tidal processes, such as mud drapes, bimodal grain size layering, bidirectional ripples and cross strata, wavy and lenticular beds with tidal signals, tidal rhythmites, and tidal bundles. Although tidal processes are not restricted to the transgressive half-cycle; they are well preserved there because of being protected from the open-coast waves.

4. Long-distance shoreline migrations within the context of overall shelf margin progradation are documented from both wireline log and core data sets. The average shoreline transit distance is more than 28.5 km (17.7 mi), and maximum migration distance is more than 51.1 km (31.8 mi). The average shoreline regressive rate is more than 176.1 km/m.y. (109.4 mi/m.y.).

The above emphasis on shelf-extensive regressive–transgressive cycles also highlights the fundamental growth mechanism of both topset and shelf aggradation and of shelf edge accretion. Significant slope accretion occurred each time the shoreline reached the shelf edge area. Autogenic stratigraphic responses during the transgressive and regressive phases of cross-shelf migration ensure that sediment is also spread along a strike distance about equal to the cross-shelf migration distance (see Olariu, 2014). This results in the shelf edge accreting forward in a more even manner than that of individual shorelines on the shelf. Although researchers have suggested several typical process regime changes that can happen on deltas in time and space during shelf transits (e.g., Yoshida et al., 2007), for example, stronger storm wave influence as the delivery system approaches the shelf edge, this aspect of regime change has not been explored here.

An analogous, high-subsidence topset succession (relief >500 m [1640 ft]) has been documented from the Pliocene Orinoco shelf margin on Trinidad (Sydow et al., 2003; Bowman and Johnson, 2014; Chen et al., 2014). Two sedimentary successions, outcropping at both outer shelf and inner shelf sites, show similar repeated transgressive–regressive cycles. The Pliocene Mayaro Formation, which is more basinward (within 20 km [12 mi] of the shelf edge) shows wave-dominated deltaic or shoreface-like deposits (Bowman and Johnson, 2014). The Pliocene Morne L’Enfer Formation, which is more landward (~100 km [62 mi] landward of shelf edge) is composed of 17 transgressive–regressive cycles with strong tidal influence in both half-cycles (Chen et al., 2014). These two studies show that the Orinoco Delta transited up to 100 km (62 mi) back and forth, contributing sand to the deep-water areas beyond the shelf edge, on a time scale of approximately 100 k.y. (Sydow et al., 2003; Chen et al., 2014).

The estimated shoreline progradation rate for Wilcox deltas (176.1 km/m.y. [109.4 mi/m.y.]) is in the range of documented modern and ancient shoreline migration rates. Forward modeling of Burgess and Hovius (1998) based on 24 modern river delta systems (without application of waves or tides) shows cross-shelf transit duration is generally less than 120 k.y., irrespective of how wide the shelf is. Calculated from their results, the regressive shoreline migration rates are commonly less than 276 km/m.y. (171 mi/m.y.). Another review of modern and ancient delta clinoforms shows the progradation rates of sand-prone subaqueous delta clinoforms, muddy subaerial deltas and shoreline clinoforms, and sandy subaerial deltas and shoreline clinoforms to be 10–500, 1–20, and 1–30 km/m.y. (6–311, 1–12, and 1–19 mi/m.y.) when obtained data are long term (>10 k.y.) (Patruno et al., 2015).

Note that the transgressive–regressive cycles detailed above show clearly that the Wilcox deltaic system cannot be referred to as a long-term sea-level lowstand system below the shelf edge, nor were the Wilcox deltas likely to have been docked at the shelf edge. We agree that Wilcox greenhouse climate would have allowed only modest eustatic sea level amplitudes, but the very low coastal gradients nevertheless allowed extensive transgressions back onto the previous coastal plains, on a time scale of approximately 300 k.y.
Influences of High Sediment Supply and Modest Accommodation on Shoreline and Shelf Edge Building

As previous studies have stated, the Wilcox system had an unusually high sediment supply (Galloway et al., 2000; Carvajal et al., 2009; Sweet and Blum, 2011). Our study provides additional data that emphasize rapid progradation and slow aggradation on a shelf transit time scale of approximately 300 k.y. The estimated shelf edge progradation rate is generally less than 20 km/m.y. (12 mi/m.y.), averaging 7.8 km/m.y. (4.8 mi/m.y.), and aggradation rate is less than 250 m/m.y. (820 ft/m.y.), averaging 86.4 m/m.y. (283.5 ft/m.y.) (Figure 8, Table 2). Comparing the estimated progradation and aggradation rates of the Wilcox Group to the 33 studied shelf margins in Carvajal et al. (2009), the Wilcox system is
characterized by a high shelf edge progradation rate and a relatively low aggradation rate (Figure 8). This rapid progradation and slow aggradation strongly suggests that the Wilcox system was characterized by low shelf storage ability, frequent sand delivery across the shelf edge, and significant potential for sand bypass even during rising sea level (Porębski and Steel, 2003; Carvajal et al., 2009). Wilcox shorelines evidently prograded fast and arrived at the shelf edge frequently. The recently discovered voluminous turbidite successions on the Wilcox basin floor far out into the Gulf of Mexico are consistent with the above notions of sediment supply and accommodation.

Greenhouse global sea level changes are known to be of low amplitude (20–30 m [66–98 ft]) and low frequency (<1 m.y.) (Miller et al., 2005a) with exceptions of occasional eustatic falls of 60–70 m (197–230 ft) (see evidence in Zarra, 2007). As a result, greenhouse highstand condition provides only modest accommodation on the shelf and provides longer time for deltas to migrate toward the shelf edge; during greenhouse times, sea level has less potential to drop below the shelf edge at sea level lowstand. We therefore conclude that shoreline migration in greenhouse Wilcox time was probably not driven by global sea level change alone. The difference between global sea level change in greenhouse and icehouse times (low amplitude and low frequency vs. high amplitude and high frequency) also has implications to deep-water sand delivery (Blum and Hattier-Womack, 2009; Carvajal et al., 2009). Greenhouse highstand with less shelf storage and longer cycles of sea-level change has greater potential to deliver sandstone to deep water than during icehouse highstand. Carvajal and Steel (2006), Covault et al. (2007), and Boyd et al. (2008) all provide good examples to show formation of sandy deep-water fans during highstand conditions. In contrast, greenhouse lowstand condition, with less sea-level drop, is less able to deliver sands to deep-water areas compared with icehouse lowstand.

Basinward movements of salt or shale caused by depositional loading of a prograding deltaic system created a local extensional regime on the Wilcox shelf margin (Winker and Edwards, 1983). This extensional regime created growth fault zones with high subsidence rates. If the sea-level fall rate did not exceed the subsidence rate, relative sea level would have continued to rise and hinder the formation of sand bypass across the shelf. Therefore, if the shoreline arrived at a growth fault–controlled outer shelf area, it may have remained there and aggraded until the growth faults became less active. For example, the shorelines of the Oligocene Frio Formation of northwest Gulf of Mexico were relatively stable within a growth faulted subbasin, showing only minor landward or seaward migration (2–4 km [1–4 mi]) (Edwards, 1995; Olariu et al., 2013).

Consideration of Wilcox Shoreline Migration in a Greenhouse Setting

Global Paleocene–Eocene sea-level change was of low frequency and low amplitude (Miller et al., 2005a, b), and so we might predict that the repeated shelf transit migration distances of Wilcox shorelines should have been insignificant, as has been suggested by Sweet and Blum (2011). However, a repeated and relatively long-distance back-and-forth shoreline migration has now been documented (above) for these shorelines. In view of this result, we provide the following points to help explain the greenhouse Wilcox shoreline migrations.

Controls on Shoreline Migration

Cross-shelf shoreline migration is controlled by the interplay between accommodation, sediment supply and autogenesis (Swift and Thorne, 1991; Muto and Steel, 1997). All of these, in turn, are controlled by the tectonic (influencing drainage area, basin relief, and subsidence rate), climate (influencing basin average temperature and precipitation), and eustatic sea level drivers. Unsteadiness in these drivers through time produces the allogenic stratigraphic responses familiar to sequence stratigraphers, whereas the autogenic responses (shoreline retreat, delta lobe switching, and a variety of changing depositional processes) arise despite steadiness in these external drivers. All these controls work at a variety of different time scales (from $10^6$ to $10^8$ yr; Figure 9), although we wish to focus on the time scale of several hundred thousand years, which is the fundamental shelf transit time scale now documented in the Wilcox sediment delivery system.

Long-term sea-level change ($10^7$ to $10^8$ yr) is caused largely by seafloor spreading rates (Miller et al., 2005a), whereas short-term sea-level change...
is ascribed to variations of ice sheet volumes ($10^4$ to $10^6$ yr), modulated by Milankovitch cycles ($10^4$ to $10^5$ yr) (Miller et al., 2005a). Milankovitch theory shows how astronomical earth movements affect climate and the variations in solar radiation. Its frequencies affecting paleoclimate are eccentricity (400 and 100 k.y.), obliquity (41 k.y.), and precession (19 and 23 k.y.) (Hays et al., 1976; Berger et al., 1989; Blum and Hattier-Womack, 2009). The late Miocene through Holocene sea level was evidently influenced by $10^4$- to $10^5$-yr-scale Milankovitch changes (Miller et al., 2005a). In contrast, the ice volume variations at Milankovitch scale from Late Cretaceous through Eocene were modest, near or below detection limit (Miller et al., 2005a). The amplitude of sea-level change was therefore also low in these greenhouse times. From Late Cretaceous to Eocene, i.e., through Wilcox deposition, sea-level change is recorded to have ranged from approximately 15 to 30 m (49 to 98 ft), at a time scale of few hundred thousand to million years (Miller et al., 2005a, b). These greenhouse sea-level changes are relatively minor, compared with Pleistocene icehouse sea-level change (>100 m [328 ft]).

Sea-level change based on isotope analysis has been fairly well documented in the Neogene. However, variation of sediment supply has been only loosely documented and is still a poorly constrained factor in sequence development. In numerical or physical modeling studies, sediment supply is commonly considered as constant through time (e.g., Kim et al., 2006; Sømme et al., 2009a); only limited modeling studies considered the factor of sediment supply variations (e.g., Burgess and Prince, 2015). However, sediment supply does change in both long term and short term (Blum and Hattier-Womack, 2009). Based on a global database of 488 rivers, Syvitski and Milliman (2007) proposed a new model

Figure 9. (A) Illustration of a source-to-sink system showing the control elements for sediment supply (indicated by 1), accommodation (indicated by 2), and autogenesis (indicated by 3) (modified from Castelltort and Van Den Driessche, 2003). (B) Time scales for the control elements. (C) Idealized model for Wilcox shoreline migration (dashed orange line) and shelf edge building (red solid line).
termed BQART (equation 2) to show that geologic factors (basin size, relief, lithology, and ice erosion) can account for 65% of the between-river (i.e., modern rivers) sediment load variability, whereas climatic factors (precipitation and temperature) can explain an additional 14%:

$$Q_i = \omega BQ^{0.31}A^{0.5}RT \quad \text{for } T \geq 2^\circ C \quad (36^\circ F)$$  \hspace{1cm} \quad (2a)$$

$$Q_i = 2\omega BQ^{0.31}A^{0.5}R \quad \text{for } T < 2^\circ C \quad (36^\circ F)$$  \hspace{1cm} \quad (2b)$$

where $Q_i$ is sediment discharge (Mt/yr), $w = 0.0006$, $B = IL(1 - T_c)E_h$ ($I$ is glacier erosion factor, $L$ is average basin-wide lithology factor, $T_c$ is trapping efficiency of lakes and reservoirs, and $E_h$ is human-influenced soil erosion factor), $Q$ is mean water discharge (km$^3$/s), $A$ is drainage area (km$^2$), $R$ is maximum relief (km), and $T$ is basin average temperature ($^\circ C$).

Over a time scale of $10^6$ yr or less, the geologic factors are relatively steady, whereas the climate factors vary significantly. We can also assume that tectonic activity was an important factor affecting sediment discharge through affecting drainage area and relief generation over time scales of $10^6 - 10^7$ yr (Blum and Hattier-Womack, 2009; Carvajal and Steel, 2012). However, during the development of the Wilcox sequences documented here, climate events such as Milankovitch cycles and Paleocene to Eocene climate hyperthermals are very likely to have been a primary control on sediment discharge over a time scale of less than $10^6$ yr. These Milankovitch cycles would have influenced both precipitation and basin temperature. Paleocene to Eocene hyperthermals (Zachos et al., 1993; Thomas and Zachos, 2000) are rapid warmings during brief periods (~100 k.y.) during the Paleogene. A positive correlation exists between sediment discharge and temperature (Syvitski and Milliman, 2007), and even minor basin-average temperature changes would affect sediment discharge significantly. Application of the BQART model to the river systems of the Texas coastal plain and shelf by Blum and Hattier-Womack (2009) indicates that sediment yields were 25%–30% less than today during the last glacial maximum because of a 5$^\circ$C (41°F) depression of temperatures. We therefore conclude that climate variation is likely to have been an important contributor to shoreline migration within Wilcox shelf transit sequences that lasted a few hundred thousand years.

Subsidence and resultant compaction in passive margin settings is commonly considered constant at the million-year time scale. The amount of subsidence increases toward the hinge zone. The only evident change from initial phase of rapid subsidence to second phase of slow subsidence commonly takes tens of millions of years (see the detailed review in Xie and Heller, 2009). Therefore, the variation of subsidence rate is less likely to influence fundamental sequence development at the time scale of hundreds of thousands of years.

Shoreline autoretreat or delta lobe switching occurs at a time scale currently thought to be less than $10^5$ yr. In the case of the Mississippi Delta, delta lobes switched every 1–2 k.y. (Frazier, 1967; Coleman, 1988). The time scale of shoreline autoretreat is still under debate but is likely below the time scale of cross-shelf transgression or regression and can create parasequence-scale (<$10^5$ yr) depositional units (Kim et al., 2006). Depositional processes (fluvial, wave, storm, and tide) are also very important to construct and modify the shoreline. They are active at the scales of days (e.g., tide), seasons (e.g., seasonal storm events and river flooding), or years (e.g., minor cycles in storm events [Hampson and Storms, 2003]). In conclusion, the influences of shoreline autoretreat, delta lobe switching, and other self-organizational depositional processes are generally at a time scale of less than $10^5$ yr.

All these controls are variably operative at the same time (Figure 9) but within the time scale of cross-shelf back-and-forth shoreline migration (324 k.y. for the Wilcox case), which creates the fundamental sequence. High-frequency change of eustatic sea level and climate-driven sediment supply change are the two controls likely to have been key factors. Two reasons exist why climate driven supply is highlighted here, on top of eustatic change. One is that eustatic change was modest (but not likely to have been negligible) in forcing cross-shelf deltaic transits and likely to have needed enhancement by sediment supply. The second reason is that an absolute high supply, as amply documented on both shelf topsets and deep-water sand volumes, was a necessary condition in view of the frequency of deltas arriving at the shelf edge and because of the great spatial scale of the northern Gulf of Mexico source-to-sink system.

How Much Can Sediment Supply Change within a Few Hundred Thousand Years?

The second question here is not only how large the sediment supply was in the Wilcox system but also
how fast the sediment supply can change at the time scale of fundamental sequences (~100 k.y.). Therefore, the key question we care about is how fast this climate effect can change and its effect on sediment supply change, especially for hyperthermal events. Despite a poor understanding of the occurrence, triggering, and duration of such hyperthermal episodes, it is agreed that the duration is between 50 and 200 k.y. (Thomas, 1989, 1990; Kennett and Stott, 1991; Röhl et al., 2000). Röhl et al. (2000) claims the largest hyperthermal, known as Paleocene–Eocene Thermal Maximum (PETM) (Kennett and Stott, 1991; Sluijs et al., 2006; McInerney and Wing, 2011) interval, spans 11 precessional cycles within a duration of 210 to 220 k.y. Lately, the duration of the PETM has been narrowed into less than 170 k.y. (Röhl et al., 2007). In addition, PETM could have increased sea surface temperatures from 8°C to 10°C in high latitudes and 4°C to 5°C in low latitudes in a brief period of time (Zachos et al., 1993, 2003). If we assume a stable tectonic background (drainage area and relief), a 4°C–10°C short-term increase of average temperature within approximately 100 k.y., and a long-term 10°C–20°C average temperature in the Wilcox system from the late Paleocene to the early Eocene (Wilf, 2000), the sediment supply in Laramide basins has the potential to increase 20%–100% within 200 k.y. based on the BQART model (equation 2, Table 3). This dramatic change of sediment supply would be enough to cause the rapid and frequent cross-shelf shoreline migrations that we have documented. At least 17 hyperthermals, thought to be possibly related to orbital eccentricity, have been documented within almost half of Wilcox time from 57.5 to 52.5 Ma so far (Cramer et al., 2003; Zachos et al., 2010), and these could be a similar factor (as PETM) to increase sediment supply. These climate events are proposed to greatly enhance the modest eustatic sea level change in driving greenhouse Wilcox shoreline migrations.

### Table 3. Sediment Supply Increase Caused by Paleocene–Eocene Thermal Maximum at the Time Scale of 100 Thousand Years, Assuming No Significant Change of Bedrock Lithology, Drainage Area, and Basin Relief

<table>
<thead>
<tr>
<th>Increased Temperature by PETM Within Approximately 100 k.y. (°C [°F])</th>
<th>Long-Term Average Temperature (°C [°F])</th>
<th>Sediment Supply Increase by Calculation of BQART Model</th>
</tr>
</thead>
<tbody>
<tr>
<td>4–10 (39–50)</td>
<td>10–20 (50–68)</td>
<td>20%–100%</td>
</tr>
</tbody>
</table>

Abbreviations: PETM = Paleocene–Eocene Thermal Maximum.

### Nonuniqueness

Though we do highlight the factor of sediment supply in the Wilcox system, the eustatic sea-level influence should not be ignored. Instead, we think these long-distance back-and-forth shoreline migrations are controlled by frequent change (~100 k.y.) of unusually high sediment supply (progradation rate of shelf margin >10 km/m.y. [6 mi/m.y.]) and modified by low-amplitude and low-frequency sea-level change.

The shoreline migration of each fourth-order sequence during the 12 Ma interval is difficult to explain by a single cause, not only because each fourth-order sequence shows some differing characteristics (different shoreline migration rates and different shelf margin progradation and aggradation rates) but also because the same or similar stratigraphic geometry can be produced by different controls. Using forward modeling, Burgess and Prince (2015), in an analysis of sequence stratigraphic surfaces and systems tracts, have shown that each type of surface, stacking pattern, or shoreline trajectory could be caused by multiple controlling factors. The bottom line is that most surfaces and geometries can be explained in multiple ways; that is, the solutions are non-unique. Stratigraphic examples, such as shown by Vakarelov et al. (2006) and Plint and Kreitner (2007), also strongly suggest that similar high-frequency greenhouse shoreline migration could have been controlled by different mechanisms: tectonic mechanisms (Vakarelov et al., 2006) or low-amplitude sea-level changes (<10 m [33 ft]) (Plint and Kreitner, 2007).

Shoreline migration of each fourth-order sequence, showing different migration distance and different stratigraphic geometry, may have responded to different magnitudes of sea-level change and sediment supply change. The varying magnitude of sediment supply may have been caused by tectonic-controlled third-order sediment supply change superimposed on
fourth-order, climate-driven sediment supply change or by different magnitudes of climate change between each fourth-order sequence. The different magnitudes of relative sea-level change may have been caused by local subsidence rate variations (active and inactive period of growth fault) or different amplitudes of global sea-level change.

Implications for Wilcox Exploration and Production Onshore and Offshore

The regressive–transgressive cycles documented in this paper are the fundamental building blocks of the Wilcox shelf margin prism. Although several canyons (or large incised valleys) are well documented in the Wilcox shelf margin (Galloway et al., 1991; Zarra, 2007; Cornish, 2013) and thought to have been important sediment pathways to deep water, these fundamental shelf transit sequences also were a critical factor to supply the canyons by delivering sediments from inner shelf to shelf margin and even possibly delivered sediments over the shelf edges without the existence of canyons (see the example in Dixon et al., 2012). These sediments could have been captured by the growth-faulted subbasins or minibasins within the slope or finally transported beyond to the basin floor. The documented shelf edge progradation/aggradation ratio above is also a good indicator for the timing of sediments delivered to deep-water basins (Carvajal and Steel, 2006; Carvajal et al., 2009). Rapid progradation and slow aggradation (e.g., initial period of lower Wilcox) is the favorable condition for the sediment bypass on the shelf.

These transgressive–regressive cycles also have implications for Wilcox onshore exploration and production. First, attention should be drawn to the existence of marine transgressive deposits within amalgamated fluvial deposits and the resultant reservoir heterogeneity and anisotropy caused by their differing reservoir quality (porosity and permeability). Second, the presence of tidal channels and bars within transgressive estuaries adds complexity to reservoir characterization and modeling (Breyer and McCabe, 1986). The changes of dominant processes are able to influence the orientation, lateral connectivity, external shape, and internal geometry of reservoir. Last but not least, the Wilcox expanded faulted zones also received a high amount of sediments and now represent good potential traps.

CONCLUSIONS

1. A revisit of some 350 wells and 450 m (1476 ft) of cores through the Wilcox succession in south Texas shows that the fundamental organization of the linked facies tracts was one of 37 transgressive to regressive units, each thickening from some 30 m (98 ft) on inner shelf to some 70 m (230 ft) on outer shelf reaches. Cores show that the transgressive half-cycle formed mainly from tidally influenced estuaries and wave-influenced barrier lagoon systems, whereas the regressive half-cycle was generated mainly from river- and wave-dominated deltas.

2. Relatively long–distance back-and-forth shoreline migration is documented. Average regressive or transgressive shoreline migration was 28.5 km (17.7 mi) for a single high-frequency sequence, with maximum exceeding 51.5 km (32.0 mi).

3. The documentation of repeated transgressive–regressive shoreline behavior indicates that the Wilcox system was not associated with a long-term sea-level lowstand below the shelf edge (Rosenfield and Pindall, 2003) and that although the delta transits were of relatively low frequency, the deltas were not docked at the shelf edge (Sweet and Blum, 2011). The general sediment supply to the gulf from the Texas coastal plain and shelf was high, causing the shelf edge to prograde greater than 10 km/m.y. (6 mi/m.y.) during lower Wilcox deposition. However, we suggest that the back-and-forth shoreline transits to the shelf edge were controlled by frequent changes in water and sediment discharge as a result of late Paleocene–early Eocene climate hyperthermals (~100 k.y.) but also modulated by low-amplitude and low-frequency eustatic sea-level changes.

4. The Wilcox shelf edge progradation rate ranged from 0 to 86.1 km/m.y. (53.5 mi/m.y.), and aggradation rate was from 37.0 to 228.0 m/m.y. (121.4 to 748.0 ft/m.y.). This rapid progradation and slow aggradation, especially during lower Wilcox deposition, suggests a low shelf storage ability, frequent sand delivery across the shelf edge, and a potential for sand bypass during rising sea level (Carvajal et al., 2009).
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