EVOLVING COASTAL PROCESS REGIMES

Title:
SPATIAL AND TEMPORAL EVOLUTION OF COASTAL DEPOSITIONAL SYSTEMS AND REGIONAL DEPOSITIONAL PROCESS REGIMES: CAMPANIAN WESTERN INTERIOR SEAWAY, U.S.A.

Authors:
MARIJN VAN CAPPELLE1,2, GARY J. HAMPSON1*, HOWARD D. JOHNSON1

1Department of Earth Science and Engineering, Imperial College London, South Kensington Campus, London SW7 2AZ, UK
2present address: PDS Petrotechnical, Rijswijk, the Netherlands

*Corresponding author e-mail:
g.j.hampson@imperial.ac.uk

Key words:
depositional process classification, tides, embayment, Cretaceous, Mesaverde Group, tidal amplification

Word count (abstract): 422
Word count (text): 12,156
Word count (references): 5,875
Word count (figure captions): 1,434
Number of figures: 11
Number of tables: 0
ABSTRACT

This paper provides a critical review and regional synthesis of Late Cretaceous shallow-marine deposits along part of the western margin of the Western Interior Seaway of North America, which contains the most extensively documented outcrop-based studies of siliciclastic coastal depositional systems in the world. The results of this synthesis are presented in the form of paleogeographic maps (covering present-day New Mexico, Utah, Colorado, and Wyoming, USA) for five “timeslices” in the Campanian. These maps are used to evaluate the spatial and temporal evolution of regional depositional process regimes along a large (>1000 m) stretch of coastline. The evolution of regional depositional process regimes is linked to tectonic and paleoceanographic controls on the Western Interior Seaway, which enables the results of this synthesis to be applied to prediction of depositional process regimes in other, less intensively studied basins.

Six gross depositional environments have been mapped for each “timeslice”: (1) alluvial-to-coastal-plain sandstones; (2) coastal-plain coals, mudstones and sandstones; (3) shoreline sandstones; (4) marine mudstones; (5) gravity-flow siltstones and sandstones; and (6) marine marls and chalk. Shoreline sandstones in each “timeslice” are interpreted in further detail using documented evidence for the three principal classes of depositional process (wave, tidal, and fluvial) and published reconstructions of coastal morphology, which is widely considered to reflect depositional process regime. Based on these interpretations, shoreline sandstones are assigned to five categories of depositional process regime: (1) regressive wave-dominated shorefaces and delta fronts; (2) regressive fluvial-dominated delta fronts; (3) regressive mixed tide- and wave-influenced delta fronts; (4) regressive tide-dominated delta fronts; and (5) transgressive barrier islands, back-barrier lagoons, and estuaries. The accuracy of and uncertainty in classification of depositional process regime are critically evaluated. Additionally, stratal thickness and sediment routing pathways have been interpreted in order to assess the impact of tectonic and paleoceanographic controls on spatial and temporal changes.

In all of the evaluated shallow-marine successions, thin tide-influenced intervals were deposited in back-barrier and lagoon systems associated with net-transgressive shorefaces and estuaries. However, it is notable that all preserved tide-dominated and tide-influenced regressive deltaic systems are located along the northern margin of the “Utah Bight”, a tectonically-related embayment that had a pronounced expression during the middle Campanian. The southern margin of the embayment, and coastlines to the north, are conspicuously more wave-dominated, which supports their exposure to a
larger wave fetch. In contrast, the northern margin of the embayment was relatively wave-protected. It is concluded that tidal range was amplified due to resonance of the principal semi-diurnal tide in the strongly embayed geometry of the middle Campanian “Utah Bight”.

[end of abstract]

INTRODUCTION

Shallow-marine depositional systems are commonly interpreted in terms of the relative interaction of wave, tidal, and fluvial processes. A classification using these three processes as end-members was first proposed by Galloway (1975) for regressive river deltas. Later, the classification was expanded to include non-deltaic and transgressive coastlines, and to categorize a range of modern depositional environments (Boyd et al. 1992). This classification scheme was extended further by Ainsworth et al. (2011) to quantify the relative importance of wave, tidal, and fluvial depositional processes for the full spectrum of siliciclastic coastal deposits. In this paper, we apply the latter classification system on a regional scale to the shallow-marine deposits of the western margin of the Campanian (Upper Cretaceous) Western Interior Seaway of North America.

The area of interest is regionally extensive, covering 600,000 km² and comprising parts of New Mexico, Colorado, Utah, and Wyoming (USA) (Fig. 1A), where shallow-marine deposits are exceptionally well exposed in a series of large outcrop belts (e.g., McGookey et al. 1972; Kauffman and Caldwell 1993; Krystinik and DeJarnett 1995) (Fig. 1C). The paleogeographic evolution of the Western Interior Seaway is well documented, including the presence of the “Utah Bight” embayment along the western margin of the seaway throughout much of its development (e.g., McGookey et al. 1972; Kauffman and Caldwell 1993; Dickinson and Gehrels 2008; Szwarc et al. 2015). In addition, the large areal extent and high quality of the outcrops have enabled detailed facies and sequence stratigraphic analysis of the strata in a combination of cliff-face and canyon exposures, which have been the subject of numerous investigations over many years (e.g., Van Wagoner et al. 1990; O’Byrne and Flint 1995; Taylor and Lovell 1995; Hampson and Howell 2005; Hampson 2010; Hampson et al. 2011; Legler et al. 2014; Van Cappelle et al. 2016). Mapping can be extended into the subsurface using well logs (e.g., Berg 1975; Boyles and Scott 1982; Palmer and Scott 1984; Tillman and Martinsen 1984; Roehler 1990; Wood 2004; Hampson et al. 2008; Hampson 2010; Burton et al. 2016) and, occasionally, seismic data (e.g., Horton et al. 2004). Cores have been taken from some hydrocarbon-bearing Campanian reservoirs (Berg 1975; Tillman and Martinsen 1987), and from research boreholes (Van Wagoner et al. 1990).
Thus, the large scale paleogeographic evolution of the western margin of the seaway is well known, and detailed sedimentologic descriptions and depositional process interpretations are available for many of the outcropping shallow-marine strata. However, the temporal and spatial evolution of the coastline has not yet been combined at a regional scale with the type and distribution of depositional processes that operated in coastal and shallow-marine environments. The main aims of this paper are, on the basis of extensive published literature: (1) to critically review the facies character of these shallow-marine deposits, and (2) to synthesize the spatial and temporal variations in coastal morphology and depositional processes in the studied strata for the first time. Of special interest is the distribution and potential origin of regressive coastlines with a regionally pronounced tidal signature, and the reason why the resulting deposits contain abundant evidence for tidal reworking (Suter and Clifton 1999; Aschoff and Steel 2011b; Steel et al. 2012). The synthesized temporal and spatial evolution of depositional process regimes is interpreted in the context of tectonic and paleoceanographic controls on the Western Interior Seaway. We then discuss the application of a similar approach, informed by an understanding of tectonic and paleoceanographic context, to predict depositional process regimes in other, data-poor basins.

**GEOLOGIC SETTING**

From the Late Jurassic to the Paleogene, the Farallon plate was subducted below the western margin of the North American plate. At the convergent plate boundary, the volcanic arc and thin-skinned fold-and-thrust belt of the North American Cordillera formed, with an associated retroarc flexural foredeep (DeCelles 2004) (Fig. 1A, B). Due to subduction of the cold Farallon slab, the asthenosphere below North America cooled and dynamic subsidence took place across much of the width of the North American plate. This caused the epicontinental Western Interior Seaway to flood the continent (Liu and Nummedal 2004; Liu et al. 2011, 2014). Thus, reconstructed subsidence profiles across the Cretaceous Western Interior Basin indicate a broad area of relatively uniform dynamic subsidence, with a superimposed narrow area of more pronounced subsidence at its western margin, due to short-wavelength flexural loading of the foredeep (e.g., Liu et al. 2014) (Fig. 1B). During most of the Cretaceous, deformation was concentrated in the north-south striking thin-skinned fold-and-thrust belt (Sevier Orogeny) (DeCelles 2004). Thrust faults are organised in a series of convex-outward ‘salients’ in which the thrusts became progressively younger to the east, in normal sequence (Fig. 1C) (Burtner and Nigrini 1994; DeCelles and Mitra 1995; DeCelles et al. 1995; DeCelles 2004; Yonkee and Weil 2010, 2015). During the latest Cretaceous and Paleogene, deformation became gradually thick-
skinned in style (Laramide Orogeny), with several basement-cored uplifts emerging within the Cordilleran foreland due to a decrease in the dip of the subducting Farallon slab (DeCelles 2004; Yankee and Weil 2015; Copeland et al. 2017). Stratal thicknesses, mineralogic-compositional data, and paleo-drainage reconstructions provide evidence for activity of some Laramide-style structures during the Campanian and Maastrichtian (Dorr et al. 1977; Lawton 1983, 1986; Bryant and Nichols 1988; Shuster and Steidtmann 1988; Steidtmann and Middleton 1991; Miall and Arush 2001; Leva López and Steel 2015). Laramide structures became active in the direction of shortening (SW-NE) in normal sequence between 90 Ma in California to 60 Ma in South Dakota (Copeland et al. 2017). From the Sevier fold-and-thrust belt, sediment was transported eastward into the Western Interior Basin, where it was deposited to form a series of wedges of alluvial, coastal-plain and shallow-marine siliciclastic deposits with an overall north-south-trending coastline.

Along the southwestern margin of the Western Interior Seaway, the northwest-southeast striking Bisbee rift basin was present in Arizona, New Mexico and Mexico (Dickinson and Lawton 2001). The shoulders of the Bisbee rift basin formed the Mogollon Highlands (Bilodeau 1986) (Fig. 1A), which together with the Cordilleran magmatic arc supplied sediment northeastwards (axially) into the Cordilleran foreland, approximately orthogonal to the eastward-directed (transverse) sediment supply from the Sevier fold-and-thrust belt (Lawton et al. 2003; Dickinson and Gehrels 2008; Lawton and Bradford 2011; Lawton et al. 2014; Szwarc et al. 2015). The “Utah Bight” lay north of, and was supplied with sediment from, the Mogollon Highlands (Fig. 1A) at various times throughout the Cretaceous (e.g., Lawton et al. 2003; Dickinson and Gehrels 2008; Lawton and Bradford 2011; Lawton et al. 2014; Szwarc et al. 2015).

Several previous studies have reconstructed regional paleoshoreline positions along the western margin of the Western Interior Seaway within the context of a well-established ammonite biostratigraphic scheme (e.g., Cobban et al. 2006). However, these studies have each concentrated on a restricted area (Zapp and Cobban 1960; Roehler 1990; Franczyk et al. 1992), or have been compiled at low temporal resolution (McGookey et al. 1972; Kauffman and Caldwell 1993). In addition, none of these studies reconstructed the depositional process regime along the regional paleoshorelines, which is one aim of this work.

During the Campanian, the study area was located at 40-55° N latitude (Fricke et al. 2010). The climate was warm temperate with year-round or seasonal precipitation, and the land was covered by deciduous broad-leaved forests with a closed canopy. Just south of the study area was a transition to
a more tropical climate harboring savannah vegetation, and just north of the study area there was a
transition to cool temperate climates harboring more open forests and shrubland (Sellwood and
Valdes 2006; Fricke et al. 2010). Oceanic circulation within the Western Interior Seaway was
dominated by anti-clockwise, thermo-haline and storm-generated currents (Erickson and Slingerland

DATASET AND METHODS

This paper does not present original primary data, but instead compiles and synthesizes secondary
data from published studies, most of which are based on outcrop data. The compiled dataset has
three aspects. First, published descriptions and interpretations of shoreline-sandstone facies
successions in the strata of interest are critically reviewed and their depositional process regime is
classified consistently using the scheme of Ainsworth et al. (2011). Second, the location and
depositional process regime of shorelines is shown in maps of five “timeslices”; each “timeslice” can
be related to periods of transgression and regression that are identified in most Tertiary Laramide
basins in the area of interest (Fig. 2). Third, spatial and temporal variations in depositional process
regime at the shorelines are discussed using the five “timeslice” paleogeographic reconstructions,
including their relationship to structures that were active during deposition.

Classification of depositional processes in shoreline sandstones

In the widely used scheme of Boyd et al. (1992), shoreline types are qualitatively positioned within a
ternary diagram that has three classes of depositional process (wave, tidal, fluvial) at its apices.
Ainsworth et al. (2011) recently proposed a more detailed classification scheme, in which the relative
influence of the three process classes (wave, tidal, fluvial) are qualitatively or quantitatively estimated
in a particular shoreline or shoreline stratigraphic-architectural element. The dominant process is
assigned a capital letter (W - wave, T - tidal, F - fluvial), the secondary process is assigned a non-
capitalized, non-italicized letter (w - wave, t - tidal, f - fluvial), and the tertiary process is assigned a
non-capitalized, italicized letter (w - wave, t - tidal, f - fluvial). We use herein the scheme of Ainsworth
et al. (2011), because it provides a concise and consistent way of summarising depositional process
interpretations in the full continuum of shoreline deposits. However, this and similar classification
schemes still have limitations, particularly when applied quantitatively; these limitations are related
to the spatial scale of analysis, and to the degree to which available data representatively sample the
unit of interest (Ainsworth et al. 2011; Rossi et al. 2017). Our analysis is focussed at relatively large
spatial scales (e.g., parasequences and parasequence sets), and we also interpret a range of
depositional process regimes to encompass uncertainty in detailed shoreline classification, in order to
mitigate the effects of potentially unrepresentative sampling.

Lithostratigraphy and biostratigraphy

The five “timeslices” have been defined based on ammonite biostratigraphy. A detailed ammonite
biostratigraphy is available for the Campanian marine mudstones of the Western Interior Seaway,
with an average age resolution of 0.5 Myr (24 zones in 12 Myr; Krystinik and DeJarnett 1995). Ten of
these ammonite zones have been dated radiometrically using volcanic ash beds (Cobban et al. 2006)
(Fig. 2). Tracing “timeslices” becomes progressively more uncertain with increasing distance from the
ammonite-bearing marine mudstones, especially in coastal-plain and alluvial deposits. Recent and
ongoing age dating of detrital zircons suggests that coastal-plain and alluvial deposits, in particular,
may differ in age by up to several million years from the ages inferred using long-distance sequence
stratigraphic correlations tied to ammonite zones and radiometrically dated bentonites (e.g., Lawton
and Bradford 2011; Szwarc et al. 2015; Pettit and Blum 2017; Bartschi et al., in press). Uncertainty in
the correlation of shoreline sandstones, which are the focus of this paper, is lower than for coastal-
plain and alluvial deposits; even if the absolute age of shoreline sandstones is uncertain, their
correlation is constrained by physical stratigraphic relationships mapped at outcrop by previous
workers and by zonal ammonites documented in intercalated marine mudstones.

The gross depositional environments (GDEs) for each “timeslice” have been reconstructed in map
view based on published regional cross sections that are largely lithostratigraphic in character. These
regional cross sections tie the ammonite zones to Campanian lithostratigraphic units preserved
within the following Tertiary Laramide basins (Fig. 2): (1) San Juan Basin (Molenaar et al. 2002); (2)
Jinnah et al. 2009; Seymour and Fielding 2013); (3) Uinta and Piceance basins (Young 1955; Cobban
1969; Gill and Hail 1975; Fouch et al. 1983); (4) Denver, Middle Park and Sand Wash basins (Izett et al.
1971); (5) Rock Springs Uplift (Roehler 1978); (6) Washakie and Wind River basins (Merewether et al.
1997); (7) Hanna and Laramie basins (Gill et al. 1970); and (8) Bighorn and Powder River basins (Gill
and Cobban 1966). In addition to the age uncertainty outlined above, there is also uncertainty in
correlation between Laramide basins at temporal scales below the age resolution of the ammonite
biostratigraphic framework. Consequently, each “timeslice” contains some degree of time-averaging
over a period of up to several hundred thousand years. The paleogeographic reconstructions
presented here for each “timeslice” are constrained by the most reliable age data that are currently available, but they will undoubtedly change as new chronostratigraphic data are collected and incorporated.

**Active structures**

Thrust faults and folds that were active during the Campanian have been mapped (Fig. 1C) (DeCelles 2004). In the Sevier fold-and-thrust belt, the following three main groups of faults were active: (1) the Absaroka Thrust in the Wyoming Salient (Burtner and Nigrini 1994; DeCelles 1994; Yonkee and Weil 2015); (2) the Charleston-Nebo and Frontal Triangle Zone thrusts in the Charleston-Nebo Salient (Horton et al. 2004; DeCelles and Coogan 2006); and (3) the Paxton Thrust, Gunnison Thrust (Lawton et al. 1993; DeCelles et al. 1995; DeCelles and Coogan 2006) and Iron Springs Thrust (Goldstrand 1994) in southwest Utah. In addition, some incipient Laramide structures were also active during the late Campanian, most notably the Wind River Uplift (Dorr et al. 1977; Shuster and Steidtmann 1988; Steidtmann and Middleton 1991), thrusts bounding the Green River, Great Divide and Washakie basins (Leva López and Steel 2015), the Uinta Uplift (Bryant and Nichols 1988; Plink-Bjorklund 2008; Steel et al. 2012) and the San Rafael Swell (Lawton 1983, 1986; Miall and Arush 2001). There is no evidence for active Laramide structures around the San Juan Basin (e.g., Rio Grande Uplift, Sangre de Cristo Uplift and Sawatch Range) during the Campanian (Bryant and Naeser 1980; Seager et al. 1997).

**GROSS DEPOSITIONAL ENVIRONMENTS AND SHORELINE-SANDSTONE CLASSES**

Based on published literature, six GDEs have been identified within the studied strata (from landward to seaward): (1) alluvial-to-coastal-plain sandstones; (2) coastal-plain coals, mudstones and sandstones; (3) shoreline sandstones; (4) marine mudstones; (5) gravity-flow siltstones and sandstones; and (6) marine marls and chalk (e.g., Young 1955; Gill and Cobban 1966; Gill et al. 1970; Izett et al. 1971; Gill and Hail 1975; Roehler 1978; Fouch et al. 1983; Merewether et al. 1997; Molenaar et al. 2002). Descriptions and interpretations of these GDEs are summarized in Figure 3. More detailed descriptions and process interpretations of five classes of shoreline sandstones are given below. Each shoreline-sandstone class is considered in terms of facies types, facies successions, depositional processes, and depositional environment (Figure 4). Bioturbation intensity is described using the Bioturbation Index (BI) scheme of Taylor and Goldring (1993). The descriptions and process interpretations presented below are based on critical review and synthesis of previously published studies of Campanian shoreline sandstones in the area of interest. The descriptions and
interpretations therefore represent facies models based on a large number of previously published case studies, rather than primary data from a particular example.

The classification of depositional process regime in shoreline sandstones (GDE #3 in the list above and in Fig. 3) is typically based on a combination of (1) facies analysis of individual beds or groups of beds (Ainsworth et al. 2011), (2) the nature of vertical genetically-related facies successions (e.g., Van Wagoner et al. 1990), and (3) lateral geometrical relationships offered by continuous outcrops and/or high-confidence in the correlation of laterally adjacent vertical sections. In this study, applying the semi-quantitative percentage-based classification of depositional processes, as proposed by Ainsworth et al. (2011) and Rossi et al. (2017), is problematic because a single sedimentary structure can be formed by several (potentially co-occurring) depositional processes. For example, cross-bedding and current-ripple cross-lamination can be formed by fluvial currents, tidal currents and wave-generated longshore currents. Hence, at the scale of this synthesis, the vertical and lateral facies relationships, or their inferred group of depositional environments, is often more indicative of the dominant process.

Wave-dominated shorefaces and deltas fronts

Description.--- These deposits consist of 10-50 m thick, coarsening-upward successions of mudstones, siltstones and very fine- to medium-grained sandstones, which are locally capped by coal seams (Fig. 4A). In the lower part of a typical succession, marine mudstones (facies OS) grade upward into interbedded mudstones, siltstones and thin (5-50 cm thick), hummocky cross-stratified (HCS), very fine- to fine-grained sandstones (facies dLSF). Higher up in the succession, HCS beds become thicker (20-100 cm), more abundant and amalgamated, and contain swaly cross-stratification (SCS) (facies pLSF). The upper part of the succession consists of fine- to medium-grained cross-bedded sandstones with paleocurrent directions approximately parallel to the paleoshoreline position, and generally southward directed (facies USF). These sandstones are overlain by medium-grained sandstones containing low angle (~1°), inclined planar bedding (facies FS). The HCS, SCS and cross-bedded sandstones show an overall upward-decreasing diversity and intensity of bioturbation (from BI=5 to BI=1) (Kiteley and Field 1984; Van Wagoner et al. 1990; Kamola and Van Wagoner 1995; Hampson and Howell 2005; Hampson et al. 2011).

Shoreline sandstones of this class predominate in the following lithostratigraphic units (Fig. 2): the Point Lookout Sandstone, Cliff House Sandstone, and Pictured Cliffs Sandstone of the San Juan Basin.
Interpretation.--- HCS and SCS (facies dLSF and pLSF) are the result of storm wave action, predominantly between storm wave base and fair-weather wave base (Dott and Bourgeois 1982; Duke 1985; Keen et al. 2012). The upward increase in abundance of HCS and SCS beds is attributed to upward-shallowing, upward-increasing wave energy and storm intensity, and/or upward-increasing sand supply (Storms and Hampson 2005). High-angle cross-bedding (facies USF) is the result of unidirectional longshore currents above fair-weather wave base, forced by waves approaching the coastline at an oblique angle and possibly augmented by southward directed, regional thermo-haline currents (Erickson and Slingerland 1990; Slingerland et al. 1996). Low-angle inclined-bedded sandstones (facies FS) are interpreted to have been deposited in the swash zone, where waves break on the foreshore. All the structures in these facies are the result of oscillatory storm and fair-weather wave action, and facies higher up in the succession are interpreted to have been deposited in shallower paleo-water depths. Therefore this facies succession represents a prograding wave-dominated shoreface (W sensu Ainsworth et al. 2011) (Fig. 4B, C).

Variation in facies motif.--- Three groups of observations indicate a divergence from the end-member interpretation of wave-dominated shoreface deposition presented above. Firstly, variability in bioturbation intensity and diversity can be attributed to variability in physicochemical stresses, such as sedimentation rate and sea-water salinity (cf. MacEachern and Bann 2008). It is interpreted that in areas close to river mouths the water salinity fluctuated between fully marine and fresh, and
deposition rates were episodically very high, with the latter coinciding with periods of peak river discharge. In areas away from river mouths, salinities were uniformly fully marine and deposition rates were uniformly low, resulting in a more diverse suite of trace fossils and higher bioturbation intensities. Secondly, the presence of current ripple cross lamination and climbing ripples in facies OS and dLSF indicate that unidirectional currents also acted in this environment, and based on paleogeographic reconstructions it is interpreted that the coastline had a locally cuspat geometry (Devine 1991; Hampson and Howell 2005; Hampson et al. 2011). Thirdly, time-equivalent strata in the lower coastal plain contain abundant fluvial channels (Hampson et al. 2012; Flood and Hampson 2014), which in some cases cut down from the top of shoreface successions, implying a genetic relation between these fluvial channels and the subjacent shoreface (O'Byrne and Flint 1995; Taylor and Lovell 1995; Pattison 2005; Pattison et al. 2007). In combination, or in isolation, these three groups of observations enable differentiation between fluvial processes in wave-dominated deltas (Wf sensu Ainsworth et al. 2011) and non-deltaic, flanking shoreface environments (W sensu Ainsworth et al. 2011) (Fig. 4B, C).

Fluvial-dominated delta fronts

Description.---These consist of upward coarsening successions 10-40 m thick that consist of bioturbated mudstones grading upwards into medium- to coarse-grained sandstones (Fig. 4D) (Olariu et al. 2010; Hampson et al. 2011). Bioturbated mudstones (facies OS) gradually coarsen upward into thin (1-10 cm thick), very fine- to fine-grained, structureless, current-ripple cross-laminated, and planar-laminated sandstone beds separated by mudstone interbeds (facies dDF). Sandstone beds may also contain rare HCS and wave-ripple cross-lamination. Higher up in the succession, mudstone beds are thinner (<20 cm) and less abundant, and sandstone beds become thicker (10-100 cm) and coarser grained (fine- to medium-grained sandstones) (facies pDF). Some beds have sharp bases, fine-upward and show an upward change from planar lamination to current-ripple cross-lamination. Other beds have gradational tops and bases and show a succession of planar lamination, current-ripple cross-lamination and planar lamination. Bioturbation intensity decreases (from BI=4 to BI=0-1) and trace fossil diversity decreases to the presence of just Ophiomorpha, Skolithos, Paleophycus and Planolites (pDF). The succession coarsens upward into laterally continuous medium-grained, planar-laminated, structureless and cross-bedded sandstones with little bioturbation (BI=0-3) (facies MB). All these facies are deposited on relatively steeply inclined (generally ~5° to a maximum of 15°) clinoforms (Enge et al. 2010; Olariu et al. 2010). Locally, sharp-based, fine- to coarse-grained, structureless or cross-bedded sandstones are present in channelized and lenticular bodies at the top of
these succession (facies DC) (Hwang and Heller 2002; Olariu et al. 2010).

This class of shoreline sandstones forms a subordinate component of the following lithostratigraphic units (Fig. 2): the Muley Canyon Sandstone of the Henry Basin (Birgenheier et al. 2009; Seymour and Fielding 2013); the Star Point Sandstone, particularly the Panther Tongue, and shallow-marine members of the Blackhawk Formation of the Uinta Basin (e.g., Kamola and Van Wagoner 1995; Hampson et al. 2011; Forzoni et al. 2015); and the upper Sego Sandstone of the southern Piceance Basin (Kirschbaum and Hettinger 2004).

**Interpretation.**—The abundance of planar-laminated beds, current-ripple cross-lamination and cross-bedding indicates deposition from unidirectional currents. The prevalence of mudstone interbeds indicates that the sand was transported intermittently. In facies dDF and pDF, fining-upward sandstone beds with a sharp base are interpreted to have been deposited by decelerating flows, possibly from turbidity currents. Beds with gradational bed boundaries record sustained, waxing and waning currents, probably river-derived hyperpycnal flows (Olariu et al. 2010; Hampson et al. 2011). The occurrence of clinoforms indicates that deposition took place on relatively steeply inclined delta fronts (Enge et al. 2010; Olariu et al. 2010). The overall upward decrease in mudstone content, upward increase in sandstone grain size, and upward increase in cross-strata size indicates an upward increase in energy on the delta front and contiguous mouth bar, where fluvial currents decelerated and deposition of sand on dunes took place at distributary channel mouths. Locally, distributary channels cut through the top of coarsening-upward delta front successions (Olariu et al. 2010) or their remnants were reworked to form lags during subsequent transgression (Hwang and Heller 2002). The facies successions indicate that deposition took place in a prograding, fluvial-dominated delta front setting (F sensu Ainsworth et al. 2011) (Fig. 4E, F).

**Variation in facies motif.**—Evidence for wave reworking is also present in some delta front successions, in the form of wave ripples and HCS beds. Therefore, fluvial-dominated delta fronts were also locally wave-influenced (Fw sensu Ainsworth et al. 2011) (Fig. 4E, F). Detailed mapping of strata in the outcrop belts shows that wave-dominated shoreface and fluvial-dominated delta front deposits were often coeval along the same paleoshoreline (Hampson and Howell 2005; Hampson et al. 2011).

**Mixed tide- and wave-influenced delta fronts**
Description.— Tide-dominated or tide-influenced delta-front deposits consist of upward-coarsening successions that are 10-40 m thick. The lower and distal parts of these successions consist of bioturbated mudstones (facies OS) which are interbedded with hummocky cross-stratified, very fine-to fine-grained sandstone beds (facies dLSF) that become amalgamated upwards as mudstone interbeds thin and become less abundant (facies pLSF) (Fig. 4G). These deposits are overlain across a low-relief erosional surface by cross-bedded fine-grained heterolithic sandstones, with evidence of bidirectional paleocurrent directions (facies LRTC). However, most paleocurrents trends are asymmetric, usually comprising a strong paleo-seaward-directed mode and a subordinate, paleo-landward-directed mode (e.g., Van Cappelle et al. 2016). Mud drapes are present along the foresets of cross-beds, and the toesets of cross-beds are heterolithic and current-ripple cross-laminated. The cross-bedded sandstones are interbedded with coarsening-upward, heterolithic, very fine- to fine-grained sandstones that contain wave-ripple cross-lamination and bidirectional current-ripple cross-lamination (facies TB). Towards the top and most proximal part of the succession, fine- to medium-grained sandstones with high-relief (up to 20-30 m) channelized erosional bases of concave-upward geometry are present (facies HRTC). These sandstones are characterized by cross-beds that contain mud drapes on their foresets and current-ripple cross-lamination in their heterolithic toesets, together with bidirectional paleocurrent populations. The latter three facies (LRTC, TB, HRTC) have a relatively low intensity (BI=0-3) and diversity of trace fossil content. Sandstones at the top of the succession (facies LRTC and HRTC) are commonly heavily bioturbated (BI=5) with prominent *Ophiomorpha* and/or *Skolithos* (Van Wagoner 1991; Willis and Gabel 2001, 2003; Legler et al. 2014; Van Cappelle et al. 2016; Burton et al. 2016).

Shoreline sandstones of this class predominate in the lower Sego Sandstone of the Uinta and Piceance basins (e.g., Willis and Gabel 2001, 2003; Painter et al. 2013; Legler et al. 2014; Burton et al. 2016; Van Cappelle et al. 2016).

Interpretation.— There is a clear distinction between deposits in the lower and upper parts of the succession. The lower part (facies OS, dLSF and pLSF) is dominated by wave-ripple cross-lamination and storm-generated HCS beds, and follows the upward-coarsening and upward-shallowing trends of a wave-dominated shoreface succession (compare Fig. 4A and 4G). In contrast, the upper part of the succession (facies LRTC, TB, HRTC) is dominated by cross-strata that contain evidence of fluvio-tidal currents in the form of bidirectional paleocurrents with a strong seaward-directed asymmetry (inferred fluvial and/or ebb-tide direction), pervasive mud drapes, and restricted trace fossil assemblages consistent with fluctuating salinity (Fig. 4G). Previously it has been interpreted that these
tide-dominated facies were deposited in incised valleys (Van Wagoner 1991), and that there was a
temporal change in process regime from an earlier (pre-valley incision) wave-dominated shoreface
(W sensu Ainsworth et al. 2011) to later (post-valley incision) tide-dominated estuaries within the
valley fills (Tf sensu Ainsworth et al. 2011). Later interpretations place many of the tide-dominated
deposits in the proximal part of regressive tide-dominated deltas (Ft sensu Ainsworth et al. 2011)
(Willis and Gabel, 2001). Wave- and tide-dominated facies are interbedded in distal locations,
implying that they were deposited in different coeval parts of a mixed tide- and wave-influenced
delta (Ftw or Fwt sensu Ainsworth et al. 2011; Legler et al. 2014) (Fig. 4H, I). In this context,
channelized cross-bedded sandstones with high basal erosional relief (facies HRTC) are interpreted as
fluvio-tidal distributary channel-fills (Ft or Tf sensu Ainsworth et al. 2011) (Legler et al. 2014; Van
Cappelle et al. 2016). These may have been overdeepened by fluvial erosion during falling relative
sea-level or by tidal scour during later abandonment and transgression (Willis and Gabel 2003; Van
Cappelle et al. 2016).

Tide-dominated delta fronts

Description.--- In the northeast of the study area, some coarsening-upward sandstone successions (up
to 20 m thick) are laterally (to the west and east) and vertically juxtaposed against mudstones. These
form elongate, linear bodies, usually oriented north-south or southwest-northeast, or a series of such
bodies that are laterally amalgamated to form more sheet-like geometries. Detailed mapping reveals
that many of these sandstones are connected at their northern extremity to time-equivalent shoreline
deposits (Mellere and Steel 1995; Mellere and Steel 2000; Hampson et al. 2008). Hence, these laterally
discontinuous sand bodies have a superficial (along-strike) resemblance to some so-called “isolated”
shallow-marine sandstones (sensu Bergman and Snedden, 1999). In the lower part of each succession,
bioturbated mudstones (facies OS) coarsen gradually upward into bioturbated, muddy, very fine- to
fine-grained sandstones (faces DB/BT) (Fig. 4J). Where preserved, sedimentary structures include
wave-ripple and current-ripple cross-lamination and rare HCS. These structures are commonly
overprinted by a high degree of bioturbation (BI=3-5) (Boyles and Scott 1982; Tillman and Martinsen
1984; Gaynor and Swift 1988; Walker and Bergman 1993; Mellere and Steel 2000; Hampson et al.
2008). The intensity of bioturbation decreases upwards (BI=1-3) and the grain size increases upwards,
into heterolithic, fine- to medium-grained sandstones (facies MB/BC) (Fig. 4J). These sandstones
contain bidirectional current-ripple cross-lamination and unidirectional cross-beds with foresets lined
by mud drapes and heterolithic toesets containing current-ripple cross-lamination. Mud clasts are
common. Rare bidirectional cross-bedding and HCS are also present locally (Walker and Bergman
1993; Mellere and Steel 1995). These various structures are superimposed on low-angle (1-3°) inclined surfaces. The upper part of a typical coarsening-upward succession comprises erosionally-based, clean, sparsely bioturbated (BI=1-3), fine- to medium-grained, cross-bedded sandstones with unidirectional paleocurrents (facies PB/BT). The basal erosional surfaces of this facies are generally low relief and laterally discontinuous, but lie parallel to, and pass downwards into, the inclined surfaces of facies MB/BC (Boyles and Scott 1982; Tillman and Martinsen 1984; Gaynor and Swift 1988; Walker and Bergman 1993; Hampson et al. 2008). In proximal (northerly) locations, the erosional base of the uppermost facies has higher local relief, forming channelized geometries (Mellere and Steel 2000). Paleocurrents in facies MB/BC and PB/BT are mainly unidirectional towards the south and southwest, and with the following orientations: (1) sub-parallel to the regional shoreline trend; (2) parallel to the elongation trend of the discontinuous sandstone bodies; and (3) oblique to the strike of inclined surfaces within the discontinuous sandstone bodies.

Such shoreline sandstones predominate in the following lithostratigraphic units (Fig. 2): the Wise Gulch Sandstone, Berry Gulch Sandstone, and Morapos Sandstone of the eastern Uinta Basin and Sand Wash Basin (Boyles and Scott 1982; Kiteley and Field 1984; Hampson et al. 2008); the Kremmling Sandstone and Hygiene Sandstone in the Middle Park Basin (Kiteley and Field 1984); the Hygiene Sandstone in the Colorado Front Range (Kiteley and Field 1984; Steel et al. 2012); and the Haystack Mountain Formation of the eastern Hanna Basin (Mellere and Steel, 1995, 2000).

**Interpretation.—** The upward increase in sandstone grain size and cross-stratification, from current-ripple cross-lamination (facies MB/BC) to dune-scale cross-bedding (facies PB/BT), indicates an upward increase in flow velocity and implies upward shallowing. Deposition of cross-beds on inclined surfaces that are oriented oblique to cross-bed paleocurrent directions indicates deposition on laterally accreting barforms (Boyles and Scott 1982; Gaynor and Swift 1988), which were elongated north-south (Boyles and Scott 1982; Tillman and Martinsen 1984; Hampson et al. 2008). The high bioturbation intensity and diversity of trace fossil assemblages in the lower part of the succession (facies DB/BT) indicates low deposition rates and fully marine salinities in the toes of these barforms (cf. MacEachern and Bann 2008). In facies DB/BT and MB/BC, the occasional presence of HCS (Walker and Bergman 1993; Hampson et al. 2008; Leva López et al. 2016) indicates episodic reworking of the sediment by storm waves (Dott and Bourgeois 1982; Duke 1985). The presence of wave-ripple cross-lamination throughout the succession also provides evidence for wave-influence. Abundant bidirectional current-ripple cross-lamination and pervasive mud drapes along cross-bed foresets and toesets (facies MB/BC and PB/BT) are indicative of tidal currents (Mellere and Steel 2000; Hampson et
Most workers agree on the description and depositional process interpretations outlined above, but interpretations of depositional environment vary widely, depending on the emphasis placed on the presence or absence of associated facies, and on stratigraphic relationships and context (Snedden and Bergman 1999; Suter and Clifton 1999). Previous interpretations include: (1) deposition as shelf-ridges on a wave-dominated shelf (Tillman and Martinsen 1984, 1987; Gaynor and Swift 1988) (W sensu Ainsworth et al. 2011); (2) the “isolated” shallow-marine sandstones have been interpreted in a sequence stratigraphic framework as tide-dominated, estuarine incised valley fills (Bergman 1999) (Tfw sensu Ainsworth et al. 2011) that were eroded directly into marine mudstones during relative sea-level falls and lowstands; and 3) deposition as lowstand shorelines during forced-regressions from which thin coastal-plain deposits were removed by later transgressive erosion, either as fully land-attached shorefaces (Walker and Bergman 1993; Bergman 1994; Bergman and Walker 1995) or spits (Nielsen and Johannessen 2008) (W sensu Ainsworth et al. 2011). The most recent, and our preferred, interpretation is that of southward-prograding tide-dominated deltas (Mellere and Steel 1995; Suter and Clifton 1999; Mellere and Steel 2000; Hampson et al. 2008; Steel et al. 2012) (Ftw sensu Ainsworth et al., 2011) (Fig. 4K, L). This interpretation supports the main observation of gradationally based, coarsening-upward successions, and the tide-dominance is in agreement with facies analysis. Some wave reworking (HCS, wave-ripples) can also be found in modern tide-dominated deltas (Dalrymple et al., 2003), especially those facing the direction of oceanic storm waves. By analogy with modern tide-dominated deltas, the elongated coarsening upward successions in the WIS are interpreted as having been deposited as fluvial- and ebb-tide-dominated mouth bars in a sub-tidal delta platform setting. Modern analogues, such as the Ganges-Brahmaputra and Chang Jiang (Yangtze), display a wide range of large tidal bedforms, including isolated and coalesced elongate sand ridges (e.g. Kuehl et al. 1997, 2005; Hori et al. 2001, 2002; Berné et al. 2002; Goodbred and Saito 2012; Xing et al. 2012; Liu et al. 2013; Xu et al. 2016; Yang 1989).

**Barrier islands, back-barrier lagoons, and estuaries**

**Description.**— Fluvial-dominated delta front and, more commonly, wave-dominated shoreface successions are overlain by thin (up to 20 m) successions of laterally continuous, heterolithic, carbonaceous mudstones, coals and lenticular, heterolithic sandstones which extend paleo-landward of their associated, underlying coarsening-upward successions (Fig. 4M). The laterally continuous, carbonaceous mudstones are interbedded with planar-laminated and wave-ripple cross-laminated
siltstones and very fine- to medium-grained, tabular sandstones on a decimetre- to metre-scale. Bioturbation intensity is variable (BI=0-4), but always with a low diversity trace-fossil suite of just Planolites, Ophiomorpha and Thalassinoides (facies LG). These heterolithic deposits and the coarsening-upward successions that they overlie are locally cut into by sharp-based, lenticular and channelized, fining-upward, very fine- to medium-grained sandstones (facies TC). Sedimentary structures in the sandstones include planar lamination, current-ripple cross-lamination, and cross-bedding with bidirectional paleocurrent directions, which are locally organised on inclined surfaces. Bioturbation intensity is variable (BI=0-4) and trace fossils include Planolites, Paleoophycus, Ophiomorpha and Thalassinoides (Donselaar 1989; Devine 1991; Kamola and Van Wagoner 1995; Olsen et al. 1999; Jordan et al. 2016). Less commonly observed are sharp-based, lenticular-to-tabular, well-sorted, very fine- to medium-grained sandstones containing bidirectional cross-bedding and planar to low-angle cross-stratification which cut down into the top of wave-dominated shoreface successions in paleo-seaward locations (facies TI) (Donselaar 1989; Olsen et al. 1999; Jordan et al. 2016). These sandstones contain sparse-to-moderate bioturbation (BI=0-4) by Planolites, Paleoophycus, Skolithos, Ophiomorpha and Thalassinoides, and some have discontinuous lags of shell hash and oyster shells along their bases (Hampson et al. 2011; Jordan et al. 2016). Lenticular and convex-upward, coarsening-upward, heterolithic, very fine- to medium-grained sandstones are also found in paleo-seaward locations. Sedimentary structures include bidirectional current-ripple cross-lamination and cross-beds that were deposited on clinoforms which dip paleo-landward (facies FTD) (Kamola and Van Wagoner 1995; York et al. 2011). In more paleo-landward positions, lenticular and convex-upward, coarsening-upward, heterolithic, carbonaceous, fine- to coarse-grained sandstones are present, containing current-ripple cross-lamination and wave ripples which were deposited on paleo-seaward dipping clinoforms (facies BHD) (Donselaar 1989; Olsen et al. 1999; Kamola and Van Wagoner 1995).

Shoreline sandstones of this class form a subordinate component of the following lithostratigraphic units (Fig. 2): the Point Lookout Sandstone, Cliff House Sandstone, and Pictured Cliffs Sandstone of the San Juan Basin (Palmer and Scott 1984; Donselaar 1989; Devine, 1991; Olsen et al. 1999; Jordan et al. 2016); the Muley Canyon Sandstone of the Henry Basin (Birgenheier et al. 2009; Seymour and Fielding 2013); the Star Point Sandstone, shallow-marine members of the Blackhawk Formation, and distal Castlegate Sandstone in the Uinta Basin (e.g., Van Wagoner et al. 1990; Kamola and Van Wagoner 1995; Van Wagoner 1995; Yoshida 2000; Hampson et al. 2008; Hampson et al. 2011; Gani et al. 2015); the upper Sego Sandstone and shallow-marine members of the Mount Garfield Formation of the southern Piceance Basin (Hettinger and Kirschbaum 2002; Kirschbaum and Hettinger 2004; Aschoff and Steel 2011a; Madof et al. 2015); the Sego Sandstone, Iles Formation, and Twentymile
Sandstone of the northern Piceance Basin and Sand Wash Basin (Masters 1967; Kiteley and Field 1984; Seidler and Steel 2001; Gomez-Veroiza and Steel 2010; York et al. 2011; Painter et al. 2013); the Rock Springs Formation in the Rock Springs Uplift area (Roehler 1978, 1990); the Haystack Mountain Formation of the Washakie Basin and western Hanna Basin (Mellere and Steel, 1995, 2000); and the Eagle Formation and Judith River Formation in the Bighorn Basin (Fitzsimmons and Johnson 2000).

Interpretation.--- The low diversity of trace fossil assemblages indicates deposition in stressed, marginal marine environments (MacEachern and Bann 2008). Fine-grained carbonaceous mudstones (facies LG) were deposited in a low-energy, brackish water environment in close proximity to the vegetated part of the coastal plain, which provided the carbonaceous debris (Kamola and Van Wagoner 1995). Therefore it is interpreted that deposition of facies LG took place in a protected lagoon behind a barrier island (Fig. 4O). The tabular sandstones in this facies either represent the product of washovers over the barrier during storms, or the product of crevassing during river floods (Donselaar 1989; Devine 1991; Olsen et al. 1999; Kamola and Van Wagoner 1995; Hampson et al. 2011). The well-sorted, lenticular-to-tabular sandstones at the top of wave-dominated shoreface successions (facies TI) are interpreted to be the result of reworking of shoreface sands, probably in tidal inlets, at the paleo-seaward margin of the lagoon (Donselaar 1989; Olsen et al. 1999; Hampson et al. 2011) (Fig. 4O). Coarsening-upward, lenticular and convex-upward sandstones containing paleo-landward-dipping clinoforms (facies FTD) are interpreted as flood tidal deltas that were genetically-related to tidal inlets (Kamola and Van Wagoner 1995) (Fig. 4O). Coarsening-upward successions that are richer in carbonaceous material and contain paleo-seaward-dipping clinoforms, and which are found in more paleo-landward locations (facies BHD), have been interpreted as prograding bayhead deltas (Donselaar 1989; Olsen et al. 1999; Kamola and Van Wagoner 1995) (Fig. 4O). Fining-upward, channelized, inclined heterolithic sandstones with bidirectional paleocurrent directions (facies TC) are interpreted as tidal channels (Donselaar 1989; Devine 1991; Olsen et al. 1999; Kamola and Van Wagoner 1995) or tide-influenced fluvial channels (Kirschbaum and Hettinger 2004; Gomez-Veroiza and Steel 2010; Aschoff and Steel 2011a). These channels could have been part of bayhead deltas (Donselaar 1989) or the marine shoreline (Devine 1991), or they may represent transgressed fluvial channels that effectively became estuaries when they were abandoned (Gomez-Veroiza and Steel 2010) (Fig. 4O). Overall, these various facies formed in a more paleo-landward position than the underlying wave-dominated shoreface and fluvial-dominated delta front deposits. The dominant depositional process in barrier island systems is waves. Tides play a role in keeping breached barriers open to create tidal inlets and associated flood tidal deltas (Wt sensu Ainsworth et al. 2011) (Fig. 4N). Additionally, there is evidence for fluvial processes in the form of bayhead deltas and tide-influenced
fluvial channels (Wtf or Wft sensu Ainsworth et al. 2011) (Fig. 4N). Except for wave ripples, the back-barrier facies of this succession contain little direct evidence for wave processes, although all workers agree on a wave-dominated setting based on their close association with underlying shoreface and paleo-seaward-lying barrier island successions.

Stacking of shoreline-sandstone successions

There is a significant discrepancy between the spatial and temporal scales of the shoreline-sandstone successions documented above (Fig. 4), and those of the “timeslice” paleogeographic reconstructions presented below (Figs. 5-9). The upward-coarsening successions described above for wave-dominated shoreface, fluvial-dominated delta front, mixed tide- and wave-influenced delta front, and tide-dominated delta front deposits record a single regressive transit of the shoreline, and correspond to parasequences (sensu Van Wagoner et al. 1990). Their transgressive counterparts comprise successions of barrier island, back-barrier lagoon, and estuarine deposits in paleo-landward locations and thin, correlative lags and open-marine-mudstone intervals in paleo-seaward locations (e.g., Hwang and Heller 2002). These individual shoreline-sandstone successions are typically up to several tens of meters in thickness (Fig. 4). Each of the studied “timeslices” (Fig. 2) contains multiple regressive and transgressive successions of these types, which are arranged in regressive-transgressive tongues (cf. parasequences and their bounding flooding surfaces sensu Van Wagoner et al. 1990) that are stacked vertically to form successions up to several hundred meters thick. The stacked successions are characterised by progradational (net-regressive), retrogradational (net-transgressive) and aggradational stratal patterns (cf. parasequence sets sensu Van Wagoner et al. 1990).

Each “timeslice” map shows the interpreted depositional process regimes in the succession of stacked shoreline sandstones (i.e. in multiple stacked regressive-transgressive tongues). The range of shoreline-sandstone classes that are present within the stacked stratal succession at a particular part of the outcrop belt(s) is recorded using a ternary diagram, after the classification scheme of Ainsworth et al. (2011) (Figs. 5-9). The regressive and transgressive components of the stacked stratal succession are indicated separately, respectively by red and blue shaded regions of the ternary diagram, because there is typically a difference in interpreted depositional process regime between them. However, a consistent depositional process regime is generally interpreted in all regressive or transgressive components of a stacked shoreline-sandstone succession. Thus, each ternary diagram shows the time-averaged depositional process regimes for the regressive and transgressive components of a stacked shoreline-sandstone succession, at spatial and temporal scales comparable to the duration of the
Each “timeslice” map contains several ternary diagrams, each of which corresponds to part of an outcrop belt(s) (Figs. 5-9). Progradational and retrogradational stratal patterns of shoreline-sandstone stacking at a particular part of the outcrop belt(s) are indicated by red and blue arrows in each “timeslice” map (Figs. 5-9).

“TIMESLICE” PALEOGEOGRAPHIC RECONSTRUCTIONS

Spatial and temporal variations in depositional processes at the scale of stacked shoreline-sandstone successions (cf. parasequence sets *sensu* Van Wagoner et al. 1990) are synthesized in paleogeographic reconstructions for the five “timeslices” indicated in Figure 2 (Figs. 5-9), which build on and extend previous paleogeographic reconstructions (e.g., McGookey et al. 1972; Kauffman and Caldwell 1993). Two isopach maps are also presented (Fig. 10) based on published cross sections (Merewether et al. 1997; Molenaar et al. 2002; Anna 2012). The coastal morphology and depositional processes along the paleoshoreline are summarized separately below for each “timeslice”, with the shorelines described from south to north.

“Timeslice 1” (c. 83 Ma)

“Timeslice 1” corresponds to the Scaphites Hippocrepis I biozone (c. 83 Ma) (Fig. 2). This includes shoreline deposits of the Point Lookout Sandstone in the San Juan Basin, which record overall northeastward progradation during the early Campanian (Molenaar et al. 2002) (Fig. 5). The Point Lookout Sandstone contains regressive wave-dominated shoreface and transgressive barrier-island and back-barrier deposits (Fig. 4A-C, M-O) that are progradationally stacked (Devine, 1991).

Biostratigraphic dating in the Kaiparowits Plateau of south-central Utah is poor (Fig. 2). Radiometric dating of bentonites in the Wahweap Formation shows that this formation is Campanian in age (Jinnah et al. 2009), whereas the John Henry Member of the underlying Straight Cliffs Formation contains Desmocaphites ammonites, which are older than Campanian (Eaton 1991). Therefore coastal-plain and fluvial strata of the lower and middle Wahweap Formation are included in this “timeslice” (Jinnah and Roberts 2011) (Fig. 5).

In the Wasatch Plateau and Book Cliffs area of east-central Utah, the Star Point Sandstone and the
lowermost part of the Blackhawk Formation are mapped in this “timeslice”. These strata contain regressive wave-dominated shoreface and fluvial-dominated, wave-influenced delta front deposits (Fig. 4A-C, D-F), and transgressive barrier-island and back-barrier lagoon deposits (Fig. 4M-O) (Van Wagoner et al. 1990; Kamola and Van Wagoner 1995; Hampson et al. 2011; Gani et al. 2015). Shoreline sandstones in the Star Point Sandstone are stacked progradationally in the southern Wasatch Plateau and retrogradationally in the northern Wasatch Plateau (Fig. 5) (Hampson et al. 2011). Also in this “timeslice”, fluvial-dominated, wave-influenced deltas of the Panther Tongue co-existed lateral to wave-dominated strandplains, with fluvial-dominated, wave-influenced deltas showing a southward-deflected, asymmetrical planform geometry (Hampson et al. 2011; Forzoni et al. 2015).

This planform morphology is attributed to wave-generated longshore currents as evidenced by southward-directed paleocurrents in adjacent wave-dominated shoreface successions (Slingerland and Keen 1999; Hampson et al. 2011). The successions of the Star Point Sandstone and lower part of the Blackhawk Formation are correlated to gravity-flow siltstones and sandstones in the Prairie Canyon Member of the Mancos Shale, which are interpreted to have been supplied from two distinct sources lying to the west and south, based on paleocurrents measured at outcrop and subsurface mapping using well-log data (Johnson 2003; Hampson 2010) (Fig. 5). In the Blackhawk Formation, gravity-flow siltstones and sandstones fed from the west have been correlated and linked to the position of fluvial channels that cut into wave-dominated shoreface successions in the contemporaneous coastline (Pattison 2005; Pattison et al. 2007). Gravity-flow siltstones and sandstones fed from the south constructed subaqueous clinoforms up to 200 m high (Johnson 2003). It is possible that they were supplied via rivers that fed the wave-dominated or wave-influenced deltas of the Muley Canyon Sandstone (Birgenheier et al. 2009; Seymour and Fielding 2013) or that fed now-eroded shorelines between the Henry Basin and San Juan Basin outcrop belts.

In Wyoming, this “timeslice” correlates to regressive wave-dominated shoreface and delta front (Fig. 4A-C), and transgressive barrier-island and back-barrier lagoon successions (Fig. 4M-O) of the lower part of the Rock Springs Formation in the Rock Springs Uplift area (Roehler 1978, 1990) and the Eagle Formation in the Bighorn Basin (Fitzsimmons and Johnson 2000). Shoreline sandstones in the Rock Springs Formation and Eagle Formation are progradationally stacked (Fig. 5) (Roehler 1978, 1990; Fitzsimmons and Johnson 2000).

“Timeslice 2” (c. 79 Ma)

“Timeslice 2” corresponds to the Baculites sp. (smooth) biozone (c. 79 Ma) (Fig. 2). In the San Juan
Basin, this comprises regressive wave-dominated shoreface (Fig. 4A-C) and transgressive barrier-island and back-barrier lagoon deposits (Fig. 4M-O) of the middle Cliff House Sandstone (Fig. 6) (Olsen et al. 1999; Molenaar et al. 2002). Subsurface mapping using well logs indicates that asymmetric sandstone bodies in the Cliff House Sandstone, possibly wave-dominated deltas, were deflected to the southeast, probably due to wave-driven longshore currents (Palmer and Scott 1984). Shoreline sandstones in the lower-to-middle Cliff House Sandstone are retrogradationally stacked (Fig. 6) (Olsen et al. 1999; Molenaar et al. 2002).

In east-central Utah and western Colorado, regressive wave-dominated shoreface successions (Fig. 4A-C) of the upper part of the Blackhawk Formation and lower Castlegate Sandstone were deposited with a progradational stacking pattern (Van Wagoner 1995; Yoshida 2000; Hampson et al. 2008). These successions were fed by rivers of the Castlegate Sandstone, while the shoreline was reworked by southward-directed longshore currents (Hampson et al. 2008; Hampson 2010) (Fig. 6).

In northern Wyoming, the regressive wave-dominated shoreface (Fig. 4A-C) and transgressive barrier-island and back-barrier lagoon deposits (Fig. 4M-O) of the Judith River Formation were deposited with a progradational stacking pattern in the Bighorn Basin (Fitzsimmons and Johnson 2000). In southern Wyoming, the Haystack Mountain Formation was deposited. The western exposures of this formation comprise regressive wave-dominated shoreface deposits (Fig. 4A-C) (Mellere and Steel, 1995, 2000). However, further east, the formation consists of regressive tide-dominated delta-front deposits (Fig. 4J-L) that are attached to regressive wave-dominated shoreface sandstones (Fig. 4A-C) (Mellere and Steel, 1995, 2000). These shoreline-connected, tide-dominated deltas correlate further south, in the Sand Wash Basin, to the apparently “isolated” tide-dominated delta-front deposits (Fig. 4J-L) of the Wise Gulch Sandstone, Berry Gulch Sandstone and Morapos Sandstone (Boyles and Scott 1982; Kiteley and Field 1984; Hampson et al. 2008). Paleocurrents in these sandstones are southwest-directed, sub-parallel to the regional paleoshoreline trend. It has been interpreted that this paleocurrent direction is the result of ebb-tide dominance in a southward prograding tide-dominated delta (Fig. 6) (Hampson et al. 2008; cf. Suter and Clifton 1999). This delta complex prograded southwards, which was, unusually, approximately parallel to the contemporary and preceding north-south shoreline trends (cf. Figs. 5, 6). This contributed to an increase in the length of the embayment that is referred to as “Utah Bight” (increasing from 225 km in “timeslice 1” to 270 km in “timeslice 2”). Increased tidal conditions have been noted at the head of the embayment at this time (Yoshida, 2000).
“Timeslice 3” (c. 77 Ma)

“Timeslice 3” corresponds to the Baculites Reduncus biozone (c. 77 Ma) (Fig. 2). Perhaps the most conspicuous feature of this “timeslice” is the pronounced embayed morphology of the “Utah Bight”, which had increased in length to approximately 450 km (Fig. 7).

In the San Juan Basin, New Mexico, “timeslice 3” encompasses the regressive wave-dominated shoreface (Fig. 4A-C) and transgressive barrier-island and back-barrier lagoon deposits (Fig. 4M-O) of the upper Cliff House Sandstone and Pictured Cliffs Sandstone. Shorelines prograded to the northeast and contain evidence for southeast-directed longshore currents along a straight, linear coast (Fig. 7) (Donselaar 1989; Molenaar et al. 2002; Jordan et al. 2016). Shoreline sandstones in the upper Cliff House Sandstone are aggradationally stacked (Fig. 6) (Donselaar 1989; Molenaar et al. 2002; Jordan et al. 2016).

In the Uinta Basin, east-central Utah, and Piceance Basin, west-central Colorado, this period includes the regressive mixed tide- and wave-influenced deltaic deposits (Fig. 4G-I) of the lower Sego Sandstone (Willis and Gabel 2001, 2003; Legler et al. 2014; Burton et al. 2016; Van Cappelle et al. 2016). Shoreline sandstones are progradationally stacked in the lower Sego Sandstone (Fig. 7) (Willis and Gabel 2001; Legler et al. 2014; Burton et al. 2016; Van Cappelle et al. 2016). Shoreline orientation switched abruptly from approximately northwest-southeast (Cliff House Sandstone) to northeast-southwest (lower Sego Sandstone); this change coincided with the approximate axis of the “Utah Bight”. Deltaic-dominated progradation on the northern side of the embayment was consistently towards the southeast (Fig. 7).

In the Sand Wash Basin, northwestern Colorado, “timeslice 3” is represented by regressive wave-dominated shoreface deposits (Fig. 4A-C) of the Iles Formation, which show progradation to the southeast and progradational stacking (Fig. 7) (Izett et al. 1971; Kiteley and Field 1984; Gomez-Veroiza and Steel 2010). Based on the abundance of fluvial and tide-influenced channels in the coeval coastal plain deposits (Gomez-Veroiza and Steel 2010), it is interpreted that these wave-dominated shoreface deposits were also influenced by fluvial and tidal currents.

In northern Colorado, the apparently “isolated” Hygiene Sandstone (Griffitts 1949; Kiteley and Field 1984) is interpreted to comprise regressive tide-dominated delta-front deposits, which prograded towards the southeast (Steel et al. 2012) (Figs. 4J-L, 7). This suggests that deltaic deposition dominated
the whole northern margin of the “Utah Bight” at this time (Fig. 7).

In the Powder River Basin, Wyoming, the Parkman Sandstone has been interpreted to comprise regressive wave-dominated shoreface deposits (Fig. 4A-C), which are progradationally stacked (Fig. 7) (Hubert et al. 1972). There was a dramatic change in shoreline orientation (approximately north-south), direction of progradation (mainly eastwards) in this northern area and the dominant sedimentary processes; the latter more closely resemble the wave-dominated shorelines of the southern margin of the “Utah Bight” (Fig. 7).

“Timeslice 4” (c. 75 Ma)

“Timeslice 4” corresponds to the Exiteloceras Jenneyi biozone (c. 75 Ma) (Fig. 2). In the San Juan Basin, New Mexico, regressive wave-dominated shoreface deposits (Fig. 4A-C) of the Pictured Cliffs Sandstone are progradationally stacked (Fig. 8) (Tokar and Evans 1993; Molenaar et al. 2002).

On the southern flank of the Piceance Basin, west-central Colorado, the Corcoran Member of the Mount Garfield Formation was deposited, while on the northern flank of the Piceance Basin and in the Sand Wash Basin, northwestern Colorado, the Trout Creek Member of the Iles Formation was deposited. Both of these units consist of regressive wave-dominated shoreface successions (Fig. 4A-C) that are progradationally stacked (Fig. 8) (Kiteley and Field 1984; Hettinger and Kirschbaum 2002; Kirschbaum and Hettinger 2004; Gomez-Veroiza and Steel 2010; Aschoff and Steel 2011a; Madof et al. 2015). Transgressive barrier-island and back-barrier lagoon deposits (Fig. 4M-O) have also been reported in the Sand Wash Basin (Masters 1967), in the context of progradationally stacked shoreline sandstones. Time-equivalent coastal plain deposits of the Neslen, Mount Garfield and Iles formations contain fluvial and tidally influenced channel-fill deposits (Hettinger and Kirschbaum 2002; Kirschbaum and Hettinger 2004; Gomez-Veroiza and Steel 2010; Aschoff and Steel 2011a; Shiers et al. 2014; Madof et al. 2015). Therefore, it is inferred that the wave-dominated shoreface deposits were also affected by tidal currents.

In northern Wyoming, the strata of “timeslice 4” have been eroded at the basal unconformity of the Teapot Sandstone Member (Figs. 2, 8) (Gill and Cobban 1966).

“Timeslice 5” (c. 73 Ma)
“Timeslice 5” corresponds to the Baculites Compressus biozone (c. 73 Ma) (Fig. 2). In the San Juan Basin, New Mexico, regressive wave-dominated shoreface deposits (Fig. 4A-C) of the uppermost Pictured Cliffs Sandstone are progradationally stacked (Fig. 9) (Tokar and Evans 1993).

In the Piceance Basin, west-central Colorado, regressive wave-dominated shoreface deposits (Fig. 4A-C) of the Rollins Sandstone Member of the Mount Garfield Formation were developed (Hettinger and Kirschbaum 2002; Kirschbaum and Hettinger 2004; Aschoff and Steel 2011a; Madof et al. 2015). The Twentymile Sandstone in the Sand Wash Basin, northwestern Colorado, also contains regressive wave-dominated shoreface and delta-front (Fig. 4A-C) and transgressive barrier-island and back-barrier lagoon deposits (Fig. 4M-O) (Masters 1967; Roehler 1990; Seidler and Steel 2001). Shoreline sandstones are progradationally stacked in the Rollins Sandstone Member and Twentymile Sandstone (Fig. 9) (Masters 1967; Roehler 1990; Hettinger and Kirschbaum 2002; Kirschbaum and Hettinger 2004; Aschoff and Steel 2011a; Madof et al. 2015).

In western, central and northern Wyoming, the alluvial-and-coastal-plain sheet sandstones of the Teapot Sandstone Member, Pine Ridge Sandstone, and Ericson Sandstone were developed (Fig. 9) (Gill and Cobban 1966; Gill et al. 1970). The base of the Teapot Sandstone Member is marked locally by an angular unconformity (Gill and Cobban 1966), which has been linked to the activity of Laramide-type thrust faults (Leva López and Steel 2015). Time-equivalent shoreline sandstones are present in the subsurface of the western part of the Powder River Basin (Fig. 9) (Fox 1993).

During this period, the coastal indentation of the “Utah Bight” was reduced (c. 200 km), the regional shoreline strike was oriented north-south, and coastal progradation was mainly eastwards (Fig. 9). Shorelines were wave-dominated with straight to linear morphologies; there is a notable paucity of tidal deposits.

**Isopach maps**

The thickness of Campanian strata (Fig. 10) provides a proxy for syn-depositional tectonic subsidence in the Western Interior Basin, because these strata comprise alluvial, coastal-plain and shallow-marine deposits that record nearly continuous and complete filling of available accommodation over their duration. Thicker strata therefore record the position of the seaway depocenter. In early Campanian times, including “timeslice 1”, the depocenter lay in the western part of the study area, close to some active structures in the Sevier fold-and-thrust belt (e.g., Gunnison Thrust; shown in red in Fig. 10A).
In the late Campanian, including “timeslices 3-5”, the depocenter had widened and migrated to the east, such that it lay eastward of active Laramide structures (shown in purple in Fig. 10B). The isopach maps define a large (c. 300-500 km diameter) and broadly circular depocenter, with up to 1 km of sedimentary fill. The paleoshoreline position also migrated eastward through time, in the same way as the seaway depocenter. The strike of thickness contours in the western part of the map approximate the strike of late Campanian shorelines and also the morphology of the “Utah Bight” embayment. This implies a primary link between syn-depositional tectonic subsidence and the shape of the embayment.

**DISCUSSION**

**Critical application of depositional process classification scheme**

Depositional process regime can be interpreted and classified at a variety of spatial scales (e.g., from “element” to “element complex assemblage set” in the architectural scheme of Vakarelov and Ainsworth 2013). This approach utilizes a combination of facies analysis and architectural-element analysis of genetically-related deposits that are bounded by sequence stratigraphic surfaces. The regional focus of this study requires the interpretation of depositional process regime at the spatial scale of regressive parasequences and their transgressive counterparts (cf. “regressive element complex assemblage sets” and “transgressive element complex assemblage sets” of Vakarelov and Ainsworth 2013) (Fig. 4), and stacked successions of these regressive and transgressive elements (cf. “parasequence set” of Van Wagoner et al. 1990). The interpretation of depositional process regime at such large spatial scales cannot capture smaller-scale variability. For example, we interpret an assemblage of barrier island and back-barrier lagoon deposits (constituting a “transgressive element complex assemblage set” of Vakarelov and Ainsworth 2013) to be wave-dominated and tide-influenced (Fig. 4M-O), but some of its components are tide-dominated (e.g., tidal inlet, TI, and flood tidal delta, FTD, deposits) or even fluvial-dominated (e.g., bayhead delta deposits, BHD). Differences in the preservation potential of these small-scale components must also be acknowledged in interpreting depositional process regime at large spatial scales. For example, wave-dominated beach ridge sets and associated shorefaces are poorly preserved in transgressive barrier-island deposits, because they are erosionally reworked during shoreface retreat (e.g., Swift 1968; Sixsmith et al. 2008), even though their inferred presence is integral to the interpretation of an overall wave-dominated process regime (Fig. 4M-O). The effects of small-scale variability and preservational biases can only be...
mitigated by detailed facies analysis and architectural-element analysis in the context of a sound sequence stratigraphic framework, prior to any interpretation of depositional process regime.

Data quantity, quality and distribution also determine confidence in interpreting depositional process regime. Data must sample representatively the studied deposits at the scale of interest in order to support a robust interpretation. The potential influence of sampling bias becomes more important as the attempted precision of an interpretation increases. Thus, semi-quantitative percentage-based classification of depositional process regime (e.g., Ainsworth et al. 2011; Rossi et al. 2017) is increasingly susceptible to both sampling bias, and human error (e.g., misinterpretation of facies). Our interpretations of depositional process regime are based on well-documented outcrop case studies that are unusually data-rich as a result of the large scale, continuous character, and near-complete exposure of Campanian outcrop belts in the Western Interior Basin. Nonetheless, our depositional process regime interpretations only allow broad areas in the “process space” of the ternary classification scheme to be defined (Fig. 4B, E, H, K, N).

Finally, different workers have argued for varying interpretations of some of these deposits, based on the application of different conceptual models to sparse datasets from incompletely preserved depositional systems over several decades of study (since the 1970’s). For example, the “isolated” shallow-marine sandstones that are here interpreted as part of tide-dominated deltas (Ft sensu Ainsworth et al. 2011) have been interpreted alternatively as wave-dominated shoreface deposits (Walker and Bergman 1993; Bergman 1994; Bergman and Walker 1995) (W sensu Ainsworth et al. 2011), storm-generated sand ridges (Tillman and Martinsen 1984, 1987; Gaynor and Swift 1988) (W sensu Ainsworth et al. 2011), and tide-dominated estuaries developed in incised valleys (Bergman 1999) (Tf sensu Ainsworth et al. 2011). This wide variety of interpretation makes it difficult to define consistently and confidently the corresponding “process space” in a ternary classification scheme. This problem is exacerbated where mixed depositional processes are recognized (e.g., Collins et al. 2018). The detail and rigor of descriptive documentation is critical in evaluating the various interpretations that may have been proposed for a particular shoreline sandstone. In addition, the variety of interpretations can reflect the quality of data support and internal logic of the proposed depositional models. This type of uncertainty is highlighted in the literature synthesis component of this study, and could be mitigated by collection of additional primary data at outcrop if time and resources allowed.

Spatial and temporal variation in depositional process regime
In all five “timeslices”, the northwest-southeast-striking paleoshorelines in the south of the study area, in the San Juan Basin of New Mexico, consisted of regressive wave-dominated shorefaces and transgressive wave-dominated barrier islands at the scale of interest (Donselaar 1989; Devine 1991; Tokar and Evans 1993; Olsen et al. 1999; Jordan et al. 2016) (Figs. 5-9). In the north-south-striking paleoshorelines in the northern part of the study area, comprising the Great Divide, Wind River, Bighorn and Powder River basins of central and northern Wyoming, regressive wave-dominated shorefaces and transgressive barrier islands also dominate in all “timeslices” that contain detailed case studies (Hubert et al. 1972; Roehler 1990; Fitzsimmons and Johnson 2000) (Figs. 5-7). Hence, paleoshorelines in both the northern and southern parts of the study area were wave-dominated throughout the Campanian.

In contrast, a broader range of depositional process regimes is evident in the “Utah Bight” embayment of the central part of the study area, comprising the Uinta, Piceance, Sand Wash, Washakie, Hanna and Laramie basins of Utah, Colorado and southern Wyoming (Figs. 5-9). In “timeslice 1” (Fig. 5), when the paleoshoreline lay in its most paleo-landward position, regressive wave-dominated shorefaces, regressive fluvial-dominated deltas, and transgressive wave-dominated barrier islands were developed (Kamola and Van Wagoner 1995; Olariu et al. 2010; Hampson et al. 2011; Seymour and Fielding 2013). In “timeslice 2” (Fig. 6), the paleoshoreline was more embayed and contained two distinctive settings: (1) regressive wave-dominated shorefaces and transgressive wave-dominated barrier islands (Van Wagoner 1995; Yoshida 2000); and (2) regressive tide-dominated deltas (Mellere and Steel 1995, 2000; Hampson et al. 2008). This “timeslice” records the first notable increase in tidal influence in regressive shoreline deposits along the northern margin of the embayment. In the third “timeslice” (Figure 7), there are three distinct paleoshoreline settings: (1) regressive mixed tide- and wave-influenced deltas (Sego Sandstone) in the Uinta and Piceance Basins (Willis and Gabel 2001, 2003; Legler et al. 2014; Van Cappelle et al. 2016); (2) regressive tide-dominated deltas (Hygiene Sandstone) in the Middle Park Basin and Colorado Front Range (Steel et al. 2012, after Kiteley and Field 1984); and (3) an intervening area comprising regressive wave-dominated shorefaces in the Sand Wash Basin (Kiteley and Field 1984; Gomez-Veroiza and Steel 2010). The “Utah Bight” was most strongly embayed during this “timeslice” (Figure 7). In “timeslice 4” and “timeslice 5” (Figs. 8, 9), the paleoshoreline advanced further eastwards, resulting in progressive infilling of the “Utah Bight” embayment. This resulted in a less embayed paleoshoreline characterized by regressive wave-dominated shorefaces, regressive fluvial-dominated deltas, and transgressive wave-dominated barrier islands (Masters 1967; Kiteley and Field 1984; Kirschbaum and
Peterson 2004; Gomez-Veroiza and Steel 2010; Aschoff and Steel 2011a). However, coeval coastal-plain deposits continued to contain tidally influenced channel-fills (Masters 1967; Kiteley and Field 1984; Kirschbaum and Hettinger 2004; Gomez-Veroiza and Steel 2010; Aschoff and Steel 2011a). Potential tidal resonance in the “Utah Bight” embayment.--- Tidal processes in regressive shoreline deposits were only prominent during “timeslice 2” and “timeslice 3”, when the “Utah Bight” had a pronounced embayed geometry (Figs. 6, 7, 11B), implying that tides may have been amplified on a regional scale at these times. Tidal amplification takes place in an embayment when its dimensions are in resonance with the quarter wavelength of the tidal wave; according to the following relationship (Pugh 1987; Sztanò and De Boer 1995):

\[ l = \frac{1}{4} T \sqrt{gh} \]  

(eq. 1)

in which \( l \) is the length of the embayment, \( T \) is the period of the tidal wave (12.42 hour for the principal semi-diurnal, M2, lunar tide), \( g \) is the gravitational constant (9.81 ms\(^{-2}\)), and \( h \) is the water depth. Paleo-water depths of 50-300 m have been typically estimated for the Western Interior Seaway during the Campanian (Erickson and Slingerland 1990). With these parameters, tidal resonance of the M2 tide would have taken place when the “Utah Bight” embayment had a length of 250-600 m. The length of the “Utah Bight” embayment during “timeslice 1”, “timeslice 4” and “timeslice 5” was 200-250 km (Figs. 5, 8, 9). However, during “timeslice 2” and “timeslice 3”, the embayment was 270-450 km long, which lies in the range for tidal resonance of the M2 tide for paleo-water depth ranges of 60-170 m (Figs. 6, 7). The small difference between the length of the “Utah Bight” embayment in wave-dominated “timeslice 4” (250 km) and tide-influenced “timeslice 2” (270 km) suggests that the embayment came into resonance with the M2 tide when the it was c. 260 km long, corresponding to a paleo-water depth of c. 55 m. Thus, tidal resonance of the “Utah Bight” embayment provides a plausible alternative to previous interpretations that tides were locally amplified due to funnelling along coastlines with increased rugosity, due to the widespread initiation and early growth of development of incipient Laramide structures in the middle Campanian (Aschoff and Steel 2011b; Steel et al. 2012). Although tidal resonance can account for the widespread occurrence of tide-dominated and tide-influenced shorelines during specific periods in the development of the “Utah Bight” embayment, it does not explain why these shorelines occur only along the northern margin of the embayment (Figs. 6, 7, 11B). The latter can be attributed to strong, wave-generated, southward-directed longshore drift.
along the western margin of the Western Interior Seaway, which created a headland at the northern
entrance to the “Utah Bight” embayment (Palmer and Scott 1984; Slingerland and Keen 1999;
Hampson and Howell 2005; Hampson et al. 2011). As a result, leeward paleoshorelines, which faced
north, were subject to a larger wave fetch and greater wave reworking than windward
paleoshorelines that faced south. This could further explain why the most tide-dominated part of the
shoreline was not simply located at the head of the embayment, as is the case in most modern
embayments, such as the German Bight (Reineck and Singh 1980), and in many ancient examples
(e.g., Yoshida et al. 2004; Wells et al. 2010a, 2010b).

Structural controls on evolution of the “Utah Bight” embayment.--- The overall position and
regional orientation of the paleoshoreline along the western margin of the Western Interior Seaway
throughout the Cretaceous was controlled by the north-south-striking Sevier fold-and-thrust belt
(DeCelles 2004; Yonkee and Weil 2015) and the northwest-southeast striking Mogollon Highlands to
the south (Dickinson and Lawton 2001) (Fig. 11). From the Cenomanian to the early Campanian
(“timeslice 1”; Figs. 2, 5, 10A), the paleoshoreline protruded eastwards of the Wyoming Salient of the
Sevier fold-and-thrust belt (McGookey et al. 1972), which formed the northeastern margin of the
“Utah Bight” (Fig. 11A) (DeCelles 1994; Yonkee and Weil 2010). The “Utah Bight” had a subdued
embayment geometry in the early Campanian, and was fringed by wave-dominated paleoshorelines
(Fig. 11A). During the middle Campanian (“timeslice 2” and “timeslice 3”; Figs. 2, 6, 7), Laramide
structures in the Wyoming foreland became active (Dorr et al. 1977; Lawton 1983; Lawton 1986;
Bryant and Nichols 1988; Shuster and Steidtmann 1988; Steidtmann and Middleton 1991; Miall and
Arush 2001; Leva López and Steel 2015), which forced the depocenter in Wyoming much further
eastward (Figs. 10B, 11B). However, Laramide structures were largely inactive in east-central Utah,
where the depocenter did not migrate as far east, thus enhancing the embayed geometry of the “Utah
Bight” (Fig. 10) and potentially initiating tidal resonance in the embayment (Fig. 11B). In the late
Campanian (“timeslice 4” and “timeslice 5”; Figs. 2, 8, 9), further eastward progradation of the
paleoshoreline in east-central Utah infilled the “Utah Bight” embayment, and coincided with a return
to wave-dominated paleoshorelines (Fig. 11C).

Application to data-poor basins.--- As outlined above, the morphological evolution of the “Utah
Bight” embayment during the Campanian, and the development of depositional process regime(s) in
the “Utah Bight”, was controlled to a large extent by the growth of active structures and related
patterns of tectonic subsidence. The latter can be deduced in basins that are much poorer in data than
the Western Interior Seaway, via analysis of stratigraphic relationships and thickness patterns (e.g., in seismic
data), and thus has potential to be used as a tool to infer regional depositional process regime.

Although such inferences will be inherently uncertain, for the reasons of large spatial scale and sparse data sampling outlined earlier, we suggest that estimating the dimensions and morphology of shallow-marine seaways and embayments will allow the potential for tidal amplification to be evaluated (cf. Sztanò and De Boer 1995) and, with paleogeographic maps of suitable quality, modelled numerically (Mitchell et al. 2010; Wells et al. 2010a, 2010b; Collins et al. 2018). When combined with available regional data describing paleoshoreline sedimentologic character, it is possible to make first-order predictions of temporal and spatial variations in coastal process regime. Such predictions do not conform simply to interpreted sequence stratigraphic setting, but must rather account for evolving basin morphology and context (cf. Yoshida et al. 2007; Ainsworth et al. 2008; Collins et al. in press).

CONCLUSIONS

Published sedimentologic descriptions of Campanian strata along the southwestern margin of the North American Western Interior Seaway (in New Mexico, Colorado, Utah and Wyoming, USA) have been critically reviewed and synthesized to define five different paleoshoreline types, each with a distinctive depositional process regime at relatively large spatial scales (e.g., regressive parasequences and their transgressive counterparts, parasequence sets): (1) regressive wave-dominated shorefaces and delta fronts; (2) regressive fluvial-dominated delta fronts; (3) regressive mixed tide- and wave-influenced delta fronts; (4) regressive tide-dominated delta fronts; and (5) transgressive barrier islands, back-barrier lagoons, and estuaries. The distribution and stacking pattern of these shoreline types has been compiled for five “timeslices”, each of which represents an ammonite biozone with a duration of up to several hundred thousand years. The five “timeslices” together illustrate the spatial distribution of depositional processes along the southwestern margin of the seaway, as preserved in the stratigraphic record, and can be related to the distribution of structural elements that were active during deposition.

The position of the Western Interior basin margin and associated paleoshoreline in the area of interest was controlled by the location of the Sevier fold-and-thrust belt to the west and the Mogollon Highlands to the south. The presence of the eastern-protruding Wyoming Salient in the Sevier fold-and-thrust belt resulted in the northern margin of a structural embayment of the basin margin in present-day Utah, and a broad, pre-existing embayment of the paleoshoreline (“Utah Bight”) was variably developed as a result.
Early Campanian paleoshorelines along the “Utah Bight” embayment were wave-dominated, except for fluvial-dominated deltas developed locally at points of sediment input.

The onset of Laramide thrusting in the foreland of Wyoming during the middle Campanian caused pronounced eastward progradation of paleoshorelines north of the “Utah Bight”. However, the position of the paleoshorelines did not change significantly in central Utah, therefore exaggerating the embayed geometry of the “Utah Bight”. Mixed-tide-and wave-influenced and tide-dominated deltas developed along the northern margin of the “Utah Bight” embayment at this time, suggesting a regional-scale tidal amplification. A possible explanation for this is that increased length of the embayment during this time period caused the embayment to be in resonance with the principal semi-diurnal (M2) tide. Paleoshorelines along the central and southern margins of the “Utah Bight” embayment, which were more exposed to the large wave fetch and wave-generated counterclockwise-circulating currents in the adjoining Western Interior Seaway, remained dominated by waves.

Regression during the late Campanian infilled the “Utah Bight”, resulting in the development of a near-linear paleoshoreline in the study area, which was dominated by wave action along its entire length.

The review, synthesis and analysis of previously published sedimentologic data presented here illustrates an approach that could be widely applied to other basins. The dense dataset for Campanian shorelines along the southwestern margin of the North American Western Interior Seaway is unusual, and reflects the availability of extensive, high-quality outcrop exposures. Consequently, our results have a relatively small degree of uncertainty in classifying shallow-marine depositional process regime at large spatial scales, which is probably inherent to the application of any shoreline classification scheme (i.e. irreducible uncertainty). When this sedimentologic synthesis is combined with an understanding of evolving structural evolution in the basin hinterland and tectonic subsidence in the basin itself, it is possible to account for temporal and spatial variations in coastal process regime.

ACKNOWLEDGEMENTS

The Department of Earth Science and Engineering of Imperial College is thanked for the award of a Janet Watson PhD scholarship to MvC. Additional funding by Shell International is gratefully acknowledged. We would like to thank Dan Collins and Chris Dean for discussions. Reviewers James Golab and Cari Johnson, Associate Editors Greg Ludvigson and Paul McCarthy, and Editor Leslie...
Melim are thanked for their critical, insightful and constructive reviews and editorial suggestions for improvement of the manuscript.

REFERENCES


COLLINS, D.S., AVDIS, A., ALLISON, P.A., JOHNSON, H.D., HILL, J., AND PIGGOTT, M.D., in press (a), Controls on tidal sedimentation and preservation: insights from numerical tidal modelling in the Late Oligocene–Miocene South China Sea, Southeast Asia: Sedimentology.


DECELLES, P.G., LAWTON, T.F., AND MITRA, G., 1995, Thrust timing, growth of structural 
culminations, and synorogenic sedimentation in the type Sevier orogenic belt, western United 

DEVINE, P.E., 1991, Transgressive origin of channeled estuarine deposits in the Point Lookout 
Sandstone, northwestern New Mexico; a model for Upper Cretaceous, cyclic regressive 

DICKINSON, W.R., AND GEHRELS, G.E., 2008, Sediment delivery to the Cordilleran foreland basin: 
insights from U-Pb ages of detrital zircons in Upper Jurassic and Cretaceous strata of the 

DICKINSON, W.R., AND LAWTON, T.F., 2001, Tectonic setting and sandstone petrofacies of the 

DONSELAAR, M.E., 1989, The Cliff House Sandstone, San Juan Basin, New Mexico; model for the 

DORR, J.A., SPEARING, D.R., AND STEIDTMANN, J.R., 1977, Deformation and deposition between 
a foreland uplift and an impinging thrust belt: Hoback Basin, Wyoming: Geological Society of 
America, Special Paper 177, 86 p.

DOTT, R.H., AND BOURGEEOIS, J., 1982, Hummocky stratification: significance of its variable 

DUKE, W.L., 1985, Hummocky cross-stratification, tropical hurricanes, and intense winter storms: 

EATON, J.G., 1990, Stratigraphic revision of Campanian (Upper Cretaceous) rocks in the Henry Basin, 

EATON, J.G., 1991, Biostratigraphic framework for the Upper Cretaceous rocks of the Kaiparowits 
Plateau, southern Utah, in Nations, J.D. and Eaton, J.G., eds., Stratigraphy, depositional 
environments, and sedimentary tectonics of the western margin, Cretaceous Western Interior 
Seaway: Geological Society of America, Special Paper 260, p. 47-64.

ENGHE, H., HOWELL, J.A., AND BUCKLEY, S.J., 2010, Quantifying cinothem geometry in a forced-
regressive river-dominated delta, Panther Tongue Member, Utah, USA: Sedimentology, v. 57, p. 
1750-1770.

ERICKSON, M.C., AND SLINGERLAND, R., 1990, Numerical simulations of tidal and wind-driven 
circulation in the Cretaceous Interior Seaway of North America: Geological Society of America 

FITZSIMMONS, R., AND JOHNSON, S., 2000, Forced regressions: recognition, architecture and 
genesis in the Campanian of the Bighorn Basin, Wyoming, in Hunt, D. and Gawthorpe, R.L.,
eds., Sedimentary responses to forced regression: Geological Society of London, Special Publication 172, p. 113-139.


HAMPSON, G.J., PROCTER, E.J., AND KELLY, C., 2008, Controls on isolated shelf sandstone ridges in the Cretaceous Western Interior Seaway, northern Utah and Colorado, USA, in Hampson,


SLINGERLAND, R., KUMP, L.R., ARTHUR, M.A., FAWCETT, P.J., SAGEMAN, B.B., AND
BARRON, E.J., 1996, Estuarine circulation in the Turonian Western Interior Seaway of North

SNEDDEN, J.W., AND BERGMAN, K.M., 1999, Isolated shallow marine sand bodies: deposits for all
interpretations, in Bergman, K.M. and Snedden, J.W., eds., Isolated shallow marine sand bodies:
sequence stratigraphic analysis and sedimentologic interpretation: SEPM, Special Publication 64,

STEEL, R.J., PLINK-BJÖRLUND, P., AND ASCHOFF, J.L., 2012, Tidal deposits of the Campanian
Western Interior Seaway, Wyoming, Utah and Colorado, USA. in Davis, R.A. and Dalrymple,

southern Wind River Range, Wyoming: implications for Laramide and post-Laramide
deforation in the Rocky Mountain foreland: Geological Society of America Bulletin, v. 103,
p. 472-485.

STORMS, J.E.A., AND HAMPSON, G.J., 2005, Mechanisms for forming discontinuity surfaces within
shoreface–shelf parasequences: sea level, sediment supply, or wave regime?: Journal of

SUTER, J.R., AND CLIFTON, H.E., 1999, The Shannon Sandstone and isolated linear sand bodies:
interpretations and realizations, in Bergman, K.M. and Snedden, J.W., eds., Isolated shallow
marine sand bodies: sequence stratigraphic analysis and sedimentologic interpretation: SEPM,
Special Publication 64, p. 321-356.

SWIFT, D.J.P., 1968, Coastal erosion and transgressive stratigraphy. Journal of Geology, v. 76, p. 444-
456.

SZTANÒ, O., AND BOER, P.L., 1995, Basin dimensions and morphology as controls on amplification

SZWARC, T.S., JOHNSON, C.L., STRIGHT, L.E., AND MCFARLANE, C.M., 2015, Interactions
between axial and transverse drainage systems in the Late Cretaceous Cordilleran foreland
basin: evidence from detrital zircons in the Straight Cliffs Formation, southern Utah, USA:

TAYLOR, A.M., AND GOLDRING, R., 1993, Description and analysis of bioturbation and

TAYLOR, D.R., AND LOVELL, R.W.W., 1995, High-frequency sequence stratigraphy and
paleogeography of the Kenilworth Member, Blackhawk Formation, Book Cliffs, Utah, USA, in
Van Wagoner, J.C., and Bertram, G.T., eds., Sequence Stratigraphy of Foreland Basin Deposits:


YOSHIDA, S., 2000, Sequence and facies architecture of the upper Blackhawk Formation and the lower Castlegate Sandstone (Upper Cretaceous), Book Cliffs, Utah, USA: Sedimentary Geology, v. 136, p. 239-276.


Figure 1

A) Regional tectonic framework and B) cross-section of western North America during the Late Cretaceous (modified after DeCelles 2004; Lawton and Bradford 2011; Yonkee and Weil 2015). C) Map of the study area showing the outcrops of Campanian strata, Tertiary (Laramide) basins and tectonically-elevated areas ('uplifts') in their present-day configuration (Tweto 1979; Love and Christiansen 1985; Hintze et al. 2000; Scholle 2003). The map also shows Sevier (thin skinned) thrusts (DeCelles 2004) and Laramide (thick skinned) structures that were active during the Campanian (see text for source references).

Figure 2

Chronostratigraphic, biostratigraphic, and lithostratigraphic framework of Campanian strata in Tertiary (Laramide) basins in the study area. The sedimentology of shallow-marine strata denoted by red lithostratigraphic names or black stars has been previously documented in detail (see top of diagram and text for source references). The ages of five mapped “timeslices” are indicated in blue. Chronostratigraphy after Gradstein et al. (2012).

Figure 3

Summary of the six main gross depositional environments (GDEs): (1) alluvial-to-coastal-plain sandstones; (2) coastal-plain coals, mudstones and sandstones; (3) shoreline sandstones; (4) marine mudstones; (5) gravity-flow siltstones and sandstones; and (6) marine marls and chalk. The same color coding is applied to Figure 2 and 5-9. Shoreline sandstones are subdivided into five classes, which are summarized in Figure 4.

Figure 4

Synthesis facies models of five interpreted shoreline sandstone classes, based on published literature (see text for details). A, D, G, J, M) Representative sedimentary logs, with regressive erosion surfaces shown in red, and various stratal surfaces developed during transgression shown in blue; B, E, H, K, N) inferred depositional process regime, plotted using the scheme of Ainsworth et al. (2011) (see text for details; W = wave, T = tidal, F = fluvial); and C, F, I, L, O) inferred plan-view morphology of shoreline showing location of representative sedimentary logs. A-C) Wave-dominated shoreface and delta-front deposits. OS= offshore marine mudstone, dLSF= distal lower shoreface, pLSF= proximal lower shoreface, USF= upper shoreface, FS= foreshore, C= coal mire. D-F) Fluvial-dominated delta
front deposits. OS= offshore marine mudstone, dDF= distal delta front, pDF= proximal delta front,
Two upward-shallowing, regressive successions separated by a flooding surface are portrayed. OS=
offshore marine mudstone, LSF= lower shoreface, LRTC= low relief tidal channel, TB= tidal bar,
HRTC= high relief tidal channel. J-L) Tide-dominated delta-front deposits. OS= offshore marine
mudstone, DB/BT= distal bar / bar toe, MB/BC= medial bar / bar center, PB/BT= proximal bar / bar top.
M-O) Barrier island, back-barrier lagoonal, and estuarine deposits. Deposits below the transgressive
surface and shown in grey represent older, regressive wave-dominated shoreface and delta-front
deposits (Fig. 4A-C); deposits above the transgressive surface and shown in black record deposition
during transgression. TI= tidal inlet, FTD= flood tidal delta, LG= lagoon, BHD= bayhead delta, TC=
tidal channel.

Figure 5
Paleogeographic map for “timeslice 1”, corresponding to ammonite biozone Scaphites Hippocrepis I
(c. 83 Ma) (Fig. 2). Interpretations of gross depositional environments (Fig. 3) are shown in bold colors
where constrained by outrops of strata contained in the “timeslice”, and in faded colors away from
outcrop control. Sketch line-drawings of paleoshorelines and ternary diagrams of interpreted
depositional process regime(s) are included where supported by detailed published work on
shoreline sandstones at outcrop (see key references and text for details). Regressive and transgressive
components of these shoreline sandstones are shown as red and blue fields, respectively, in the
ternary diagrams (cf. Fig. 4). In “timeslice 1”, the paleoshoreline occupied its most western position
and consisted mainly of wave-dominated shorefaces and some fluvial-dominated deltas (e.g., Star
Point Sandstone, Muley Canyon Sandstone) with related gravity-flow siltstones and sandstones
(lower part of Prairie Canyon Member, Mancos Shale). Marine marls and chalk (Niobrara Formation)
were deposited towards the center of the seaway.

Figure 6
Paleogeographic map for “timeslice 2”, corresponding to ammonite biozone Baculites sp. (smooth) (c.
79 Ma) (Fig. 2). Interpretations of gross depositional environments (Fig. 3) are shown in bold colors
where constrained by outrops of strata contained in the “timeslice”, and in faded colors away from
outcrop control. Sketch line-drawings of paleoshorelines and ternary diagrams of interpreted
depositional process regime(s) are included where supported by detailed published work on
shoreline sandstones at outcrop (see key references and text for details). Regressive and transgressive
components of these shoreline sandstones are shown as red and blue fields, respectively, in the
ternary diagrams (cf. Fig. 4). In “timeslice 2”, the paleoshoreline was predominantly wave-dominated, with a large tide-dominated delta (Haystack Mountains Formation, Morapos Sandstone, Berry Gulch Sandstone, Wise Gulch Sandstone) interpreted in southern Wyoming and northwestern Colorado.

**Figure 7**

Paleogeographic map for “timeslice 3”, corresponding to ammonite biozone Baculites Reduncus (c. 77 Ma) (Fig. 2). Interpretations of gross depositional environments (Fig. 3) are shown in bold colors where constrained by outrops of strata contained in the “timeslice”, and in faded colors away from outcrop control. Sketch line-drawings of paleoshorelines and ternary diagrams of interpreted depositional process regime(s) are included where supported by detailed published work on shoreline sandstones at outcrop (see key references and text for details). Regressive and transgressive components of these shoreline sandstones are shown as red and blue fields, respectively, in the ternary diagrams (cf. Fig. 4). In “timeslice 3”, the paleoshoreline was wave-dominated in the north (Parkman Sandstone Member) and south (Cliff House Sandstone), with a tide-dominated delta (Hygiene Sandstone Member) interpreted in northwestern Colorado, and mixed tide- and wave-influenced deltas (Sego Sandstone) interpreted in east-central Utah. This “timeslice” also corresponds to the maximum indentation (length) of the embayment at c. 450 km.

**Figure 8**

Paleogeographic map for “timeslice 4”, corresponding to ammonite biozone Exiteloceras Jenneyi (c. 75 Ma) (Figure 2). Interpretations of gross depositional environments (Fig. 3) are shown in bold colors where constrained by outrops of strata contained in the “timeslice”, and in faded colors away from outcrop control. Sketch line-drawings of paleoshorelines and ternary diagrams of interpreted depositional process regime(s) are included where supported by detailed published work on shoreline sandstones at outcrop (see key references and text for details). Regressive and transgressive components of these shoreline sandstones are shown as red and blue fields, respectively, in the ternary diagrams (cf. Fig. 4). In “timeslice 4”, the paleoshoreline was wave-dominated throughout the study area, and the length of the embayment has been reduced to c. 250 km.

**Figure 9**

Paleogeographic map for “timeslice 5”, corresponding to ammonite biozone Baculites Compressus (c. 73 Ma) (Figure 2). Interpretations of gross depositional environments (Fig. 3) are shown in bold colors where constrained by outrops of strata contained in the “timeslice”, and in faded colors away from outcrop control. Sketch line-drawings of paleoshorelines and ternary diagrams of interpreted
depositional process regime(s) are included where supported by detailed published work on
shoreline sandstones at outcrop (see key references and text for details). Regressive and transgressive
components of these shoreline sandstones are shown as red and blue fields, respectively, in the
ternary diagrams (cf. Fig. 4). In “timeslice 5”, the paleoshoreline was wave-dominated throughout the
study area.

**Figure 10**
Isopach maps for: A) lower Campanian strata (ammonite biozones Scaphites Leei III to Baculites
Asperiformis, 84-78 Ma, containing “timeslice 1”; Figs. 2, 5); and B) upper Campanian strata
(ammonite biozones Baculites Perplexus to Baculites Eliasi, 78-72 Ma, containing “timeslices 3, 4, and
5”; Figs. 2, 7, 8, 9). Colored lines represent interpreted paleoshoreline positions at maximum
regression in each “timeslice”, with associated ages in bold text. Thicknesses are taken from locations
(filled black circles) in published cross sections (Merewether et al. 1997; Molenaar et al. 2002; Anna
2012). Active Sevier and Laramide structures are shown in red and purple, respectively (OCU = Owl
Creek Uplift, RSU = Rock Springs Uplift, SMU = Sierra Madre Uplift, SRU = San Rafael Uplift, UU =
Uinta Uplift, WRU = Wind River Uplift). The seaway depocenter, which is represented by thicker
strata, and paleoshoreline position both migrated eastward through time.

**Figure 11**
Interpreted tectonic controls on basin geometry and depositional process regime. A) Early
Campanian wave-dominated paleoshorelines fringed the subdued embayment of the “Utah Bight”
(Fig. 5), the geometry of which was controlled by the position of the Sevier fold-and-thrust belt and
the Mogollon Highlands. B) In the middle Campanian, active Laramide structures forced the
paleoshoreline further eastward in Wyoming and enhanced the embayed geometry of the “Utah
Bight”, which may have caused tidal amplification and development of tide-dominated and tide-
influenced paleoshorelines (Figs. 6, 7). C) Continued eastward progradation of the paleoshoreline in
east-central Utah during the Late Campanian reduced the length of the “Utah Bight”, which was
again fringed by wave-dominated shorelines (Figs. 8, 9).
Fig. 1B

Fig. 1C

A

B

C

National boundary
State boundary
Fault
Plate boundary
Accreted terrains and accretionary wedge
Magmatic arc
Sevier fold-and-thrust belt
Foreland

Farallon Plate

Mesaverde Group outcrop (shallow marine sandstones and non marine siliciclastics)
Mancos Shale outcrop (marine mudstones)
Sevier thrust active during the Campanian (thin skinned)
Laramide thrust active during the Campanian (thick skinned)
Laramide anticline/dome active during the Campanian
Laramide basin (Campanian strata in the subsurface)

Accretionary wedge
Magmatic arc
Sevier fold-and-thrust belt
Western Interior Basin

Accreted terrains and accretionary wedge
Magmatic arc
Sevier fold-and-thrust belt
Foreland

Bighorn Basin
Powder River Basin
Wind River Basin
Granite Mountains Uplift
Laramie Uplift
Hartville Uplift

Rock Springs Uplift
Washakie Basin

Sierra Madre Uplift
Medicine Bow Uplift
Owl Creek Uplift

Laramie Uplift
Front Range Uplift

San Luis Uplift
Sawatch Range
San Juan Basin

San Rafael Uplift
White River Uplift

Uinta Uplift

Bighorn Uplift

Farallon Plate

Wind River Uplift

Fig. 1B

Mesaverde Group outcrop (shallow marine sandstones and non marine siliciclastics)
Mancos Shale outcrop (marine mudstones)
Sevier thrust active during the Campanian (thin skinned)
Laramide thrust active during the Campanian (thick skinned)
Laramide anticline/dome active during the Campanian
Laramide basin (Campanian strata in the subsurface)

Accretionary wedge
Magmatic arc
Sevier fold-and-thrust belt
Western Interior Basin

Accreted terrains and accretionary wedge
Magmatic arc
Sevier fold-and-thrust belt
Foreland

Bighorn Basin
Powder River Basin
Wind River Basin
Granite Mountains Uplift
Laramie Uplift
Hartville Uplift

Rock Springs Uplift
Washakie Basin

Sierra Madre Uplift
Medicine Bow Uplift
Owl Creek Uplift

Laramie Uplift
Front Range Uplift

San Luis Uplift
Sawatch Range
San Juan Basin

San Rafael Uplift
White River Uplift

Uinta Uplift

Bighorn Uplift

Farallon Plate
<table>
<thead>
<tr>
<th>#</th>
<th>GDE</th>
<th>Sketch log</th>
<th>Lithology</th>
<th>Description</th>
<th>Dimensions</th>
<th>Interpretation</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Alluvial-to-coastal-plain sandstones</td>
<td></td>
<td>sandstone (80-100%), mudstone (0-20%)</td>
<td>amalgamated, erosively based fine- to coarse-grained, fining-upward, cross-bedded and cross-laminated sandstones.</td>
<td>10-100 m thick</td>
<td>densely stacked fluvial channel-fill and sheet sandstones</td>
<td>Gill and Cobban (1966); Van de Graaf (1972); Van Wagoner (1995); Martinsen et al. (1999); Yoshia (2000); Hampson (2010); Aschoff and Steel (2011a); Leva López and Steel (2015)</td>
</tr>
<tr>
<td>2</td>
<td>Coastal-plain coals, mudstones and sandstones</td>
<td></td>
<td>sandstone (0-80%), mudstone (20-100%), coal (0-20%)</td>
<td>coals, rooted mudstones and thin, fine-grained sandstone sheets, with subordinate (and variably amalgamated) erosively based, fine-to coarse-grained, fining-upward, cross-bedded sandstones.</td>
<td>10s-100s m thick</td>
<td>fluvial channel-fill sandstones interspersed in fine-grained coastal-plain deposits</td>
<td>Yoshida (2000); Jinnah and Roberts (2011); Hampson et al. (2012); Lawton et al. (2014); Shiers et al. (2014); Flood and Hampson (2014, 2015)</td>
</tr>
<tr>
<td>3</td>
<td>Shoreline sandstones</td>
<td></td>
<td>sandstone (30-100%), mudstone (0-70%)</td>
<td>coarsening-upward, very fine- to medium-grained sandstones with interbedded mudstones. Variable internal facies characteristics and architectures (e.g. Fig. 4).</td>
<td>5-10 m thick</td>
<td>regressive and transgressive shoreline sandstones</td>
<td>see text for references for five shoreline-sandstone classes</td>
</tr>
<tr>
<td>4</td>
<td>Marine mudstones</td>
<td></td>
<td>sandstone (0-10%), mudstone (90-100%)</td>
<td>sparsely to intensely bioturbated mudstones, containing Zoophycos and Cruziana ichnofacies.</td>
<td>10s-1000s m thick</td>
<td>siliciclastic-rich offshore mudstones</td>
<td>MacQuaker et al. (2007); Hampson (2010)</td>
</tr>
<tr>
<td>5</td>
<td>Gravity-flow siltstones and sandstones</td>
<td></td>
<td>sandstone (0-50%), mudstone (50-100%)</td>
<td>very fine- to medium-grained, planar- and cross-laminated sandstones, with interbedded mudstones.</td>
<td>10 s m thick</td>
<td>river-fed turbidity currents and hypervenal flows, and wave-supported gravity flows.</td>
<td>Johnson (2003); Pattison (2005); Pattison et al. (2007); Hampson (2010)</td>
</tr>
<tr>
<td>6</td>
<td>Marine marls and chalk</td>
<td></td>
<td>calcareous mudstone (100%)</td>
<td>decimeter-scale interbedded marls and chalk, containing Zoophycos ichnofacies.</td>
<td>10s-100s m thick</td>
<td>siliciclastic-poor offshore mudstones</td>
<td>Savrda and Bottjer (1989); Da Gama et al. (2014)</td>
</tr>
</tbody>
</table>
Timeslice 1: Scaphites Hippocrepsis I
~83 Ma

Kamola & Van Wagoner (1995)
Olariu et al. (2010)
Hampson et al. (2011)

Seymour & Fielding (2013)
Roehler (1990)

Lawton et al. (2014)

alluvial-to-coastal-plain sandstones
coastal-plain coals, mudstones, and sandstones
shoreline sandstones
marine mudstones
gravity-flow siltstones and sandstones
marine marls and chalks

sediment dispersal pathway
progradationally stacked shoreline sandstones
retrogradationally stacked shoreline sandstones

length of embayment = ~225 km
Timeslice 2: Baculites Sp. (smooth) ~79 Ma

- Sediment dispersal pathway
- Progradationally stacked shoreline sandstones
- Retrogradationally stacked shoreline sandstones

- Alluvial-to-coastal-plain sandstones
- Coastal-plain coals, mudstones, and sandstones
- Shoreline sandstones
- Marine mudstones
- Marine marls and chalks
- Gravity-flow siltstones and sandstones

Length of embayment = ~270 km

Olsen et al. (1999)
Yoshida (2000)
Fitzsimmons & Johnson (2000)
Hampson et al. (2008)
Fitzsimmons & Johnson (2000)
Lawton et al. (2014)
Timeslice 3: Baculites Reduncus ~77 Ma

- Alluvial-to-coastal-plain sandstones
- Coastal-plain coals, mudstones, and sandstones
- Shoreline sandstones
- Marine mudstones
- Gravity-flow siltstones and sandstones
- Marine marls and chalks

Legler et al. (2014)
Van Cappelle et al. (2016)

Kiteley & Field (1984)

Hubert et al. (1972)

Gomez-Veroiza & Steel (2010)

Steel et al. (2012)

Pierre Shale
Mancos Shale
Lewis Shale

“Capping Sst.” of the Wahweap Fm.

length of embayment = ~450 km

Donselaar (1989), Jordan et al. (2016)
Timeslice 4:
Exiteloceras Jenneyi
~75 Ma

Masters (1967)
Kiteley & Field (1984)
Gomez-Veroiza & Steel (2010)

Kirschbaum & Hettinger (2004)
Aschoff & Steel (2011a)

length of embayment = ~250 km
Timeslice 5:  
**Baculites Compressus** 
~73 Ma

- **sediment dispersal pathway**
- **progradationally stacked** shoreline sandstones
- **retrogradationally stacked** shoreline sandstones

**Masters (1967)**  
**Kiteley & Field (1984)**  
**Gomez-Veroiza & Steel (2010)**

**Pictured Cliffs Sst.**

**length of embayment** = ~200 km

**alluvial-to-coastal-plain sandstones**  
**coastal-plain coals, mudstones, and sandstones**  
**shoreline sandstones**  
**marine mudstones**  
**gravity-flow siltstones and sandstones**  
**marine marls and chalks**
early Campanian
B. Asperiformis
-S. Leei III
79-84 Ma

late Campanian
B. Perplexus
-B.Eliasi
72-78 Ma
late Campanian

Wyoming salient

Mogollon Highlands

Bisbee Basin

early Campanian

pronounced embayment geometry of “Utah Bight” generates potential for tidal resonance

open embayment geometry of “Utah Bight” facilitates extensive wave reworking

middle Campanian

late Campanian

alluvial-to-coastal-plain sandstones

coastal-plain coals, mudstones, and sandstones

shoreline sandstones

marine mudstones, marls and chalks

active thin-skinned (Sevier) Sevier structure

active basement-involved (Laramide) structure

longshore drift

500 km