Facies architecture of the fluvial to tidal transition of mixed-influence deltas

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30 January 2017
Declaration of originality

I, Marijn van Cappelle, hereby declare that this thesis is the product of my own work at Imperial College London under the supervision of Dr. Gary J. Hampson and Prof. Howard D. Johnson. All other published and unpublished work used in this thesis has been appropriately referenced to.

Marijn van Cappelle

30 January 2017, London
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Abstract

Coastal deposits are often classified based on the depositional processes (wave, tide, fluvial) which operate in the depositional environment. The mix of processes operating during deposition have an effect on the size, shape, orientation and internal heterogeneities of sand- and sandstone bodies. It is important to know these properties for predicting and modelling fluid flow through aquifers. Facies models for fluvial-dominated deltas and wave-dominated shorefaces are well established and widely used. However, although there are many modern tide-influenced deltas, research in facies models for tide-influenced deltas lags behind to their better studied fluvial- and wave-dominated counterparts. This is partially because tide-dominated deposits are often interpreted in a sequence stratigraphic framework as transgressive tide-dominated estuaries. The aim of this study is to present a facies model for mixed-influence deltas, with emphasis on the preservation of channelised fluvial- and tidal channels in a progradational setting. Two case studies are part of this thesis. Firstly, outcrops of the Upper Cretaceous Sego Sandstone (Utah, USA) have been studied in order to investigate the facies architecture of a mixed tide- and wave influenced deltaic deposit. Secondly, cores of the mixed influence Lower to Middle Jurassic Ror, Tofte and Ile formation from the subsurface of the Halten Terrace (offshore mid-Norway) have been examined. Both case studies were deposited in structurally controlled embayments which favoured amplification of tides. Both case studies show an initial progradational phase with little evidence for tidal processes. In contrast, deposits on the delta plain show abundant evidence for tidal currents. In this study, it has been interpreted that these channelised tidal channels are part of continuous progradation of a tide-influenced delta. This is in contrast to sequence stratigraphic models in which tide-dominated deposits are interpreted as part of a transgressive phase. During progradation, these tidal and fluvial channels erode down in underlying deposits and they are deposited in progressively more basin-ward positions. More tide-dominated channels deposited in relatively distal or off-axis locations have a more planar geometry. More fluvial-dominated channels deposited in relatively proximal or on-axis location are more lenticular.
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Chapter 1. Introduction

1.1 Rationale

Clastic coastal depositional systems are often classified based on their geomorphology resulting from the interaction of nearshore depositional processes (fluvial currents, waves and tides), which transport, deposit and rework sediment (Galloway, 1975; Boyd et al., 1992; Orton and Reading, 1993; Ainsworth et al., 2011). The mix of processes operating during deposition have an effect on the size, shape, orientation and internal heterogeneities of sand- and sandstone bodies. It is important to know these properties for predicting and modelling fluid flow through aquifers (Ainsworth et al., 2011), for example in order to optimize production from hydrocarbon reservoirs. For modelling fluid flow through porous rock, it is important to know the net reservoir content over total rock volume (nett to gross, N/G) and the variations thereof, both vertically and horizontally. Also the shape (e.g. elongation direction), dimensions and orientation of sandstone bodies are important parameters for modelling fluid flow through rock formations.

Modern river deltas are readily classified according to the relative influence of these depositional processes (fluvial currents, waves and tides) based on their geomorphology and hydrodynamics of the receiving basin. However, ancient deposits are often interpreted as an end-member based on existing facies models, especially as wave- and fluvial-dominated river deltas (Elliott, 1978; Reading and Collinson, 1978; Bhattacharya and Walker, 1992). Ancient deltaic deposits are less frequently interpreted as the product of tide-dominated deltas (Martinius et al., 2001; Martinius et al., 2005; Pontén and Plink-Björklund, 2007; Pontén and Plink-Björklund, 2009; Tänavsuu-Milkeviciene and Plink-Björklund, 2009; Legler et al., 2013; Ichaso et al., 2016). In sequence stratigraphy, tide-dominated deposits are interpreted to be deposited in during transgressions (Van Wagoner et al., 1990; Van Wagoner, 1995; Catuneanu, 2006), despite there being many examples of modern tide-influenced regressive deltas. This common fixation on the sequence stratigraphic model is a reason that research on the preservation of tide-dominated regressive deltas has lagged behind to research on fluvial- and
wave-dominate deltas. Attempts have been made to construct facies-models for tide-dominated deltas (Willis, 2005; Dalrymple and Choi, 2007; Plink-Björklund, 2012). Dalrymple and Choi (2007) conceptualize how depositional processes and the texture of sediment vary spatially in tide-dominated deltas in analogy to well-known models for tide-dominated estuaries and some modern tide-dominated deltas. However, they do not explain the vertical relationship between deposits and how deposits stack vertically during progradation. Willis (2005) and Plink-Björklund (2012) review the facies which can be expected in tide-dominated deltas and give examples of them from ancient deposits. Despite that modern tide-dominated deltas formed under relatively steady sea-level rise since the last glacial, models for deltas which were tide-dominated during progradation, often have a significant sequence stratigraphic aspect to them. In these models, “highstand system tracts”, “sequence boundaries”, “lowstand system tracts” and “transgressive system tracts” play an important role in explaining all the elements (Willis, 2005).

This latter point of the need of sequence stratigraphic explanations feeds into the other reason why tide-dominated deltas are not very often interpreted in the geological record: Tide-dominated deposits are often attributed to transgressive estuaries as opposed to regressive river deltas (Van Wagoner, 1991; Boyd et al., 1992; Dalrymple et al., 1992; Van Wagoner, 1995; Dalrymple, 2006a). Nevertheless, there are many modern examples of tide-dominated and tide-influenced deltas (Allen and Chambers, 1998; Hori et al., 2001; Hori et al., 2002; Ta et al., 2002; Dalrymple et al., 2003; Salahuddin and Lambiase, 2013). The misinterpretation of a sequence boundary can have large implications for predictions made on the basis of the sequence stratigraphy model. For example, in sequence stratigraphy, the formation of a sequence boundary is contributed to a decrease in relative sea level. In initial models this was contributed to eustatic sea level changes (Van Wagoner et al., 1990), however, also other allogenic forcing mechanisms such as basin subsidence and sediment supply are considered (Catuneanu, 2006). The implication of interpreting these allogenic forcing mechanisms is that they operate on a regional scale, and that the same sequence boundary and associated change in facies can be found elsewhere in the basin, or even around the world. Also, the sequence stratigraphic
model predicts that during the formation of a sequence boundary, sediment is by-passed to be deposited in on deep marine fans down-dip from the coastline. However, when sequence boundaries and associated incised valley fills are re-interpreted as part of a tide-influenced delta, these predictions for erosion and deposition elsewhere in the basin do not hold.

In fluvial-dominated deltas, deposition is concentrated at a point in the river mouth. Here a transition takes place from confined fluvial currents in the fluvial channel to unconfined flow in the open sea. Due to the expansion of the flow, current velocities sharply drop, and sediment is deposited at the river mouth on mouth bars (Wright, 1977). In wave-dominated settings, sediment is transported by wave-generated currents. When waves reach the shallow water-depths under an angle, a longshore current is created. Due to these longshore currents, sediment is transported and deposited in a shore-parallel fashion (Wright, 1977). However, in tide-influenced settings, shore-normal sediment transport via tidal channels is common, and sediment is eroded, reworked and re-deposited in channels and bars with an orientation perpendicular to the shoreline. The depositional model for tide-dominated, transgressive estuaries (Dalrymple et al., 1992) is widely accepted and applied. Theoretically, a similar transition from fluvial-dominated distributary channel at the apex of the delta, to marine processes (tides and waves) in distal location should apply to tide-influenced deltas (Dalrymple and Choi, 2007). However, it remains unclear how this would be preserved in the geological record.

The main aim of this thesis is to present a facies model for regressive tide-influenced deltaic deposits. Of particular interest is the preservation of the fluvial-to tidal transition, which is widely recognized in tide-dominated transgressive estuaries (Dalrymple et al., 1992).

1.2 Outline

In chapter 2, a literature review of existing knowledge on the sedimentology of regressive deltas and tide-influenced siliciclastic systems is given. This gives background against which a facies model for the fluvial-to tidal transition of regressive deltaic deposits will be developed.
In chapter 3, the western outcrops of the lower Sego Sandstone in the Book Cliffs (Campanian of the Western Interior Seaway, North America) will be described. This unit is an overall regressive system which shows abundant proof for tidal processes in combination with wave processes (Van Wagoner, 1991; Willis and Gabel, 2001; Willis and Gabel, 2003; Wood, 2004; Legler et al., 2014; Burton et al., 2016). Although the lower Sego Sandstone has been studied in detail in its eastern and central outcrops of the Book Cliffs by previous workers, the western outcrops have previously only been described on a large scale. This western part of the Sego Sandstone outcrop have been examined for evidence of the fluvial-to-tidal transition in proximal deposits of the tide-influenced “lower Sego Delta”. The benefits of these outcrops are that detailed facies analysis is possible, that the geometry of the facies can be determined and that the facies can be traced out laterally and vertically, in order to get a complete view of the facies architecture. A modified version of chapter 3 has been published in Sedimentology (2016) with Stephen Stukins, Gary Hampson and Howard Johnson as co-authors. Stephen Stukins performed the palynological analysis (<~10% of the data) in this chapter, while Gary Hampson and Howard Johnson discussed and commented on the technical content of the chapter.

In chapter 4, the sedimentology and facies of the mixed influenced regressive-to-transgressive clastic wedge of the Ile Formation (Early to Middle Jurassic, Halten Terrace, offshore mid-Norway) will be explored using core data. The deposits of the Ile Formation form one of the reservoir units from which oil and gas is produced. A better understanding of these deposits will have economic implications for the exploration, development and production strategy of these oil and gas fields. The cores of this succession are very well recovered, show a mix of depositional processes (Harris, 1989; McIlroy, 2004; Ravnås et al., 2014) and enable detailed observations. Correlation of the wireline logs of the cored wells gives a perspective on the vertical and lateral stacking of the deposits, and how the depositional system and processes varied spatially and temporally over the southern Halten Terrace. A major challenge in subsurface studies is the uncertainty in correlation and the absence of data between wells. Outcrop studies with more detail in lateral variability of the deposits (such as chapter 3) can be used as analogues to reduce uncertainty. A modified version of this chapter has been submitted to Marine
and Petroleum Geology and is under review. Rodmar Ravnås, Gary Hampson and Howard Johnson are co-authors who have discussed and commented on the technical content of the chapter.

In chapter 5, the spatial and temporal variability in depositional process along the southwestern coastline of the Western Interior Seaway (Campanian of North America), including the Sego Sandstone, is explored by synthesising the extensive published literature for these well-exposed coastal deposits. The presence of tide-influenced and tide-dominated progradational depositional environments in an elsewhere mostly wave-dominated basin is anomalous and it has long been questioned what is the cause for the increase in tide-influence (Aschoff and Steel, 2011b; Steel et al., 2012). This synthesis gives insight into the relationship between active tectonics and regressive tide-influenced deltaic deposits. A modified version of this chapter will be submitted to the Journal of Sedimentary Research. Gary Hampson and Howard Johnson are co-authors who have discussed and commented on the technical content of the chapter.

The work presented in chapters 3, 4 and 5 is synthesized in chapter 6, where the deposits of the Sego Sandstone and the Ile Formation are compared and contrasted. The common features of these two deposits are distilled in order to present a new facies model for tide-influenced regressive deltaic deposits, which is compared to other examples from literature. Also, the regional, basin-scale settings of tidal influence in the Campanian of the Western Interior Seaway (Chapter 5) and the Early-to-Middle Jurassic Halten Terrace (Chapter 4) are compared and contrasted in order to predict where to find tide-influenced deltaic deposits.
Chapter 2. Literature review

The aim of this literature study is to review tidal forces, its effect on currents in the open ocean and shallow-marine coastlines where tidal currents erode, redistribute and deposit sediment. The focus lays on how tidal currents affect the deposition of sediment in the shallow marine realm and how the sediment preserves in the geological record. Firstly, the origin of tidal forces will be briefly discussed. Also, the effect of tides on the transport and deposition of sediment will be reviewed. Then classification schemes for shallow marine coastal deposits will be discussed. Then modern and ancient examples and facies models for tide-influenced depositional environments will be shown. Special focus will be lain on deltaic deposits Also wave- and fluvial-dominated progradational deltaic systems will be shown for comparison.

2.1 Tidal force

The definition of ‘tide’ refers to the diurnal (daily) or semidiurnal (twice daily) cycle of rise and fall of sea level as observed along coastlines. This is driven by the unequal distribution of the gravitational force of a second celestial body over the surface of the first celestial body. This unequal distribution of the gravitational force deforms a solid celestial body, but when large amounts of fluid are present covering vast areas of a celestial body, the tidal force also deforms this fluid mass, for examples to generate tides in Earth’s oceans (Stewart, 2006).

2.1.1 Periods

There is a range of periods associated with the tidal force. The most important are the semidiurnal, diurnal and the neap-spring cycle. The semidiurnal and diurnal periods are explained by the tidal force of the Sun and Moon on Earth. The neap-spring cycle is the interplay between both tidal forces. These are the major periods of the tidal force that can be readily observed on the coast. However, many astronomical parameters can change the gravitational forces between Sun, Moon and the Earth such as the eccentricity of the orbits of the Earth (100 kyr period) and Moon (no fixed period, e.g. Garrick-Bethell et al., 2006; Ćuk, 2007; Iorio, 2011), and the precession of the orbits of the Earth (110 kyr) and
Moon (19 yr). In modern systems, the tides are predicted using a numerical model that contains periods between 12 hours and 19 years.

2.1.2 Distribution over the earth

A tidal wave cannot maintain the propagation velocity necessary to keep up with the angular velocity of the sun and moon because the oceans are too shallow to sustain such velocities and moreover, continents are present on Earth and block the motion of any wave propagating all the way around the Earth. Therefore the tidal wave lags the tidal force and because of the retardation the tidal wave will be deflected by the Coriolis force. In oceanic basins amphidromic points or tidal nodes are present around which tidal waves rotate (Figure 2.1A). On wide continental shelves and rugged coastlines, retardation and amplification of the tidal wave can produce complicated rotation about amphidromic points. Propagation of the tidal wave through the North Sea is a well-known example (Figure 2.1B) (Houbolt, 1968). The tidal range at amphidromic points is 0 and increases away from amphidromic points. When the wave-length of standing tidal wave is in resonance with the distance of the amphidromic point and the coastline, amplification takes place (Allen, 1997). Examples can be seen in Figure 2.1A.

![Figure 2.1](image.png)

Figure 2.1. (A) M2 twice daily lunar constituent of the tide from the FES2004 mode (Lyard et al., 2006). (B) Amphidromic points in the North Sea (Houbolt, 1968).
2.1.3 Tidal amplification and tides in sedimentary basin

Along coastlines, tides can be amplified by several mechanisms: Shallowing of the bathymetry, funneling and tidal resonance. Tidal amplification by shallowing and funneling is basically the same, but in a different dimension, funneling is the restriction of the tidal current in the horizontal plane, and shallowing of the bathymetry is a restriction of the tidal current in the vertical plane. Because a constant volume of water is forced through an increasingly restricted space, the flow velocity of the tides must increase, and the water is forced upward to create an increased tidal range (Pugh, 1987). Another mechanism for tidal amplification is tidal resonance. Tidal amplification takes place when the length and depth of an embayment are in resonance with the tidal wave (Allen, 1997). When the embayment is also wide enough, an amphidromic point can occupy the center of the embayment (Pugh, 1987; Sztanó and De Boer, 1995), and the tidal wave with an amplified amplitude will travel along the coastline of the embayment around the amphidromic point. Tides can travel along shallow shelves and epicontinental seaways, where tides can be locally amplified by both funneling, shallowing and tidal resonance, providing the distance along the shallow seaway is not too long, so the water from the open ocean can flow into the seaway (Erickson and Slingerland, 1990; Sztanó and De Boer, 1995; Wells et al., 2005a; Wells et al., 2005b; Wells et al., 2010a; Wells et al., 2010b).

2.2 Tidal bedforms

In tide-dominated environments, sand from the bed-load is transported and deposited by ebb- and flood tidal currents. During slack water, current velocities stagnate and clay from the suspension load settles out to form a mud drape of the bedding surface. The latter process is not merely the results of stagnating currents velocities, but is also aided by clay flocculation due to salt particles in areas where fresh and salt water mix (Dalrymple and Choi, 2007; Ichaso and Dalrymple, 2009; Guo and He, 2011). Usually the resulting heterolithic nature of deposits is attributed to reversing and stagnating tidal currents (Reineck and Wunderlich, 1968). However care must be taken to classify heterolithic deposits to be the result of tidal currents, because similar heterolithic deposits can also originate from steady
flows (Baas et al., 2015). Therefore other pieces of evidence for tidal currents must be obtained before interpreting tidal processes.

Along coastlines which experience semi-diurnal tides, the daily inequality in tidal range and therefore flow velocity between the two tides causes one of the sand beds between mud-drapes to be thinner than the other sand-bed (Johnson et al., 1982). Therefore the presence of so called “double mud-drapes” of “paired mud-drapes” is evidence for semi-diurnal tidal currents. Such structures do not form along coastlines experiences diurnal tides, because there is only one tidal cycle per day, so no daily inequality occurs. Another tidal cycle which can be preserved in tidal deposits is the spring-neap cycle. This cycle results from the monthly rotation of the moon about the sun, and the two constructive and de-constructive interference between the gravity of the Sun and Moon on the Earth’s oceans (Kvale, 2012). The result of this is that in a month, the tidal range increases and decreases twice every subsequent tidal cycle. This results in increasing and decreasing amounts of sands transported and deposited between slack water mud-drapes (Johnson et al., 1982). This can be recorded in relatively quiet and stable environments such as bay fill deposits (Kvale et al., 1989; Kvale, 2006), but also in dune scale cross-bedding (Allen, 1981a; Allen, 1981b).

Dune scale cross-bedding formed under reversing tidal currents can be distinguished from cross-bedding formed under steady flows (Figure 2.2-1.3). Firstly tidal cross bedding can show evidence for reversing currents by way of reactivation surfaces. Usually, either the flood- or ebb-tide is the dominant current in a channel due to the presence of mutual evasive channels and because ebb- and flood-tide occupy different elevation levels in a channel (Boersma, 1969; Van Veen et al., 2005). The result is that any dune is mainly constructed by either flood- or ebb-tidal currents. However, currents can stagnate during slack-water, and weak-reversal of flow can take place. This will results in some erosion of the lee-side of the dune. After reactivation by the tide which constructs the dune, the deposition takes place again on the lee-side of the dune. The result of these processes is that a reactivation surface is preserved which is indicative for tides (Boersma, 1969; Allen, 1980).
The deposition of mud-drapes (as discussed above) and the preservation pattern thereof is also typical evidence for tidal currents: Due to higher currents velocities at the crest of dunes, mud-drapes at- or close to the crest of dunes are liable to erosion when the tides start to make. However, at the foreset and especially the toesets in troughs between dunes, the mud drapes are protected against erosion. Therefore there is a higher preservation potential for mud-drapes in the toeset of tidal dunes. Additionally, in a similar fashion, current-ripple cross-lamination formed by back flow and/or reversed tidal currents have a higher preservation potential to form in the troughs of dunes, resulting in heterolithic, current-ripple cross-laminated toesets of dune (Oomkens and Terwindt, 1960; Boersma, 1969; de Raaf and Boersma, 1971; Allen, 1980; Van den Berg et al., 2007).

Figure 2.2.Model for tidal cross-stratification. The figure show the locations of mud-drapes (black, marked by “M”) in the upper figure and reactivation surfaces (solid lines in the lower figure) (after Allen, 1980).
Figure 2.3. Dune scale tidal cross bedding from the Holocene Rhine-Meuse Delta, the Netherlands. Note the elongated, asymptotic toesets and cross bedding in opposite direction at the top of the pictures which are both typical for tidal cross beds (Van den Berg et al., 2007).

Figure 2.4. Dunes with a flood directed (right) palaeocurrent directions with a relative fine sand grain size and high proportion of mud and ebb/fluvial downstream (left) directed cross beds with a relative coarse sands grain size and low proportion of mud. Miocene of the Lower Rhine Embayment, Germany (Van den Berg et al., 2007).
2.3 Classification of shallow marine coastal deposits

Coastal depositional systems are usually classified on the basis of the mix of depositional processes (waves, tides, rivers) which operate in the coastal depositional environment. Galloway (1975) was the first to propose a classification scheme which applied to river deltas (Figure 2.5). In this classification scheme, the sub-aerial geomorphology and the hydrodynamics of modern systems are used to define the operating depositional processes. The method enables the classification of a mix of two or three of the depositional processes.

Boyd et al. (1992) expanded this classification scheme beyond just river deltas to other coastal depositional environments which are recognized along modern shorelines (tidal- and wave-dominated estuaries, tidal flats, wave-dominated strand plains/shorefaces, and beach-barrier-back-barrier lagoon systems). They also showed the importance of recognizing weather a coastline is transgressive or regressive (Figure 2.6). Along regressive coastlines tidal flats (tidal-dominated), river deltas (fluvial-dominated) and strand plains/shorefaces (wave-dominated) develop depending on the dominant depositional process. When these depositional environments are transgressed, tidal flats, estuaries and barrier-island-back-barrier-lagoon systems develop respectively. These environments see a decrease in fluvial-influence and an increase in wave- and more notably tide-influence.

More recently, Ainsworth et al. (2011) proposed a modified version of this classification scheme (Figure 2.7). The rationale behind this scheme is that any coastal system can be objectively classified on the basis of the proportions of depositional processes interpreted from observations from facies analysis.
Figure 2.5. Depositional process classification of deltas (after Galloway, 1975).
Figure 2.6. Depositional process classification of shallow marine coastal deposits. Left: Shallow marine depositional environments along progradational and transgressive coastlines. Right: Classification of these depositional environments based on the depositional process (waves, rivers, tides) and direction of shoreline migration (after Boyd et al., 1992).

Figure 2.7. Depositional process classification of shallow marine coastal deposits (insets) and examples of morphologies of shorelines with a mix of fluvial, wave and tidal processes (after Ainsworth et al., 2011).
2.4 Tide-influenced depositional environments

A wide variety of tidal depositional environments can be found in the present, from longitudinal bar systems on the continental shelf to tide-influence in rivers far up-river from a coastline. In this literature review, different Holocene environments will be discussed together with ancient deposits which have been interpreted to have been deposited in an equivalent environment in order to understand sedimentological facies and facies architectures of these environments. The term ‘tidal bar’ is very loosely used in literature to refer to all kind of different tide-influenced bars, e.g. coarsening upward successions above marine mudstones and fining upward channel-fill deposits (both side-attached points bars and side-detached longitudinal mid-channel bars) overlaying an erosional surface. In my view, ‘tidal bar’ most frequently refers to coarsening upward successions above marine mudstones (i.e. tide-dominated distributary mouth bars). However, in order to prevent confusions, the words ‘tidal bar’ are avoided in this thesis. Instead, an addition is added to clarify the type of ‘tidal bar’, e.g. ‘coarsening upward tidal/tide-dominated bar’ for coarsening upward tidal bars above marine mudstones (i.e. tide-dominated distributary mouth bar) and ‘longitudinal tidal/tide-dominated bar’ for mid-channel, side-detached, longitudinal tide-dominated bars deposited as channel-fill above an erosion surface. Tide-dominated point bars are referred to as such, or when preserved in the geological record as inclined heterolithic strata (IHS).

2.4.1 Longitudinal bars on the open shelf

On the open continental shelf, where tidal currents are strong enough to reach the shallow seafloor, tidal currents can rework sediment that is available on the sea floor, forming laterally migrating tidal ridges. Examples of these tidal ridges can be found in the southern North Sea (Houbolt, 1968; Berné et al., 1994; Trentesaux et al., 1999), Irish Sea (Scourse et al., 2009) and East China Sea (Yang, 1989; Berné et al., 2002; Hori et al., 2002; Yoo et al., 2016). Sometimes large dunes that migrate down-current can form in a similar way on the continental shelf (Ikehara and Kinoshita, 1994). Sedimentation on the continental shelf is usually characterized by deposition of clay sized particles that are not readily re-suspended. However during and after a transgression, relict lowstand shoreline sands can be reworked
to form tidal ridges on the shelf (Houbolt, 1968; Scourse et al., 2009). Tidal ridges are long but narrow sand bodies that form roughly parallel to the main tidal current directions. The ridges are asymmetrical with a shallowly dipping stoss side where net erosion takes place and a steep lee side where net deposition takes place (Figure 2.8). Superimposed on the ridge, dunes migrate obliquely (~45°) to net-tidal current direction and the dip direction of the tidal ridge. The dominant tidal currents on either side of the ridge can be mutually evasive, so that flood tidal currents concentrate on one side of the ridge while the ebb tidal currents dominate flow on the other side (Houbolt, 1968). These tidal ridges are often rich in sand-content, and mud is rare, resulting in clean dune-scale cross-bedding. The surface over which the ridge migrates (obliquely to tidal currents) is an erosional surface that might be lined by a lag (Houbolt, 1968).

Tidal ridges are often quoted as an analogue for ancient depositional systems which show evidence for tidal currents and are encased in offshore sediment (Anderton, 1976; Suter and Clifton, 1999; Olariu et al., 2012). However, often these ancient examples do not closely resemble their modern analogues in facies, geometry and size. Olariu et al. (2012) however described a deposit in detail with respect to sediment textures, palaeocurrent directions, inclined surfaces and geologic context which does closely resemble modern tidal ridges Figure 2.9.
Figure 2.8. Model of the architecture of a tidal ridge or longitudinal tidal bar based on observations of Holocene deposits in the southern North Sea (after Houbolt, 1968).

Figure 2.9. Simplified cross section of a lateral migrating longitudinal bar on a shelf (Esdolomada Member, Eocene, Tremp-Graus Basin) (after Olariu et al., 2012).
2.4.2 Barrier island- back barrier lagoon systems

The first research in tidal sedimentology was performed in the back-barrier tidal flats of the Wadden Sea (Van Straaten and Kuenen, 1957; Reineck, 1958; Van Straaten, 1961; Reineck, 1967). These deposits form landward of barrier islands that protect the lagoons behind them. Where no tides and large storm surges, but strong longshore currents are present, tens to hundreds of kilometre long spits form. When storm surges and tides occur, especially the combination of storm surges and spring tides, the spit can be breached and a permanent inlet to the lagoon can be formed, transgressing the coastline further landward. This creates a series of barrier islands separated from each other by tidal inlets. Where tides are present, ebb and flood tidal currents will alternate through this inlet channel, creating an ebb tidal delta on the seaward side of the inlet channel and a flood tidal delta on the lagoonal side. The seaward side of the barrier is prone to high wave action from the open sea, hence it is wave-dominated. On the other hand, the ebb and flood tidal deltas form where tidal currents are concentrated and contain many tidal structures (flaser, wavy and lenticular bedding and bidirectional paleocurrents) in a thin (metre-scale) coarsening upward deltaic succession. The lagoon can be muddy, sandy or mixed, depending on the flow velocities and the availability of sediment (Van Straaten, 1961; FitzGerald et al., 2012).

Tide-influenced barrier-back barrier systems have also been interpreted in the geological record. These type of deposits form during transgressions (Ainsworth et al., 2011) and contain a wave-dominated shoreface succession and adjacent back-barrier deposits which can be sandy and/or muddy. Metre thick coarsening upward deposits that are interpreted as flood tidal deltas are often preserved where barrier and back-barrier deposits border each other. However, ebb-tidal deltas, foreshore and dune deposits of the barrier island are generally poorly preserved, because barrier-back-barrier systems often from during transgressions, during which these deposits are prone to erosion (Kamola and Van Wagoner, 1995; Sixsmith et al., 2008).
2.4.3 Tidal flats

Tidal flats are found along meso- to macrotidal shorelines under both transgressive and regressive conditions. They can be found at shorelines without major fluvial input, either with or without the protection of barrier islands, and they can also be found at the margins of tide-dominated deltas and estuaries. Tidal flats are vertically restricted by the tidal range and horizontally by the slope of the tidal flat and the tidal range. Tidal flats can be subdivided in two subcategories: Sandy tidal flats with only shrub vegetation on supratidal salt marshes in high latitudes and mangroves in the tropics.

2.4.3.1 High latitude tidal flats and salt marshes

In high latitudes, the lower parts of the tidal flat are sand flats, grading landward into more clay rich mixed flats and eventually up into mudflats. The sand flat is characterized by ripple lamination with flasers. The mixed flat is characterized by flaser, wavy and lenticular bedding, and the mud flat by mud with lenticular bedding. Bedforms in each of the three sub-environments are sand ripples transported by flood and ebb tidal currents with more or less mud deposited during slack water (Häntzschel, 1955; Van Straaten, 1961; Reineck, 1967; Kellerhals and Murray, 1969; Choi and Dalrymple, 2004; Choi, 2010). Sand grain size decreases and mud content increases away from the main tidal channels and toward the supratidal area (Van Straaten and Kuenen, 1957). The mudflats are generally quite dark due to high pyrite content (Häntzschel, 1955). Supratidal salt marshes are vegetated areas that are only flooded infrequently during storm surges, but not by normal tidal currents. These salt marshes are dissected by muddy meandering channels that exhibit low rates of lateral accretion, but result eventually in point bar deposits deposited on the banks of the inner bend of the meandering channels (Häntzschel, 1955; Van Straaten and Kuenen, 1957; Fenies and Faugères, 1998; Fagherazzi et al., 2004). Lower, in the intertidal zone where the substrate is more sandy, shells debris can accumulate by combined storm-generated and tidal currents in the form of cheniers (Van Straaten and Kuenen, 1957; Weill et al., 2012).
In the ancient, tidal flats are often interpreted as thin intervals between the top of channel-fills and coal deposits (Legler et al., 2013; Ainsworth et al., 2015), although also thicker intervals of sediment have been interpreted as tidal flats (Eide et al., 2016).

2.4.3.2 Tropical mangroves

Along coastlines in the tropics, salt-tolerant flora (mangroves) grow in the intertidal area (Staub and Gastaldo, 2003). The roots of the mangrove trees capture sediment. Therefore coarse sediment (sand) is usually deposited in- and close to channels. Further away from channels within mangrove forests, current velocities are lower and only mud and in-situ produced organic carbon is deposited.

Because salt tolerant flora only evolved in the Cenozoic (Greb et al., 2006), mangrove deposits can be expected to be common in the Cenozoic, but not in older deposits. Although many modern coastlines are tide- and mud-dominated (Nyberg and Howell, 2016), such deposits are not frequently interpreted in the ancient (Fan, 2012; Meor et al., 2013).

2.4.4 Estuaries

Estuaries form along transgressed shorelines, both on unconfined lower delta plain settings, but estuaries also form in settings where the position is confined in transgressed river valleys. Depending on the dominant process at the shorelines, sediment from seaward positions is transported into the estuary by tidal currents which operate shore-normal (tide-dominated estuary), or when wave-processes are dominant, a spit forms across the mouth of the estuary (wave-dominated estuary) (Dalrymple et al., 1992; Dalrymple and Choi, 2007).

2.4.4.1 Tide-dominated estuaries

Estuaries are extremely widespread on transgressive Holocene coastlines. Estuaries are indentations of coastlines, often with a funnel shape, that are filled with brackish water because rivers discharge water into their inner parts. They often have a relatively large tidal range that is amplified by the funnel shape of the estuary. From a sedimentological perspective an estuary can be divided into three domains. First there is the outer part where marine processes (waves and tides) dominate. Secondly
there is a mixed zone, and thirdly an inner, fluvial dominated zone where there is still some marine influence. Proximally and distally the estuary is bounded by areas where respectively fluvial and fully marine processes are the only factor.

In a tide-dominated estuary, the tidal currents are strong enough to rework any sediment that is brought into the mouth of the estuary by longshore currents. Tidal currents decrease and increase in a daily rhythm during which the tidal energy travels forth and back through the estuary, mixing and reworking sediment throughout the estuary (Figure 2.10, Figure 2.11).

Rivers that discharge in estuaries often have a low sediment yield. Because flood tidal currents are stronger than ebb tidal currents (Johnson et al., 1982) there is net sediment transport from the sea into the estuary (Green and MacDonald, 2001). Where the landward transport of the flood tidal currents and the basinward transport of fluvial processes meet, net convergence of sediment transport takes place. This is the location in the estuary with the lowest energy, and hence where sedimentation of the finest (clay) sediment takes place. In estuaries this is often in the apex of the funnel shape, where the change from channelized flow to non-channelized flow takes place (Dalrymple et al., 1992; Dalrymple and Choi, 2007). Slightly seaward from this point of net sediment convergence, tidal currents are amplified due to funneling, reaching a maximum at the location called the tidal maximum. Both landward and basinward of the tidal maximum, the tidal range and energy of the tidal currents diminishes.

2.4.4.1.1 Outer estuarine sand bars

Around the point where the tidal currents are at a maximum in the outer estuary, flow is non-channelized and efficient sediment transport builds up inter-channel, longitudinal, fining upward sandy bars with a similar architecture and processes to tidal ridges on the shelf. Bed forms on these bars are usually sandy, with some mud-drapes along the foresets (Reineck, 1958; Oomkens and Terwindt, 1960; Boersma, 1969; de Raaf and Boersma, 1971; Nio et al., 1980; Clifton, 1983; Harris et al., 1992; Harris, 1988; Fenies and Tastet, 1998; Masselink et al., 2009). Ebb and flood tidal currents
often concentrate on different sides of the longitudinal bars to create mutually evasive flow (Harris, 1988), or occupy different depths (Boersma, 1969).

2.4.4.1.2 Inner estuary inclined heterolithic strata

Landward of the elongated intra-channel bars of the outer estuary, fluvial currents become stronger, but tides remain the dominant depositional process. The result are high sinuosity channels in the inner estuary (Dalrymple et al., 1992). Because of the meandering channel pattern, side-attached point bars inclined normal to the currents and characterized by regular, repeated changes in grain size due to tidal current fluctuations are deposited resulting in inclined heterolithic strata (IHS) (Thomas et al., 1987; Choi and Dalrymple, 2004; Fagherazzi et al., 2004; Choi, 2010). In such point bars, the combined ebb tidal and fluvial currents follow the course of the normal, continuous thalweg in the outer bend of the meander. However the flood tidal currents tend to flow over the inner bend of the associated point bar, creating a shallow channel with flood directed structures (Fenies and Tastet, 1998). These flood dominated channels are often disconnected from each other (Dalrymple and Choi, 2007). These side-attached point bars of the inner estuary become gradually detached longitudinal bars in the outer estuary downstream, and the point bars become gradually more fluvial-influenced toward the fluvial-to-tidal transition in more proximal locations.

2.4.4.1.3 Fluvial-to-tidal transition

Landward of this meandering zone, channels are straighter and tidal currents diminish until only fluvial processes occur (Dalrymple and Choi, 2007). In some parts of the inner estuary there is still a (twice) daily rise and fall of water level, but further upstream there are no flood directed currents, but the ebb directed currents increase and decrease with the tidal rhythm. Only further upstream is there no influence of the tides. The points where these transitions occur can vary over time. During spring tides tidal influence is felt further upstream, in contrast with neap tides when tidal modulation becomes less important. Also changes in river discharge (for example due to storms and seasonal variations) change how far downriver the fluvial flow is dominant over tidal currents (Dalrymple and Choi, 2007).
In the fluvial-tidal transition zone, deposition takes place on side-attached point bars of a low-sinuosity meandering channel. During periods of high river discharge, relative coarse grained sand is deposited, while during low river discharge, tidal currents are more effective and transport fine sand in a landward direction. Also during low river discharge, currents slacken during slack tide, and mud is deposited, resulting in heterolithic deposits. Deposition usually takes place as dunes in at the base of the channel, and current ripples higher up the sides of the channel (Dalrymple et al., 2015a; Dalrymple et al., 2015b; Gugliotta et al., 2016).

2.4.4.1.4 Tide-dominated estuaries in the geological record

In the geological record, tidal deposits are mainly documented from transgressive deposits in incised valleys based on sequence stratigraphic models (Figure 2.18) (Van Wagoner et al., 1990; Van Wagoner, 1995; Dalrymple et al., 2006). In these models a valley has been incised within older deposits during a drop in relative sea-level. During subsequent transgression, the valley is transgressed and due to the funnel-shape of v-shaped valleys which are being flooded, tides are amplified. Because the estuaries form in incised valleys which are being flooded, base level gradually rises, and it can be expected that channel-fill deposits stack in a multi-story fashion during this base-level rise.

In other cases, tide-influenced channel fills are encased in lower delta plain deposits and for single-story architectures. These deposits usually correlate to otherwise wave-dominated shoreline deposits (Lanier et al., 1993; Plink-Björklund, 2005; Hassan et al., 2012; Pontén and Plink-Björklund, 2009; Gomez-Veroiza and Steel, 2010; Shiers et al., 2014). The presence of some of these deposits is has been explained by changes in relative sea-level, but the localized presence of estuaries can also be due to the localized presence of rivers entering the sea and switching of sediment input to the shoreline.
Figure 2.10. Model for deposition in a tide-dominated estuary (after Dalrymple et al., 1992).
Wave-dominated estuaries form when longshore currents are strong enough to produce a sand spit across the mouth of the estuary and tidal- and fluvial currents are not strong enough to rework this spit. Because fresh water is transported into the estuary by a river, usually a narrow channel is present at the down-drift end of the estuary where the river water is discharged into the sea. When tides are present (but weak), flood tidal deltas form landward from this inlet. In the central part that is protected from wave action by the sand spit, there is a muddy central basin with, at its landward limit a small
bayhead delta where the river discharges into the estuary (Figure 2.12) (Dalrymple et al., 1992; Cooper, 2001; Anthony et al., 2002).

In the geological record, wave-dominated deltas are less often interpreted (Hubbard et al., 2002; Milli et al., 2013). This might be partially due to the similarity in facies with barrier island-back barrier lagoon systems with a muddy lagoon. In both environments a wave-dominated shoreface and a muddy lagoon are the dominant facies. Elements such as tidal inlets, flood tidal deltas and bayhead deltas which cover a smaller area and are therefore less detectable are not unique to either environment neither. The big difference between the environments is the geometry of the geomorphology: Wave-dominated
estuaries are oriented shore-normal, while lagoons are shore-parallel. This needs careful and detail mapping in 3D, which is often not possible in the geological record.

Figure 2.12. Model for deposition in a wave-dominated estuary (after Dalrymple et al., 1992).
2.4.5 Deltas

2.4.5.1 Fluvial- and wave-dominated deltas

In fluvial-dominated river deltas, fluvial currents are confined to distributary channels until the fluvial current become unconfined at the location of the river mouth. Due to the expansion of the flow at this point, flow velocities decreases (Wright, 1977). Because of this decrease in flow velocity, the coarsest sediment transported by the river is deposited at the river mouth and progressively finer sediment is deposited further away from the river mouths. Due to this localized deposition at the river mouth, a mouth bar builds up which become eventually emerges above sea level. At this point in time the river bifurcates in two distributary channels and the river mouth processes migrate to the new river mouth locations of the two new distributary channels (Edmonds and Slingerland, 2007). When the system progrades in this fashion, the river gradient can become very low due to the lengthening of the system. When during a flooding a levee in a proximal location breaches, a new route to the basin with a steeper gradient can be created. Due to higher flow velocities it can become hydrodynamic more favorable to for the river to use this breach to enter the basin. In this way river switches and creates a new delta lobe (Coleman, 1988; Roberts, 1998).

This results in relatively mud-rich facies successions because mud from the suspension load of the river is deposited in front of the mouth bar, and no process for re-suspending the mud is present (Figure 2.13). When the delta progrades, these prodelta mudstones coarsen upward into planar laminated and current-ripple cross-laminated sand beds which are deposited on the delta front from hyperpicnal flows originating in the river mouth (Bhattacharya and MacEachern, 2009). These interbedded mudstones and sandstones coarsen upward into mouth bar deposits where the coarsest sediment from the bed-load of the distributary river are deposited on dunes (Bhattacharya and Walker, 1992). Often the mouth bar deposits sharply overlie the delta front deposits (Overeem et al., 2003; Fielding et al., 2005a).
When storm-waves and wave-generated longshore currents are present, mud will be kept in suspension and transported further offshore or alongshore (Aslan et al., 2003). Also sand will be transported from the river mouth along the shore line, resulting in elongated sand bodies parallel to the coastline (Correggiari et al., 2005; Fanget et al., 2014; Olariu, 2014). When sediment supply is large enough the beach ridge or spits prograde together with the river mouth bar and wave-dominated shoreface successions will be present lateral to fluvial-dominated mouth bar successions (Hampson et al., 2011). Typical wave-dominated shoreface successions (Bhattacharya and Walker, 1992) (Figure 2.13). The sedimentology of wave-dominated delta deposits is similar to wave-dominated linear shorelines. Offshore mudstones deposited in low energy environments below storm wave-base are overlain by interbedded mudstones and hummocky cross-stratified sandstones deposited between storm- and fair-weather wave-base where sand is only deposited during storm-event. Above fair-weather wave-base, longshore currents transport continuously forming cross-bedded intervals overlain by plain laminated facies deposited in the surf-zone (Van Wagoner et al., 1990). They can be distinguished from each other using trace fossil content based on the fact that deltas prograde rapidly so bioturbation intensity is lower and deltas have reduced salinities due to the input of fresh water from the river causing a lower trace fossil diversity (MacEachern and Bann, 2008). Also the planform geometry is different. Wave-dominated linear shorelines are straighter while wave-dominated deltas can be expected to have a more cuspate geometry. Even in the most-cited example of a fluvial-dominated delta, the Mississippi delta, wave-reworking of abandoned delta lobes results in a reduced coastline rugosity (Penland et al., 1985; Penland et al., 1988).
2.4.5.2 **Tide-dominated deltas**

Some of the largest river deltas on Earth are tide-dominated, including the Amazon, Ganges-Brahmaputra, Yangtze (Changjiang) and Fly River deltas. In plan view, these rivers tend to have only a few distributaries that widen as a funnel shape towards an indentation in the coastline. The fundamental difference between tide-dominated estuaries and deltas is that estuaries form during transgression, while deltas actively prograde and net deposition takes place. However, estuaries can become deltas during progradation (Ta et al., 2002; Milli et al., 2013). In tide-dominated deltas, the point where the energy of the fluvial currents suddenly decreases travels up and down river with the flood and ebb tides. Therefore the sediment load of the river is initially deposited over a larger area in tide-dominated deltas. Additionally, strong tidal currents are able to rework the sediment efficiently. Due to the large dispersal of the sediment over the delta, tide-dominated deltas have a much lower gradient than fluvial dominated deltas (Goodbred and Saito, 2012).
The coarsening upward delta front (or distributary tidal mouth bar) successions of tidal deltas are characterized by tidal indicators such as lenticular, wavy and flaser bedding. Towards the top of the delta front succession, cross bedding with mud draped foresets is present. Above that, in the intertidal zone, fining upward tidal flat deposits are present grading up into sub aerial deposits (Figure 2.15)(Harris et al., 1993; Hori et al., 2002; Ta et al., 2002; Allison et al., 2003; Dalrymple et al., 2003; Hampson et al., 2008; Legler et al., 2013). From the top of these coarsening upward successions, tidal and fluvial channels cut down (Figure 2.16). Similar to tide-dominated estuaries, the opposing dominant flood tidal currents (Johnson et al., 1982) and fluvial currents concentrate suspension load muds in one point of the delta. Whereas in estuaries this point is in the apex of the funnel shape, in tide-dominated deltas with high water discharge and high sediment load, this point is hypothesized to be further downstream, near the edge of the delta plain. In some instances the net convergence of sediment can even take place offshore of the delta on the continental shelf, for example in the Amazon delta (Figure 2.14) (Dalrymple and Choi, 2007; Li and Zhang, 1998; Goodbred and Saito, 2012).

In tide-dominated deltas, many types of bar and dune types can form in the distributary channels (Willis, 2005; Plink-Björklund, 2012). Fining up longitudinal bars are deposited between the distributaries in a similar fashion to estuaries. However, in quickly prograding systems, these bars can also be coarsening upward, depending if sediment reworking by tidal flow or initial deposition from river flow is dominant (Dalrymple et al., 2003). Grain size trends can even be more complex when fluid mud is deposited at the seabed and river flow, overlain by a fining up sand grainsize (Willis, 2005; Legler et al., 2013). After avulsion and lobe switching, older parts of tide-dominated deltas still contain old channels, but they lack fluvial discharge and are still connected to the sea. Effectively, because of the lack of fluvial sediment load, these inactive lobes become estuarine (Goodbred and Kuehl, 2000; Allison et al., 2003). On the delta plain, both tidal and fluvial flow is concentrated in channels. No levees form along these channels because flow through the channel is not relatively constant, but highly variable due to the (twice) daily rise and fall of sea level due to the tides. Because of the lack of levees, no crevasse splay deposits are found in the delta plain deposits. During flood-tide, the amount of water
transported through the channels becomes larger than the channels can contain, and the water will spill out of the confines of the channel to form tabular deposits of tidal flats (Willis, 2005).

In contrast to the abundance of transgressive tide-dominated estuarine deposits in the geological record, regressive tide-dominated deltas are more rarely interpreted. In the Devonian of the Baltic states, large scale tide-dominated and tide-influenced deltaic deposition took place, persisting during both regressions and transgressions (Pontén and Plink-Björklund, 2007; Pontén and Plink-Björklund, 2009; Tänavesu-Milkeviciene and Plink-Björklund, 2009). These examples are documented using limited outcrops and extensive core description. Davies et al. (2003) describe Miocene tide-dominated deltaic deposits along the Bay of Bengal and use the nearby Holocene Ganges-Brahmaputra delta as a modern analogue. Legler et al. (2013) describe deposits from the Eocene in Egypt that contain both heterolithic coarsening upward successions and heterolithic channels fills, both with abundant tidal indicators. These deposits represent a delta that was tide-dominated during progradation into a protected embayment with tidal amplification and wave-suppression. The facies and facies architecture are similar to the modern Fly River delta (Dalrymple et al., 2003).

Jurassic syn-rift deltaic deposits from East Greenland have been reported to show evidence for tide-dominance or at least tide-influence (Surlyk, 1990; Surlyk, 1991; Alsgaard et al., 2003; Engkilde and Surlyk, 2003; Ahokas et al., 2014a; Surlyk, 2003; Ahokas et al., 2014b; Eide et al., 2016). Time equivalent strata on the conjugate margin of the rift between Greenland and the Fennoscanndian shield, tide-dominated deposits deltaic deposits have been reported from the subsurface (Martinius et al., 2001; McIlroy, 2004; Martinius et al., 2005; Ichaso and Dalrymple, 2014; Ravnås et al., 2014; Thrana et al., 2014; Ichaso et al., 2016).

The Jurassic Lajas Formation in Argentina has been quoted to be a tide-dominated delta (McIlroy et al., 2005). However, recently these deposits have been re-interpreted to be fluvial-dominated and tide-influenced (Gugliotta et al., 2015; Gugliotta et al., 2016) or mixed-influenced deltas (Rossi and Steel, 2016). Tide-influence from mixed-influenced deltas has been reported from several examples. In the
modern Mekong delta, wave-influence increases during progradation (Ta et al., 2002), or shows at least more evidence for wave-processes in the distal parts of the delta. Another example of mixed-influence deposits where wave-influence is concentrated on the distal parts of the delta, and tide-influence on the delta plain are the Pliocene and Quaternary deposits of the Orinoco delta (Bowman and Johnson, 2014; Chen et al., 2014) and the Cretaceous Sego Sandstone in the Western Interior Seaway of North America (Willis and Gabel, 2001; Willis and Gabel, 2003; Legler et al., 2014). Mainly based on the latter example, Willis (2005) created a model for tide-influenced deltas (Figure 2.17). In this model tide-dominated delta front as described before prograde during the highstand system tract (terminology after Van Wagoner et al., 1990) and the sandstones become sharp-based with low-relief during forced regression during the late highstand- to lowstand system tract. During the lowstand system tract, narrow valleys are incised within earlier deposited sediments. These valleys are filled by tide-dominated facies during the subsequent transgressive system tract.
Figure 2.14. Model for spatial variations (A) of depositional process (B) and texture of the sediment (C) in tide-dominated deltas in analogy to tide-dominated estuaries (after Dalrymple and Choi, 2007).
Figure 2.15. 2D models for tide-dominated deltas. Left: Mekong delta (after Ta et al., 2002), right: Yangtze delta (after Hori et al., 2002). Both models show the general motive of mudstones overlain by coarsening upward delta front successions representing delta progradation. The coarsening upward succession is followed by fining upward successions representing tidal flats and channel fills deposited on the delta plain.
Figure 2.16. Model for deposition in the Holocene Fly River Delta. The vertical successions (B-D) are observed in cores and synthesized in the model in (A) (after Dalrymple et al., 2003).
Figure 2.17. Model for the evolution (A-D) and final state of tide-dominated deltas (after Willis, 2005).

2.4.5.3 **Sequence stratigraphy**

Sequence stratigraphy is the study of how coastlines prograde and retrograde. The developed sequence stratigraphic models originated in seismic stratigraphy, but the models are also widely applied in outcrop and subsurface studies as a predictive tool (e.g. Van Wagoner et al., 1990; Catuneanu, 2006). The model assumes that allogenic (external) mechanisms force the shoreline of deltaic systems to prograde or retrograde. Originally it was thought that eustatic sea-level change with a sinus-shaped curve was the dominant external force. This resulted in terminology being heavily suggestive for- and affiliated with sea-level change, e.g. ‘flooding surface’, ‘highstand system tract’, ‘lowstand system tract’, ‘transgressive system tract’. In addition, different workers use different terminology, which might communication difficult. In recent years, it has been conceded that not only (eustatic) sea-level change is an allogenic force, but also sediment supply and rate of basin subsidence (Catuneanu, 2006).

The sequence stratigraphic model predicts that at relative ‘highstands’, deltas prograde into the basin forming the ‘highstand system tract’ (Figure 2.18). When relative sea level starts to drop, the shoreline trajectory descends accordingly. When the sea-level starts to fall abruptly, rivers start to incise into the lower delta plain, eroding previously deposited sediment, forming an incised valley, and by-passing sediment to be deposited in deep-marine fans (lowstand system tract). This erosion surface is called
the sequence boundary, and separates two genetically un-related sequences: highstand and lowstand system tracts below, and transgressive system tract above. During subsequent transgressions, the incised valley is back filled with fluvial and estuarine deposits with a steadily rising base-level (transgressive system tract). When sea-level rise stops and sea-level stabilizes, the delta starts to prograde again to form another highstand system tract (Van Wagoner et al., 1990). In this model, the shorelines during progradation are wave-dominated, or less frequently, fluvial-dominated. All the tide-influenced facies are placed above the sequence boundary in the transgressive system tract, and the tide-influenced facies are not genetically related to any deposition in the other sequences. More recently, it has been recognized that the highstand system tract deposits can also be deposited in tide-dominated deltas. However, it was still interpreted that the deposits above the first erosion surface (interpreted as sequence boundary) were genetically un-related transgressive sequence tract incised valley fill deposits (Figure 2.16A, Figure 2.17) (Willis, 2005).

Figure 2.18. Model for sequence stratigraphy showing the systems tracts (HST= highstand system tract, LST= lowstand system tract, TST= transgressive system tract) and sequence stratigraphic surfaces (flooding surface, transgressive surface, maximum flooding surface, sequence boundary) (after Van Wagoner et al., 1990).

2.5 Tides in rift basins

Tide-influenced deposits are commonly reported from rift systems. Tectonics activity is in rift basins is concentrated on certain faults such that complex bathymetries and topographies are generated. In
such embayed environments the tides can be amplified due to funneling and wave reworking is dampened due to a shorter fetch in embayed settings.

The syn-rift Nukhul Formation in the Miocene Suez Rift (Carr et al., 2003; Jackson et al., 2005) is an example of a tide-dominated estuary deposited in half-grabens. During deposition, the crest of the tilted half-grabens was emerged and deposition was restricted to the sub-merged deeper parts of the half-graben.

Tide-dominated deltaic systems which prograde along the axis of a rift system have also been reported. Most notably from several Mesozoic rift systems along the north-eastern Atlantic margin. Mellere & Steel (1996) describe a graben bound tide-dominated deltaic deposits which prograded along the axis of a Jurassic rift on the Isle of Skye (Inner Hebrides, United Kingdom). Wonham et al. (1997) report that the tide-influenced deposits of the Bentheim Sandstone (Lower Saxony Basin, Germany) are bound to- and prograde along the axis of grabens. Jurassic tide-influenced deltaic systems are also reported from east Greenland, which are located on the Greenland side of the rift between Greenland and the Fennoscandian Shield. In east Greenland, the tide-influenced and dominated deposits which were confined to N-S trending grabens and prograded along the axis of the rift to the south (Surlyk, 1990; Surlyk, 1991; Alsgaard et al., 2003; Engkilde and Surlyk, 2003; Ahokas et al., 2014a; Surlyk, 2003; Ahokas et al., 2014b; Eide et al., 2016). On the conjugate margin of the rift, on the Halten Terrace, time-equivalent analogue strata in the sub-surface have been reported to be tide-influenced and tide-dominated deltas (Martinus et al., 2001; McIlroy, 2004; Martinus et al., 2005; Ichaso and Dalrymple, 2014; Ravnås et al., 2014; Thrana et al., 2014; Ichaso et al., 2016). Legler et al. (2013) describe a tide-dominated delta complex from the Western Desert of Egypt (Eocene) which does not show any evidence for fluvial or wave processes. It is interpreted that the delta complex was structurally bound to a graben, so deposits are not present on the horst blocks.

A model has been developed for the evolution of shallow marine systems in rift basin (Ravnås and Steel, 1998; Ravnås et al., 2000). The model takes in account the evolution of deep marine fan systems,
shallow marine systems and alluvial systems. The main idea of the model is that the depositional environment will retreat upstream when a rift starts to become active. Examples such transgressions are the Nukhul Formation in the Miocene Suez Rift (Carr et al., 2003; Jackson et al., 2005) and the Åre Formation on the Jurassic Halten Terrace (Thrana et al., 2014). During a (temporary) pause in rift activity, systems can start to prograde again. This is perhaps exaggerated by the increase in sediment supply from intra-basin highs created during the rift episode. Probably the tide-dominate progradational systems mentioned earlier all represent either a pause or final termination in rifting when they are respectively overlain by marine mudstones or alluvial deposits (Ravnås and Steel, 1998; Ravnås et al., 2000).

2.6 Hydrocarbon reservoir implications

Tide-influenced deposits are often challenging hydrocarbon reservoirs because 1) internal heterogeneity in mudstone and sandstone content and 2) because of the limited lateral extend of tide-influenced facies. Because of their heterogenic nature, it often hard to recognize clear patterns in geophysical wireline logs. Hence, there is a strong reliance on cored sections, which are increasingly expensive to cut, so there is limited data on the facies in the first place. The second challenge is in correlation due to the limited lateral extend of facies away from wells, and the related modelling of fluid flow away from wells (Martinius et al., 2005; Ainsworth et al., 2011). Due to these limitation, outcrop analogues with a wide lateral extend are important. These outcrops can provide clues about the lateral extend of facies away from vertical data sources (i.e. wells).
Chapter 3. Fluvial-to-tidal transition in proximal, mixed tide-influenced and wave-influenced deltaic deposits: Cretaceous lower Sego Sandstone, Utah, USA

3.1 Introduction

Tide-dominated deltaic deposits are the result of a complex interplay of depositional processes, which change from proximal to distal locations along the axes of distributary channels and onto the delta front. Fluvial processes dominate in landward locations, but these increasingly interact with tides in a coast-normal orientation, along the landward to seaward sediment transport path (Plink-Björklund, 2005; Van den Berg et al., 2007; Salahuddin and Lambiase, 2013). The changing balance of fluvial and tidal processes causes predictable variations in hydrodynamic energy, grain size, sand-mud ratio and facies types (Dalrymple and Choi, 2007). Wave reworking at the coastline may further complicate the depositional systems, for example via along-shore redistribution of sediment (Legler et al., 2014). The relative strength and interplay of these riverine, tidal and wave-driven depositional processes exerts a primary control on the dimensions, geometry, orientation and distribution of sediment bodies and facies belts in the geological record (Ainsworth et al., 2011).

Most ancient tide-influenced and tide-dominated deposits are interpreted to be deposited along transgressive coastlines. Many of the depositional models of these ancient transgressive tide-influenced systems (Pontén and Plink-Björklund, 2009; Sixsmith et al., 2008; Dalrymple, 2006a; Plink-Björklund, 2005; Yoshida et al., 2004; e.g. Yoshida et al., 2001) are based on early work on well-studied modern estuaries and barrier-back barrier systems in high latitudes which developed during the Holocene transgression (Terwindt, 1971; Reineck, 1967; Evans, 1965; Van Straaten, 1961; Oomkens and Terwindt, 1960; e.g. Van Straaten and Kuenen, 1957). Detached tidal ridges are interpreted in many ancient transgressive shelf mudstone successions, consistent with early work on modern analogues (Zhuo et al., 2014; Olariu et al., 2012; Gaynor and Swift, 1988; Tillman and Martinsen, 1984; Boyles and Scott, 1982; e.g. Houbolt, 1968). A second line of approach to interpret tide-influenced deposits uses sequence stratigraphic models in which they are interpreted to be deposited in incised
valleys during transgressions (Posamentier and Allen, 1999; Dalrymple et al., 1992; e.g. Dalrymple, 1992). A common feature for tide-dominated deposits in both types of interpretation mentioned above is that they form in transgressive settings, although we know from modern examples that many prograding tide-influenced and tide-dominated deltaic systems also exist. In particular, large modern tide-dominated river deltas in the tropics have been studied in detail (Hasan Sidi et al., 2003; Dalrymple et al., 2003; Ta et al., 2002; Allen and Chambers, 1998; e.g. Harris et al., 1993). However, due to pre-existing models which place tide dominated facies in transgressive environments such as estuaries (Boyd et al., 1992) and within a sequence stratigraphic framework within a “transgressive system tract” (Van Wagoner et al., 1991), prograding tide-dominated (Chen et al., 2014; Legler et al., 2013; Tänavsuu-Milkeviciene and Plink-Björklund, 2009; McIlroy et al., 2005; Martinius et al., 2001; Maguregui and Tyler, 1991; Yang and Nio, 1989), mixed tide-fluvial influenced (Pontén and Plink-Björklund, 2007) and mixed tide-wave influenced (Willis et al., 1999; Mellere and Steel, 2000; Willis and Gabel, 2001; Seidler and Steel, 2001) deltaic systems have for a long time not been interpreted frequently in ancient strata. Facies architecture and stratigraphic relationships have been documented in several ancient tide-influenced or tide-dominated deltas around the world. However, the facies architecture of the area where fluvial and tidal processes interact has yet to be characterised in such deltaic deposits, while this fluvial- to tidal transition is widely recognized in corresponding transgressive systems (e.g. Dalrymple et al., 1992; Plink-Björklund, 2005).

In order to develop a sound understanding of the facies relationships and stratigraphic architectures resulting from the interaction of fluvial and tidal processes, well-documented outcrop examples are needed in which vertical and lateral facies dimensions and spatial relationships can be measured. One such outcrop dataset is provided by the Late Cretaceous lower Sego Sandstone, which is exposed in the almost continuous Book Cliffs (Utah and Colorado, USA) over a distance of c. 150 km (e.g. Young, 1955). Overall the Sego Sandstone in this area is widely interpreted as an ancient tide-dominated or tide-influenced delta complex, because it records a succession of nearshore sandstones bearing a suite of tidal structures, which prograded over marine mudstones and which are overlain by non-marine
deposits (e.g. Young, 1955; Willis and Gabel, 2001; Legler et al., 2014). This interpretation includes subordinate components of the lower Sego Sandstone, which are considered to be estuarine (Van Wagoner, 1991; Willis, 2000). The central and eastern parts of the outcrop belt, which contain medial to distal deltaic deposits in which tidal and wave processes interacted, have been extensively documented in detail in previous studies (Van Wagoner, 1991; Willis and Gabel, 2001; Willis and Gabel, 2003; Wood, 2004; Burton and Wood, 2011; Legler et al., 2014; Burton et al., 2016). In these studies, tide-influenced facies are to always at least to some extent interpreted to be deposited in incised valleys overlying a sequence boundary. In contrast, more proximal deposits in the western part of the outcrop belt have been described to date only in large, regional studies (Yoshida et al., 1996; Willis, 2000; McLaurin and Steel, 2000; Hettinger and Kirschbaum, 2002; Aschoff and Steel, 2011a). In this paper, we aim to develop an improved understanding of the proximal deposits of the Sego Sandstone, in which fluvial and tidal processes interacted. This includes re-interpretation of incised valley fills. The hypothesis is that at least some of the deposits previously interpreted as incised valley fills are tide-influenced fluvial distributary channels. This aim is achieved by documenting facies characteristics within a framework of carefully traced stratigraphic surfaces and architectural elements in the part of the Sego Sandstone outcrop belt that has previously only been covered by regional studies. The observations made in these continuous, well exposed outcrops will be used to construct a depositional- and facies model. These models can be used as a guide for correlations and interpretations of the fluvial to tidal transition in proximal parts of tide-influenced deltaic deposits in more data-poor areas elsewhere.

3.2 Geological setting

The Sego Sandstone is part of the Mesaverde Group, which was deposited during the Campanian (Late Cretaceous) in the foreland of the North American Cordillera. To the west, the Farallon Plate was being subducted below the North American continent, resulting in the formation of a magmatic arc and the thin-skinned Sevier Orogen fold-and-thrust-belt. A combination of flexural subsidence due to loading of the orogen and dynamic subsidence due to subduction of the cold Farallon Plate slab below the
continent resulted in the creation of a wide basin in the centre of the North American continent (DeCelles, 2004; Liu and Nummedal, 2004; Liu et al., 2014). This wide basin was eventually flooded to form the Western Interior Seaway (Figure 3.1A). Sediment was transported from the Sevier Orogen eastward into the Western Interior Seaway to deposit a series of clastic coastal sediment wedges, which record repeated shoreline progradation and retrogradation (e.g. Young, 1955).

The Sego Sandstone forms part of the progradational component of one such wedge (Figure 3.2) and has a duration of about 2.5 Myr (Gill and Hail, 1975; Cobban et al., 2006). It overlies the marine Buck Tongue of the Mancos Shale and is overlain by coal-bearing, lower coastal plain deposits of the Neslen Formation. The Sego Sandstone is informally sub-divided into lower and upper intervals that are separated by the marine Anchor Mine Tongue of the Mancos Shale. Towards the east, both intervals of the Sego Sandstone pass into the Mancos Shale (Young, 1955). Towards the west, the Sego Sandstone correlates with the middle, relatively mudstone-rich part of the Castlegate Sandstone (Yoshida et al., 1996; McLaurin and Steel, 2000; Willis, 2000; Aschoff and Steel, 2011a). This mudstone rich middle part of the Castlegate Sandstone is generally poorly exposed due to the high clay content. Where exposed, the facies are similar to the Neslen Formation (Young, 1955; Shiers et al., 2014) and include overbank mudstones, fluvial channels, tide-influenced fluvial channels and mudstone rich channel fills which are all interpreted to be deposited in lower delta plain environments (Willis, 2000; Aschoff and Steel, 2010a).

The lower Sego Sandstone crops out in a 150 km long outcrop belt in the Book Cliffs between the towns of Green River, Utah, and Grand Junction, Colorado (Figure 3.1B). The central portion of this outcrop belt is oriented SW-NE, parallel to the interpreted palaeo-coastline trend (i.e. along depositional strike). In this central area, the lower Sego Sandstone consists of offshore mudstones, storm-generated hummocky cross-stratified (HCS) sandstones, and heterolithic sandstones with a strong tidal signature (Van Wagoner, 1991; Willis and Gabel, 2001; Willis and Gabel, 2003; Wood, 2004; Burton and Wood, 2011). The eastern part of the outcrop belt is a W-E-trending section, which is oriented oblique to
depositional dip. In this section, the lower Sego Sandstone comprises erosionally based tidal sandstones that interfinger with, and pass eastward (offshore) into, wave-rewilded sandstones. This relationship implies that waves and tides operated at the same time, at least in the proximal part of the delta front. In such settings, preservation bias may result in the stratigraphic partitioning of tide- and wave-dominated deposits in shallower and deeper water environments, respectively (Legler et al., 2014).

Tidal deposits are erosionally based, with either tabular or lenticular geometries. Originally, such tidal sand bodies were all interpreted as the infill of incised valleys, with their erosional bases defining nine sequence boundaries (Van Wagoner, 1991; Yoshida et al., 1996; Willis, 2000). More recently, many of these erosion surfaces have been attributed to tidal scour inherent in the formation, migration and abandonment of tide-influenced deltaic distributary channels (Willis and Gabel, 2001; Willis and Gabel, 2003). These latter authors have interpreted three regressive-transgressive tongues in the lower Sego Sandstone, each comprising (1) a coarsening-upward succession of offshore mudstones passing gradationally into hummocky cross-stratified sandstones, which are interpreted as wave-dominated shoreface deposits, and (2) an overlying, erosionally based unit comprising heterolithic tidal sandstones, which are interpreted as tidal bar and channel-fill sand bodies. The high-relief (up to 30 m) erosion surfaces that cut down from near the top of each upward-coarsening succession are interpreted as sequence boundaries (Willis and Gabel, 2003). The stratigraphic architecture of each regressive-transgressive tongue was interpreted to record a temporal change in process regime, with wave-dominated deposits assigned to an early highstand systems tract, and overlying tide-dominated deposits assigned to late highstand system tracts and the high-relief sequence boundaries being formed during lowstand systems tracts (Willis and Gabel, 2001). In other recent works, the high relief erosion boundaries are also interpreted as sequence boundaries forming an incised valley which has a transgressive infill (Burton et al., 2016). Therefore, since first detailed work on the Sego Sandstone (Van Wagoner, 1991), the interpretation of the location and abundance of sequence boundaries has changed over time. In this study, the aim is to re-evaluate the interpretation of the high-relief erosion
surfaces as sequence boundaries forming the base of incised valleys. The hypothesis is that these high relief erosion surfaces can also be distributary channels. The best location to test the interpretation of a distributary channel is in the proximal, depositional up-dip locations in the western part of the outcrop belt of the Sego Sandstone, because distributary channels are usually deposited in relatively proximal locations.

Figure 3.1. (A) Palaeogeographic reconstruction of the North American continent during the Campanian. The black lines indicate modern coastline positions. The Sego Sandstone outcrop belt (Figure 3.1B) is indicated by the red box (modified after Blakey, 2013). (B) Map showing the study area in the context of the Sego Sandstone outcrop belt in the Book Cliffs and the gross interpreted palaeogeography of the unit (modified after Willis and Gabel, 2001). (C) Map of the study area showing the Sego Sandstone, and the location of stratigraphic logs and photo panoramas discussed in this paper (modified after Doelling, 2002; Gualtieri, 2004; Witkind, 2004).
Figure 3.2. Lithostratigraphy of the Sego Sandstone and neighbouring units. The red box indicates the studied interval (Cobban et al., 2006; Kirschbaum and Hettenger, 2004; McLaurin and Steel, 2000; Fouch et al., 1983; Gill and Hail, 1975; modified after Young, 1955).

3.3 Dataset and methods

Twenty-nine stratigraphic sections have been measured in a 25 km long, continuously exposed section of the Book Cliffs outcrop belt between Tusher Canyon and Crescent Canyon in east-central Utah (Figure 3.1C). Strata in this area have a uniformly low (<5˚) tectonic dip towards the north. Stratigraphic logs are measured where canyons cut through the cliff face. In between the measured logs, stratal units have been traced in cliff faces using photo-panoramas.

Facies associations are identified based on lithology, sedimentary structures and ichnology, and are characterised and interpreted using facies analysis. Then the vertical stacking of facies associations in successions is defined and estimates for palaeo-water depth relative to effective storm-
fairweather-wave base are made in order to interpret sequence stratigraphic surfaces and the units that they bound. Lastly, the dimensions, geometries and lateral distributions of facies associations are characterised within the context of sequence stratigraphic units. This is achieved using 29 of the measured sections, with an average spacing of ~1km, and photo-panoramas of the intervening cliff exposures. The resulting architectural interpretations therefore follow the rugose map-view pattern of the Sego Sandstone outcrops.

In addition, eight samples that are representative of the sedimentological facies were processed and analysed for their palynological content, and their palynofacies are documented herein. Palynological studies have been rarely undertaken in the Book Cliffs (Oboh-Ikuenobe, 1996), but may provide important additional information for facies analysis of tide-influenced deltaic systems (Stukins et al., 2013). The samples were treated with concentrated hydrochloric acid (37%) where necessary to remove carbonate material and then treatment with hydrofluoric acid (40%) to remove silicates and leave an organic residue. All residues were sieved using a 15 µm polyester mesh and then mounted onto glass slides using a visible light curing mounting medium (Rapidslide).

3.4 Facies Associations

Five facies associations (FA) have been recognized based on observations of lithology, grain size and sorting, sedimentary structures, palaeocurrents, body fossils, and trace fossils. Bioturbation intensity is recorded using the bioturbation index (BI) scheme of Taylor & Goldring (1993). A summary of the facies association descriptions is given in Table 3.1. Examples of facies are given in photo-panels (Figure 3.3-6), example logs showing the vertical facies successions (}
Figure 3.7), and a correlation panel (
Figure 3.8) and panoramas (Figure 3.9) showing the lateral facies architecture are given as well.

Palynofacies analysis was undertaken using a simple count scheme of categories (Table 3.2), and a minimum of 300 particles were counted for each sample before normalising as a percentage. Other organic particles (e.g. fungal remains, monolete spores) that do not appear in Table 3.2 were not encountered in the palynofacies count. One sample (FA4-1) was barren of any discernible organic residue. Analysis of the palynofacies data used concepts from Traverse (1994) and Tyson (1995). The interpretation of the palynofacies method used herein is relatively simple and distinguishes pollen and spores of terrestrial flora (Trilete Spores, Bisaccate Pollen, and Other Pollen in Table 3.2), marine microfauna (Dinoflagellates) and land-derived plant and wood debris (other columns in Table 3.2). A first order palynofacies interpretation is based on where the studied organic material is derived from (land and/or marine).
<table>
<thead>
<tr>
<th>FA#</th>
<th>FA Name</th>
<th>Lithology</th>
<th>Primary sedimentary structures</th>
<th>Biogenic structures</th>
<th>Geometry</th>
<th>Dimensions</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Bioturbated mudstones</td>
<td>Mudstones</td>
<td>Siltstone and very fine grained sandstone beds (&lt;10 cm thick) with sharp bases and sharp and gradational tops, internally low angle cross stratification</td>
<td>BI=0-4. Palaeophycus, Planolites, Schaub cylindrichnus, Thalassinoides, Skolithos, Arenicolites</td>
<td>Planar</td>
<td>5 to more than 10 m thick x more than 25 km wide</td>
<td>Offshore mudstones</td>
</tr>
<tr>
<td>2</td>
<td>Humocky cross stratified sandstones</td>
<td>Sandstones and mudstones</td>
<td>Humocky cross stratification</td>
<td>BI=0-2. Skolithos, Palaeophycus, Ophiomorpha, Planolites, Arenicolites, Dilipocrateron</td>
<td>Planar</td>
<td>2 to 10 m thick x 10 to more than 25 km wide</td>
<td>Lower shoreface</td>
</tr>
<tr>
<td>3</td>
<td>Ripple cross-laminated heterolithic sandstones</td>
<td>Heterolithic sandstones</td>
<td>Lenticular, wavy and flaser bedding. Symmetrical and a symmetrical cross-laminations. Wave ripples and current ripples, Bidirectional paleocurrent directions</td>
<td>BI=0-2. Ophiomorpha, Thalassinoides, Skolithos, Planolites, Palaeophycus, Gyrochorte</td>
<td>Planar or lenticular</td>
<td>0 to 4 m thick x 1 to 8 km wide</td>
<td>Tide-dominated delta front and tide-dominated channel fills</td>
</tr>
<tr>
<td>4</td>
<td>Cross-bedded sandstones</td>
<td>Sandstones</td>
<td>Cross-bedding with draped foresets. Rippled surfaces in the foresets.</td>
<td>BI=0-2. Ophiomorpha, Skolithos, Palaeophycus, Planolites, Gyrochorte, Arenicolites, Rosselia, Cylindrichnus, Teredolites</td>
<td>Planar</td>
<td>1 to 10 m thick x 1 to more than 25 km wide; Lenticular: 1 to 15 m thick x 300 m wide x 2 to 4 km long</td>
<td>Longitudinal tide-influenced mid-channel bar and tidal channel fills</td>
</tr>
<tr>
<td>5</td>
<td>Sharp-based mudstones and heterolithic mudstones</td>
<td>Mudstones and heterolithic mudstones</td>
<td>Lenticular and wavy bedding. Rich in organic matter.</td>
<td>BI=0-1. Planolites</td>
<td>Lenticular</td>
<td>0.5 to 5 m thick x 0.5 to 2 km wide</td>
<td>Abandoned channels and tidal creeks</td>
</tr>
</tbody>
</table>

Table 3.1. Summary of facies associations.
3.4.1 Facies Association 1: Bioturbated mudstones

3.4.1.1 Description

FA1 consists of dark grey to rarely black homogenous claystones that grade into silty claystones, mudstones and pale yellow siltstones. In the claystones and mudstones, thin (<10 cm) siltstone and very fine-grained sandstone beds with sharp or gradational boundaries are present (Figure 3.3A). Some siltstone beds contain low-angle cross-lamination and have a sharp lower boundary and a gradational, sharp or symmetrically rippled top. Lenticular siltstone-mudstone alternations and wavy siltstone-mudstone alternations are present locally. Bioturbation is generally low to moderate in intensity (BI of 2-3), but is locally absent (BI of 0) and high (BI of 4). Trace fossils include *Palaeophycus*, *Planolites*, *Schaubcyclindrichnus*, *Thalassinoides*, *Skolithos* and *Arenicolites*, which in combination constitute the Cruziana ichnofacies (MacEachern and Bann, 2008).

The palynofacies of FA1 is characterised by a dominance of translucent phytoclasts (67-74%) and the lowest assemblage percentage for cuticle material. There are no marine palynomorphs (dinoflagellates or foraminiferal test linings) or Amorphous Organic Matter (AOM) within the count, but pollen and spores are present.

3.4.1.2 Interpretation

The fine grain size and occurrence of the Cruziana ichnofacies indicate deposition in a quiet, open-marine environment below effective storm-wave base (MacEachern and Bann, 2008). Siltstone and very fine-grained beds indicate episodic transport of coarser-grained sediment, probably during high river discharge and/or storm events (Wright, 1977; Bhattacharya and MacEachern, 2009). Although identified by the ichnofabric as marine, there are no marine palynomorphs in the palynofacies count. This is most likely because the assemblage is dominated by phytoclasts that drown out values of other organic particles. Further qualitative examination of only the palynomorphs shows an abundant assemblage of dinoflagellate cysts and several specimen of foraminiferal test linings, therefore
indicating that marine surface- and benthic conditions were present. The dominance of translucent phytoclasts in the palynofacies assemblage exceeds the presence of opaque phytoclasts, which is consistent with a deltaic environment that was relatively distal to any trunk channel system (Hart et al., 1994; Tyson, 1995). This facies association corresponds to the mudstone part of the “transgressive and highstand system tract facies” of Van Wagoner (1991), the mudstone part of facies C of Willis (2000), facies 1 of Willis & Gabel (2001) and to the bioturbated mudstone intervals of facies association 1 of Legler et al. (2014).

### 3.4.2 Facies Association 2: Hummocky cross stratified sandstones

#### 3.4.2.1 Description

FA2 consists of thin (10-50 cm), tabular beds of very fine- and lower fine-grained sandstone which are interbedded with mudstones with the same characteristics as FA1. Individual beds can be traced laterally over distances of up to 1 km, and groups of beds can be traced over distances of up to 10 km. The beds have a sharp base and either a sharp, rippled or a gradational top. Internally the beds contain HCS with a wavelength of 0.5-2m and a preserved bed thickness of 10-50 cm (Figure 3.3B). Bioturbation is generally absent to low in intensity (BI of 0-2). Trace fossils include *Skolithos, Palaeophycus, Ophiomorpha, Planolites, Arenicolites* and *Diplocraterion*, constituting a mixture of the Cruziana and Skolithos ichnofacies (MacEachern and Bann, 2008). The sandstone beds alternate with bioturbated mudstones of FA1. In general, FA2 gradationally overlies FA1, and occurs in coarsening-upward successions in which sandstone beds are increasingly erosionally amalgamated towards the top. This FA is overlain by bioturbated mudstones of FA1 or is erosionally overlain by cross-bedded sandstones of FA4. No organic material is preserved in the sandstones of this FA, so no palynofacies could be determined.

#### 3.4.2.2 Interpretation

The presence of sandstone beds displaying HCS indicates deposition by episodic storm events on the lower shoreface in water depths between the effective storm wave base and fairweather-wave base
This interpretation is supported by the occurrence of a mixed Cruziana and Skolithos ichnofacies (MacEachern and Bann, 2008). The upward increase in the abundance and amalgamation of storm-event beds indicates that storms became more frequent towards the top of the succession, due to an increasingly energetic wave climate and/or progressive shallowing of water depth (Storms and Hampson, 2005). This facies association corresponds with the “transgressive and highstand system tract facies” of Van Wagoner (1991), facies C of Willis (2000), facies 2 of Willis & Gabel (2001) and with intervals of HCS sandstone interbedded with bioturbated mudstones in facies associations 1 and 2 of Legler et al. (2014).

Figure 3.3. Facies associations 1 and 2. (A) Mudstone with thin siltstone beds and sparse bioturbation by *Planolites* (FA1). (B) Very fine-grained sandstone bed containing HCS (FA2).
3.4.3 Facies Association 3: Ripple cross-laminated heterolithic sandstones

3.4.3.1 Description

This facies association consists of very fine- to fine-grained sandstones and mudstones that exhibit flaser, wavy and lenticular bedding (Reineck and Wunderlich, 1968). The sandstone-to-mudstone ratio in the facies association is 0.5-0.9. Ripple cross-lamination and both symmetrical and asymmetrical rippled surfaces are present (Figure 3.4A, B, D). Asymmetrical ripples with bedform heights of 1-2 cm and wavelengths of 5-10 cm display apparent bidirectional palaeocurrent patterns. Along individual rippled surfaces, lateral changes from asymmetrical to symmetrical ripples occur. Up to centimetre-scale mudstone clasts are common. Coal fragments up to 1 cm in diameter and carbonaceous mudstone drapes are also present. Bioturbation is generally absent to low in intensity (BI of 0-2), and trace fossils include *Ophiomorpha*, *Thalassinoides*, *Skolithos*, *Planolites*, *Palaeophycus* and *Gyrochorte* (Figure 3.4C). The trace fossil assemblage constitutes an impoverished, mixed Cruziana and Skolithos ichnofacies (MacEachern and Bann, 2008). This facies association alternates with cross-bedded sandstones of FA4. In these alternations, the lower boundary of FA3 is marked by a sharp decrease in sandstone grain size and an increase in mudstone content. The upper contact of FA3 is defined by the erosional base of FA4. Towards the top of the lower Sego Sandstone, FA3 is deposited above or below FA5, and contacts between the two facies associations are sharp.

There is a variable palynofacies characterisation for FA3, however there is consistently a relatively high percentage of cuticle material in all samples. The highest numbers of dinoflagellate cysts and resin are present in palynofacies counts from FA3 samples. Phytoclasts are the dominant organic particle. The type of phytoclast however, is not consistent. The samples containing the dinoflagellate cysts within the palynofacies count have a far higher proportion of translucent phytoclast than the samples, which lack dinoflagellate cysts and contain a greater proportion of equidimensional, opaque phytoclasts.
3.4.3.2 Interpretation

The abundance of asymmetrical ripple forms and current ripple cross-lamination indicate that sand was predominantly deposited as ripples migrating in response to unidirectional subaqueous currents. The occurrence of symmetrical ripple forms also shows that oscillatory waves were dominant episodically. The interbedding of sandstone and mudstone beds indicates temporal changes in current velocities. The resulting heterolithic facies contain apparent bidirectional palaeocurrents, which supports deposition by tidal processes (Van den Berg et al., 2007; Terwindt, 1971; Reineck and Wunderlich, 1968; Oomkens and Terwindt, 1960). The tidal reworking of the sediment on a daily basis suggests deposition above fairweather wave base. The low intensity of bioturbation by an impoverished Cruziana and Skolithos ichnofacies is consistent with deposition in a stressed environment characterised by fluctuating energy levels, a mobile sediment substrate and/or restricted marine salinities (MacEachern and Bann, 2008). These stressed environments and the close association and alternations with FA4 suggest deposition in a lower energy, tide-dominated channels. The subtle coarsening upward grain size trend within FA3 is not necessarily in contradiction with deposition in a tide-dominated channel, coarsening upward and cleaning upward trends have been observed in tide-dominated channels in both the modern (Dalrymple et al., 2003) and ancient (Legler et al., 2014). The heterolithic nature of this facies association is reflected in the palynofacies values. The higher percentage of cuticle material in samples of FA3 than those of FA1, may indicate a closer proximity to a significant sediment fairway (Hart et al., 1994; Tyson, 1995). The greater ratio of equidimensional to bladed opaque phytoclasts corroborates this conclusion. Samples of FA3 also contain a greater percentage of opaque phytoclasts that were most likely deposited in association with a proximal supply of terrestrial sediment. This facies association corresponds with the thinly bedded, finer grained part of the “lowstand system tract facies” of Van Wagoner (1991) and facies 3 of Willis & Gabel (2001).
Figure 3.4. Facies association 3. (A), (B) and (C) Current rippled and cross-laminated heterolithic sandstones exhibiting apparent bidirectional palaeocurrents. The black and grey arrows indicate cross lamination and current ripples of apparent opposed palaeoflow directions. (D) Three dimensional current ripples indicating palaeoflow towards the SE.
3.4.4 Facies Association 4: Cross-bedded sandstones

3.4.4.1 Description

Facies association 4 occurs in units with sharp, planar to erosional bases that are locally lined by wood debris, mudstone clasts and rare shell debris (Figure 3.5C). Above the basal surface, trough cross-bedded, fine- to medium-grained sandstones are present, marking an abrupt increase in grain size compared to the underlying very fine- to fine-grained sandstones of FA2 or FA3. Trough cross-beds occur in sets that are generally 20-40 cm thick. Sets are stacked into cosets that occur in sandstone beds defined by a sharp erosional base and with a top marked by abrupt fining in grain size or by a sharp erosional contact at the base of the overlying unit. In such beds, cross-sets indicate unidirectional palaeocurrents with a large spread of c. 90°. Locally, bidirectional palaeocurrents in cosets are observed (for all measured palaeo-current direction see Figure 3.10). The sandstones may be devoid of finer grained material, but commonly contain variable amounts of claystone drapes, carbonaceous mudstone drapes, a combination of claystone- and carbonaceous mudstone drapes, mudstone flasers and mudstone clasts. Mud-drapes line foresets, which merge and thicken down-dip into mudstone and sometimes rippled heterolithic toesets. Down depositional dip, mudstone drapes are rhythmically more widely spaced and more closely spaced due to thinning of intervening sandstone foresets. There are no clear structural cycles in these groupings of mudstone drapes (Figure 3.5A, B). Reactivation surfaces are absent. Units of this sandstone-dominated facies (>90% sand) contain internal low-angle (generally 5° and up to 10°) accretion surfaces dipping towards the NE and SW, but other dip directions occur as well. These inclined surfaces have been observed in large cliffs (100’s of m’s wide) where such subtle inclined surfaces can be traced out. The orientation of these surfaces can be determined in amphitheatres where two or three apparent dip directions can be combined to reconstruct the true orientation of these planes. Coaly clasts and wood debris are sparse and are bored by rare *Teredolites*. Bioturbation is generally absent (BI of 0), but sparse bioturbation (BI of 1-2) is locally present. Trace fossils are dominated by *Ophiomorpha* and *Skolithos*, but also include *Palaeophycus*, *Planolites*, *Gyrochorte*, *Arenicolites*, *Rosellia* and *Cylindrichnus*. This assemblage constitutes a mixture of the
Cruziana and Skolithos ichnofacies (MacEachern and Bann, 2008). The top of FA4 is often free of mudstone, it is usually completely reworked by *Ophiomorpha* and less commonly by *Skolithos* (BI of 5-6) (Figure 3.5D), and contains spherical red concretions (up to c. 50 cm in diameter). No organic material is preserved in the sandstones of this FA, so no palynofacies could be determined.

### 3.4.4.2 Interpretation

The prevalence of cross-bedding indicates deposition from subaqueous dunes that migrated in response to relatively strong currents. The geometry, abundance and distribution of mudstone drapes are typical of those generated by tidal currents above fairweather wave base, which fluctuate on a daily basis (Allen, 1980; Visser, 1980; Van den Berg et al., 2007). Where carbonaceous mudstone drapes are abundant, it is interpreted that the channel was a main conduit for down-stream currents carrying organic debris form the lower delta plain. Where claystone drapes are more abundant, it is interpreted that most of the water flowing through the channel was derived from the sea, with no, or just a minor connection for downstream flow from the lower delta plain. The mudstone drapes are organised in denser and less dense intervals. This could be the result from spring-neap cycles, or the results of waxing- and waning downstream currents, which can be seasonal. The interpretation of tidal currents is further supported by the presence of bidirectional palaeocurrents. Based on regional palaeogeographic considerations (Willis and Gabel, 2001; Wood, 2004; Burton et al., 2016), palaeocurrents directed towards the NW are interpreted as flood tidal currents and those directed towards the SE are interpreted as ebb and fluvial currents. The low diversity of the trace fossil assemblage is consistent with deposition in a high-energy marine environment, possibly with brackish water conditions. The sharp, lag-lined bases of units of FA4 are interpreted as tidal erosion surfaces, possibly enhanced by fluvial discharge, at the base of tidal channels. The cosets of inclined surfaces indicate dune migration across larger bar-forms. Where units of FA4 exhibit planar, laterally continuous bases, inclined bedding may also indicate the presence of highly mobile bar-forms. This interpretation indicates that multiple parallel bar forms covered, and migrated in an area with a water depth at which tidal currents affected bed-load transport in the form of subaqueous dunes (Houbolt,
The abundance of wood debris, coal clasts and carbonaceous mudstone drapes indicates nearby vegetated areas, consistent with an emergent deltaic to coastal plain setting. The intensely bioturbated tops of units of FA4 indicate an increase in biogenic activity, probably associated with reduced sedimentation following channel abandonment. This facies association corresponds with the coarser grained cross stratified part of the “lowstand system tract facies” of Van Wagoner (1991), facies D of Willis (2000), facies 4 and 5 of Willis & Gabel (2001) and facies associations 6 and 7 of Legler et al. (2014).
Figure 3.5. Facies association 4. (A) and (B) Mud draped, cross-bedded heterolithic sandstones of FA4. Palaeocurrents are unidirectional except for the top beds in (A) which show an opposed apparent palaeoflow directions (grey and black arrows). Toesets consist of cross-laminated wavy and lenticular bedding. In (A) mudclasts are randomly scattered and aligned along bed bases and foresets. (C) Lag of mudstone clasts and wood debris at the erosional base of a unit of FA4. (D) Bioturbation by Ophiomorpha at the top of a bed in FA4.
3.4.5 Facies Association 5: Sharp-based mudstones and heterolithic mudstones

3.4.5.1 Description

Facies association 5 consists of dark grey to black, carbonaceous mudstones. Very fine-grained sandstone is present in intervals of lenticular and wavy bedding (c. 0.1 to 10 cm thick). The sandstone-to-mudstone ratio of the facies association is 0-0.5. Units of FA5 have sharp concave-upward, erosional bases above which there is an upward increase in mudstone content (Figure 3.6A, B). In some cases, sharp internal boundaries are present which are overlain by very fine grained sandstones that rapidly fine upwards into mudstones (Figure 3.6C). Each unit is thus channelised, with a lateral extent of up to 1 km. Some internal bedding surfaces within channelised units of FA5 are inclined, and associated with abrupt lateral changes in sandstone-to-mudstone ratio (inclined heterolithic strata, IHS, sensu Thomas et al., 1987). Bioturbation is generally absent (BI of 0), with sparse Planolites present locally (BI of 1). The palynofacies for FA5 is very similar to that of FA1, however with the addition of dinoflagellate cysts and AOM (Amorphous Organic Matter) included in the count data.

3.4.5.2 Interpretation

The concave-upward erosional bases of units of FA5 record deposition in channels. The low sandstone-to-mudstone ratio of the facies association indicates deposition predominantly from suspension due to low flow velocities, high suspended sediment concentrations and/or a deficiency of sand supply. The occurrence of lenticular and wavy bedding signifies intermittent current activity through the channels (Legler et al., 2013; Staub and Gastaldo, 2003; Oomkens, 1974; Van Straaten and Kuenen, 1957). Furthermore, IHS indicates accretion of a mobile substrate within laterally migrating channels (Choi et al., 2004; Thomas et al., 1987; Reineck, 1958). The abundance of carbonaceous organic material indicates that nearby vegetation was eroded by and transported through the channels (Oomkens, 1974; Staub and Gastaldo, 2003). The near absence of bioturbation indicates a highly stressed environment (MacEachern and Bann, 2008). FA5 is interpreted as the fills of low-energy to inactive (abandoned) deltaic distributary channels in a shallow water environment on the lower delta.
plain. It is likely that these channels retained an open connection with the sea during abandonment, forming estuarine channels (Dalrymple et al., 1992). Abundant organic material is inferred to have been derived from surrounding supra-tidal salt marshes on the lower delta plain. The similarities of the palynofacies with FA1 indicate that the deposition was not directly connected to a fluvial source, but is instead consistent with the deposition of fine grained material by tidal currents within abandoned channels. Funnelling of more saline waters along abandoned tidal channels can often extend the range of marine palynomorphs to potentially more proximal locations, explaining the occurrence of dinoflagellate cysts and AOM (Amorphous Organic Matter) at this position (Hubbard et al., 2011). This FA corresponds with facies A of Willis (2000). None of the FA’s show evidence for sub-aerial exposure.
Figure 3.6. Facies association 5. (A) Lenticular channel fill of FA5 that erosionally truncates deposits of FA4. Mudstones drape the concave-up base of the channel and are themselves truncated by an overlying channel-fill sandstone of FA4. (B) Organic-rich mudstones of FA5 that truncate underlying rippled heterolithic sandstones (FA3) and which are erosionally overlain by cross-bedded sandstones (FA4). (C) Stacked mudstone channel fills of FA5. Individual channel fills fine upward and their erosional bases are marked by sharp increases in grain size. Triangles mark grain size trends within each channel fill.
3.5 Facies successions

3.5.1 Wave-dominated successions $S_1$ and $S_0$

Two upward-coarsening successions characterise the lower part of the studied interval, each comprising FA1 gradationally overlain by FA2. The two facies successions extend across most of the study area.
Figure 3.7, 3.8). The lower upward-coarsening succession is named S₁ by extending the nomenclature of Willis & Gabel (2003) for the Sego Sandstone in exposures directly to the east of the study area. It has a thickness of 3 to 11 m. The overlying upward-coarsening succession, S₀, has a thickness of 3 to 18 m.

The upward-coarsening trend in each succession may represent an increase in storm energy, an increase in sand supply and/or an upward shallowing in palaeo-water depth (Storms and Hampson, 2005; Sømme et al., 2008).

3.5.2 Mixed, wave- and tide-influenced succession S₁

A third upward-coarsening succession comprising FA1 and FA2 is erosionally overlain by cross-bedded sandstones of FA4 (sandstone S₁ after Willis & Gabel, 2001) across a laterally extensive, planar erosion surface (}
Figure 3.7). This facies succession has a thickness between 8 and 18 m, with an average of 13 m. The facies succession is present over the whole extent of the study area.
Figure 3.7, 3.8, and also in other parts of the Book Cliffs outcrop belt, where abrupt fining at the top of this facies succession marks a regionally correlative flooding surface (Van Wagoner, 1991; Willis and Gabel, 2003).

It is interpreted that the facies succession of FA1, FA2 and FA4 is a genetic succession. The transition from FA2 to FA4 is attributed to an upward shallowing in palaeo-water depth from below to above the maximum depth to which currents scoured the sea bed when spring tides, storms and high fluvial discharge coincided to form the highest flow velocities. This scour depth is likely to have varied over different temporal scales, ranging from tidal cycles through to longer term periods, possibly seasonal, that could reflect periods when tidal currents were augmented by storm and river flood events. Spatial variations also occur due to fluctuations in the relative strength of tidal currents relative to wave-induced and river currents (Dalrymple and Choi, 2007; Legler et al., 2014). Alternatively, the erosion surface may record erosion during lowering of relative sea-level and forced regression (Willis and Gabel, 2001). The top of the facies succession is marked by an abrupt upward fining across a sharp surface.

3.5.3 Tide-dominated successions $S_2$ and $S_3$

The upper part of the studied interval (sandstones $S_2$ and $S_3$ after Willis & Gabel, 2001;
Figure 3.7, 3.8) comprises alternations of FA3 and FA4, with FA5 also present locally. The architecture of these strata is complex, and varies laterally from east (palaeoseaward) to west (palaeolandward) ( 
Figure 3.7, 3.8), as described in the next section of the paper. Multiple upward-coarsening successions of variable thickness (1-10 m) consist of FA3 overlain by FA4 (
Figure 3.7). Each succession is characterized by: (1) an increase in grain size from very fine- to fine-grained sandstone, (2) an upward decrease in mudstone content, (3) an upward increase in sandstone bed thickness, and (4) an increase in the size of sedimentary structures, from centimetre-scale ripple cross-lamination to decimetre-scale dune cross-bedding. The tide-dominated deposits of $S_2$ and $S_3$ are 18-32 m thick, with an average of 25 m. A prominent sandstone ($S_3$) with a thickness of 1-15 m forms the uppermost part of the lower Sego Sandstone over most of the study area.
Since both FA3 and FA4 were deposited principally by tidal currents, this part of the succession is considered to be tide-dominated. The presence of coal clasts and carbonaceous drapes may indicate erosion and transport on the lower delta plain during high river discharge, and therefore the indirect influence of fluvial processes. It is interpreted that the deposits of FA3 and FA4 are deposited in the sub-tidal zone because no indications for subaerial exposure are observed (Van Straaten, 1961; Allen, 1980; Visser, 1980). The deposits of FA5 are interpreted to represent abandoned distributary channels in the inter-tidal zone, such that the presence of FA5 indicates an upward shallowing in palaeo-water depth (Van Straaten and Kuenen, 1957; Oomkens, 1974; Staub and Gastaldo, 2003). Also, the upper sandstone (S₃) has a high relief erosional surface at its base, which is interpreted to have been cut by distributary channels (as discussed in a later section). Distributary channels mark the shallowest palaeo-water depth in delta front successions (Li and Bhattacharya, 2014). Therefore the whole succession of the lower Sego Sandstone is interpreted to be a generally upward-shallowing succession from offshore mudstones to the inter-tidal zone. The absence of any evidence of subaerial exposure can be explained by deposition during a forced regression (Posamentier and Morris, 2000; Prince and Burgess, 2013) and/or by transgressive erosion (ravinement) at the base of the Anchor Mine Tongue and/or simply by interpreting that the facies belt with supra-tidal marsh deposits did not prograde as far as the study area, and that at this location only sub- and inter-tidal palaeo water-depths were present such supra-tidal marsh deposits were not deposited in the first place.

3.5.4 Marine Anchor Mine Tongue

Marine mudstones (FA1) of the Anchor Mine Tongue define the top of the lower Sego Sandstone in the eastern part of the study area (east of Horse Mesa; Figure 3.1C, 3.8). Farther west, the Anchor Mine Tongue passes into hummocky cross-stratified and intensely bioturbated sandstones (FA2 and FA4). The Anchor Mine Tongue records an increase in water depth across a regionally extensive
flooding surface (Van Wagoner, 1991; Willis, 2000; Willis and Gabel, 2001; Hettinger and Kirschbaum, 2002; Aschoff and Steel, 2011a; Legler et al., 2014).
Correlative surfaces that can be traced across the study area are shown between the two logs, and delimit upward-coarsening successions (named according to the nomenclature of Willis & Gabel, 2001; names are encircled). The surfaces at the top of $S_1$ and $S_3$ have been interpreted as flooding surfaces, and the surfaces at the top of $S_2$ and $S_0$ as abandonment surfaces (see text). The base of the Anchor Mine Tongue (AMT) at the top of the lower Sego Sandstone is used as a datum. The arrows indicate grain size trends. For explanation of the colours representing the facies associations (FA), see
Figure 3.8.
Figure 3.8 (previous page). Correlation panel through stratigraphic logs in the study area, from west (palaeolandward) to east (palaeoseaward) (Figure 3.1B, C), showing detailed facies architecture in the lower Sego Sandstone. The base of the Anchor Mine Tongue and coincident top of the lower Sego Sandstone (Figure 3.2) is used as a datum. Key stratigraphic surfaces that have been mapped over the study area, or whose correlation is inferred are shown. See text for details.
3.6 Stratigraphic architecture

The vertical facies successions described above can be traced laterally for kilometres to tens of kilometres throughout the study area ( 
Figure 3.8). In some of the cliff-face amphitheatres, reconstructions of stratigraphic architecture and facies-association distributions over several hundreds of metres have been documented (Figure 3.9). Palaeocurrent data from the various stratigraphic intervals, and their respective facies successions, provide extensive documentation of sediment transport directions along large tracts of the cliff-face outcrop belt (Figure 3.10).

3.6.1 Wave-dominated successions $S_1$ and $S_0$

Upward-coarsening successions $S_1$ and $S_0$ exhibit minor lateral changes in thickness and have a uniform internal facies composition and relatively tabular geometries. Succession $S_1$ has a thickness of at least 14 m. Hummocky cross-stratified, lower shoreface sandstones (FA2) of succession $S_1$ are 2-3 m thick in the central part of the study area (logs 10-13, around Horse Mesa; Figure 3.1C, 3.8), and pinch out laterally towards the west and east. Succession $S_0$ is 3-6 m thick, of which the upper 1-2 m comprises lower shoreface sandstones (FA2) where it overlies succession $S_1$ in the Horse Mesa area (logs 10-13,
Figure 3.8). Towards the west, the thickness of succession $S_0$ decreases to 3 m and the lower shoreface sandstones pinch out. Towards the east, where succession $S_1$ is absent, succession $S_0$ thickens to more than 18 m, with lower shoreface sandstones comprising the upper 9 m.

The regional coastline was oriented NE to SW (Willis and Gabel, 2001; Wood, 2004; Burton et al., 2016). Therefore it is interpreted that succession $S_1$ represents southeastwards progradation. The eastern (palaeoseaward) limit of lower shoreface sandstones (FA2) lies just east of Horse Mesa (log 13,
During subsequent southeastward progradation, which resulted in deposition of succession $S_0$, lower shoreface sandstones (FA2) extended across the entire study area. The thickening of succession $S_0$ on the eastern flank of laterally discontinuous lower shoreface sandstones (FA2) in succession $S_1$ (only present in the western part of the study area,}
Figure 3.8) implies infilling of depositional topography developed at the top of succession $S_2$. These thickness patterns imply lateral shifting of depocentres between successions $S_1$ and $S_0$, which may reflect compensational stacking of delta lobes, providing the correlations of $S_0$ and $S_1$ across non-exposed areas is correct.

3.6.2 Mixed, wave- and tide-influenced succession $S_1$

Upward-coarsening succession $S_1$ is 8-18 m thick in the study area, averaging 13 m (
Figure 3.8). The facies-association components display irregular lateral variations in thickness, but average 5 m for FA1, 5 m for FA2 and 4 m for FA4 (
Figure 3.8). These three FAs are interpreted to form one genetic upward-coarsening succession of mixed wave- and tide-influenced deposition. Palaeocurrent directions derived from cross-bedding in FA4 are multi-directional, with a dominant direction towards the NW and subordinate directions towards the SE and NE (Figure 3.10E). In the context of the NE-SW-oriented palaeoshoreline (Figure 3.1B; Willis & Gabel, 2003; Wood, 2004; Burton et al., 2016), the dominant NW-ward direction is attributed to flood-tidal currents, and the subordinate SE-ward direction to fluvial and ebb-tidal currents. NE-directed palaeocurrents may be attributed to local wave-driven longshore drift, which could have deflected deltaic distributary and tidal channels, although the regional longshore drift direction in the older Blackhaw Formation was directed towards the SW (Slingerland and Keen, 1999; Hampson et al., 2014a). Alternatively, the NE-directed palaeocurrents can be attributed to dunes that obliquely over the complex morphology of nearshore tidal sand ridge (Houbolt, 1968). Intense bioturbation and abruptly decreasing grain size at the top of succession S1.
Figure 3.7, 3.8) are attributed to a pronounced decrease in sedimentation rate, probably related to an increased accommodation space generation (rising relative sea-level) because the same conditions occur at the top of succession S1 to the east of the study area (Figure 3.1; Van Wagoner, 1991; Willis & Gabel, 2003), which supports this being a regional flooding surface.

3.6.3 Tide-dominated succession S2

There is an overall upward-coarsening trend from the top of succession S1 towards the top of succession S3. This overall upward-coarsening trend is constructed via the vertical stacking of multiple thin (1-10 m) upward-coarsening successions of tidal sandstones (FA3, FA4). In the east of the study area (between logs 23 and 29, Floy Wash and Crescent Canyon in Figure 3.1C), four such upward-coarsening successions are generally present, although the number varies from three to seven (
Figure 3.8, 3.9B). Within upward-coarsening successions there is lateral variability in mudstone content between FA3 (10-50% mudstone) and FA4 (0-10% mudstone) (Figure 3.9B). A typical succession is characterised in its lower part by upward increases in grain size, sandstone content and sandstone bed thickness in FA3.
Figure 3.7). The upper parts of these successions are typically capped by low-relief, nearly planar erosional surfaces overlain by blocky, cross-bedded sandstones of FA4.
Figure 3.7). Further west, between Floy Wash and Horse Mesa (between logs 10 and 23 in Figure 3.1C), the upward-coarsening successions display uniform thicknesses (5-15 m), tabular geometries and greater lateral continuity.
Palaeocurrent directions are variable and lack clear trends (Figure 3.10B-D). Inclined surfaces (generally with a dip between 5° and up to 10°) within units of cross-beded sandstones (FA4) are observed to dip towards the NE and SW in locations where a 3D view in the cliffs is possible, while cross stratification shows SE and NW directed palaeocurrents, which suggests that dune-scale bedforms migrated over bars which accreted laterally between channels. These inclined surfaces have been observed in large cliffs (100’s of m’s wide) where such subtle inclined surfaces can be traced out. The orientation of these surfaces can be determined in amphitheatres where two or three apparent dip directions can be combined to reconstruct the true orientation of these planes. The dimensions of these bars can be roughly reconstructed from the dimensions of the sandstone (FA4) bodies in which they form.
Figure 3.8). The orientation of the bars is hard to reconstruct because of the two dimensional outcrop belt, however, it can be assumed that the bars were elongated in a dimension perpendicular to the dip of the inclined planes (Houbolt, 1968), which is a NW-SE elongation. This is consistent with the orientation of sandstone bodies of the Sego Sandstone in the subsurface (Wood, 2004; Burton et al., 2016). Channelised mudstones (FA5) with a thickness of 1-5 m and a width of less than 1 km wide are present in the upper half of this succession, except for two examples in the lower half of this succession in stratigraphic log 20. The channelised mudstones (FA5) are also more common in the western (more proximal) part of the outcrop belt between stratigraphic logs 1 and 10, although they are present further east. In locations where individual channels can be traced along curved cliff lines, it is observed that the long axes of the channels are oriented roughly NW-SE
Figure 3.8, 3.9A, C, 3.10G-I).

3.6.4 Tide-dominated succession S3

Channelised, cross-bedded sandstones (FA4) dominate interval S3 to the north and north-west of Horse Mesa (between logs 10 and 20 in Figure 3.1C), where the basal erosion surface has up to 10 m of relief (Figure 3.9A). The largest channelised sandstone body in this stratigraphic interval is up to 15 m thick. The tortuous outline of the outcrop belt north of Horse Mesa has enabled four WNW-ESE-trending channels to be mapped (red lines numbered C1-C4 in Figure 3.10F). Palaeocurrents in these channelised sandbodies have a dominant SE-ward (fluvial and/or ebb-tidal) direction and a subordinate NW-ward (flood-tidal) direction.

Figure 3.9 (next page). Uninterpreted photo-panoramas (top) and interpreted line drawings (bottom) of selected stratigraphic architectures and facies distributions exposed in cliff-face amphitheatres (located in Figure 3.1C, 3.8). (A) High-relief erosional surface at the base of succession S3. On the left (north), the erosional surface cuts out a channelised mudstone (FA5). (B) Facies architecture in the eastern part of the study area. Erosion surfaces at the base of cross-bedded sandstones (FA4) in successions S1, S2 and S3 have low relief, and there are lateral changes in sandstone content (between FA3 and FA4) within succession S2. (C) Thick channelised mudstone on the left (white arrows) and inclined surfaces in the prominent sandstone of FA4 in S2 (black arrows). AMT = Anchor Mine Tongue (of the Mancos Shale); US = upper Sego Sandstone.
Figure 3.10. Maps of (A-E) palaeocurrent distributions and (F-J) facies-association distributions for five stratigraphic levels.

(1)
Figure 3.8: (A, F) maximum regression in succession S3; (B, G) upper part of succession S2; (C, H) middle part of succession S2; (D, I) lower part of succession S2; and (E, J) maximum regression in succession S1. These stratigraphic surfaces are indicated in
3.7 Discussion

3.7.1 Depositional model for the proximal, lower Sego Sandstone

It is interpreted that the lower Sego Sandstone in the study area was deposited in a net-regressive, mixed tide- and wave- influenced deltaic environment. The upward-coarsening succession of FA1, FA2 and FA4 in interval S₁, which gradually overlie marine mudstones of the Buck Tongue, is interpreted to represent upward shallowing and regression of genetically-linked facies tracts on a wave-dominated delta front. Marine mudstones of the Mancos Shale are present in depositional down-dip locations (southeast), and the tide-influenced marginal marine mudstones and sandstones of the middle Castlegate Sandstones, which were deposited in lower delta plain settings are present in depositional up-dip (northwest) locations. This indicates that the sediments of the Sego Sandstone were deposited in coastal to marginal marine settings.

The abrupt change in grain-size across the planar erosion surfaces between FA2 and FA4 may record sediment bypass during erosion at the base of mobile, lateral migrating tidal channels. The lateral continuity of the sandstone-rich and flood-current dominated interval S₁ indicates that mud was removed from the proximal delta front. This was probably caused by a combination of wave reworking (Legler et al., 2014) and strong tidal currents seaward of the turbidity maximum zone (Dalrymple and Choi, 2007). The wave reworking of the delta front is demonstrated by the prevalence of HCS (FA2), and provides a mechanism for transporting sand and mud alongshore into the mouths of inactive distributary channels (estuaries)(Figure 3.12B).

In intervals S₂ and S₃, FA3, FA4 and FA5 interfinger laterally and vertically to form a tide-dominated facies-association complex that overlays the flooding surface above S₁. Tidal currents responsible for the cross-bedded sandstones (FA4) were concentrated through approximately coast-normal channel belts. Lateral migration of these channels, and their intimately associated tide-influenced mid-channel
bars, are interpreted to have resulted in the development of extensive, low-relief scour surfaces. Active and abandoned distributary channels are defined by sharp-based channelised successions of FA4 and FA5, respectively.

Palaeocurrent data from the tide-dominated deposits of FA3, FA4 and FA5 within interval S1 display mainly onshore (flood-tidal) directions with subordinate offshore (ebb-tidal and/or fluvial) directions (Figure 3.10E). Palaeocurrent directions become more variable in interval S2 (Figure 3.10B-D). However, in the channelised sandstones in the upper part of interval S3 there is a dominant offshore-directed (ebb-tidal and/or fluvial) trend, with a subordinate onshore-directed (flood-tidal) trend (Figure 3.10A). The cross-bedded sandstones (FA4) with offshore-directed palaeocurrents that infill the high relief, but relatively narrow, channels in the upper part of interval S3 are interpreted as fluvial-dominated, tide-influenced distributary channels. The S2 interval displays more variable palaeocurrent patterns, although there is a higher proportion of onshore (flood-tide)-directed flows. Interval S1 is dominated by onshore-directed palaeocurrents, implying local dominance of flood-tides. Flood-tidal currents that enter shallow estuarine channels from the ocean are subject to less friction, and thus attain higher velocities, than ebb-tidal currents. Therefore, in transgressive estuarine systems without a major contribution of river discharge to the ebb tide, the higher velocity flood tide transports much more sediment than the ebb tide, such that flood-directed palaeocurrent directions are likely to dominate in such settings (Dronkers, 1986; Van der Spek, 1997; Johnson et al., 1982). The lateral variability in palaeocurrent directions in the lower Sego Sandstone (Figure 3.10A-E) can also be attributed to the presence of mutually evasive flood- and ebb-tidal channels. Mutually evasive channels are characteristic of tide-influenced coastlines (Van Veen et al., 2005; Harris, 1988; Dalrymple et al., 1990; Ahnert, 1960). These mutually evasive channels can be recognized in the ancient with the preservation of opposed current directions within separate tidal channel sandbodies, each displaying a superficially unidirectional trend (e.g. Legler et al., 2013). The presence of mutually evasive tidal channels can skew preserved palaeocurrent trends, especially in areas of limited outcrop distribution.
However, the laterally extensive nature of the Sego Sandstone outcrops enables greater confidence to be placed in the depositional significance of the palaeocurrent trends.

Overall, the Sego Sandstone succession seems to preserve a temporal change that, from bottom to top, records an overall, net regressive, distal to proximal change: (1) onshore-directed currents dominated in estuary/distributary mouth settings ($S_1$), (2) bidirectional currents characterized intermediate/lower delta plain settings ($S_2$), and (3) offshore-directed currents ($S_3$) recorded the combined dominance of fluvial and ebb-tidal currents in proximal estuarine/upper delta plain locations. This is consistent with models developed for tide-dominated estuaries and deltas (Dalrymple et al., 1992; Dalrymple and Choi, 2007; Salahuddin and Lambiase, 2013; Harris et al., 2004; Dalrymple et al., 2003).

The occurrence and distribution of mud-rich facies (cross-laminated heterolithic sandstones of FA3 and mudstones and muddy heterolithic of FA5) may reflect two alternative, but not mutually exclusive, mechanisms (Figure 3.11). Firstly, the preferred model for the distribution of the mud-rich facies is that they represent the infill of the following types of channel: (1) abandoned channels, (2) less active, relatively low-energy distributary channels, or (3) smaller, laterally-migrating tidal channels without any updip fluvial connection (i.e. tidal channels that drain lower delta plain/outer estuarine areas, including intertidal areas). In this model, FA3 may have been deposited in relatively medial locations, near the widened, funnel-shaped mouths of less active and/or smaller channels in which sands of FA4 were deposited (Figure 3.12B). These seaward-connected channels will behave more like estuaries and hence will receive, in part, a seaward-derived, landward-oriented sediment flux. Such units of FA3 are likely to fine in a palaeolandwards direction, becoming more mud-rich as they approach intertidal areas and/or gradually grading into narrower channel-shaped deposits of FA5. The abundance of organic material in FA5 supports close proximity to a vegetated lower delta plain. This, and the similar stratigraphic position to the distributary channels of FA4, suggests that the deposits of FA5 were
deposited in the lower delta plain. Inactive distributaries behave like estuaries, in which flood-tidal currents can transport marine-derived sediment onshore (Dalrymple et al., 1992).

The second mechanism to explain the distribution of the mud-rich intervals results from the development of a turbidity maximum zone. This model is often applied in other studies, but cannot convincingly explain the observations made in the lower Sego Sandstone. The turbidity maximum zone occurs where the suspension load of fresh water rivers mixes with saline waters, which can result in clay flocculation. It has been hypothesized that in tide-dominated river deltas, the turbidity maximum zone occurs further seaward than in estuaries, mainly because fluvial currents are stronger relative to tidal currents in river deltas compared to estuaries (Dalrymple and Choi, 2007). More sand-rich deposits are proposed to occur both landward and seaward of the turbidity maximum zone (Figure 3.12C). Sand-rich deposits on the landward side of the turbidity maximum zone are dominated by offshore-directed fluvial currents. In contrast, sand-rich deposits on the seaward side are characterised by more bidirectional palaeocurrent patterns, in places dominated by onshore-directed, flood-tidal currents (Dalrymple and Choi, 2007). This partly resembles the distribution of mudstone that is present in the Sego Sandstone. Hence, cross-bedded sandstones (FA4) with onshore-directed palaeocurrents in interval S1, which prograded over offshore mudstones and shoreface sandstones (FA1 and FA2), could represent deposits located seaward of the turbidity maximum zone. During continued progradation, relatively mudstone-rich deposits with bidirectional palaeocurrents (interval S2) may represent deposition within the turbidity maximum zone. Subsequently, (interval S3) cross-bedded sandstones (FA4) with offshore-directed palaeocurrents, could comprise fluvial-dominated distributary channels located landward of the turbidity maximum zone (Figure 3.12C). The absence of cross-beds formed by dunes in FA3, which is most abundant in the relatively mudstone-rich interval S2, is problematic for the interpretation of this interval as the preserved turbidity maximum zone, because such cross-beds were present in deposits (FA4) interpreted to form landward and seaward of the turbidity maximum zone, while it is expected that bedforms of uniform size are present along the axis of a single distributary channel (Dalrymple and Choi, 2007).
3.7.2 Analogues

The first depositional model outlined above, in which sandstones were deposited in active distributaries and relatively muddy deposits are more common in abandoned distributaries, is similar to inactive delta-plain regions of the present day Ganges-Brahmaputra and Mahakam river deltas (Figure 3.12A). Both of these modern deltas are tide-influenced and contain regions with active distributaries separated by abandoned regions of the lower delta plain in which inactive distributary channels have become estuaries (Allison et al., 2003; Storms et al., 2005; Salahuddin and Lambiase, 2013). The tide-dominated Fly and Yangtze river deltas are potential modern analogues for the second depositional model, in which mud is deposited in the turbidity maximum zone (Figure 3.12A). Both the Fly and Yangtze deltas have an overall funnel shape, such that distributary channel geometries and orientations are similar to those in tide-dominated estuaries. The analogy of the lower Sego Sandstone with the Fly and Yangtze river deltas is another argument for the weakness of the model in which the mud-rich facies are deposited in the turbidity maximum zone because the modern analogues are situated in a confined setting, while the lower Sego Sandstone consists of two sandstone tongues which extend seaward of their intervening coastline (Willis and Gabel, 2003; Young, 1955). The modern analogues cited above provide two end-member mechanisms to explain the relative distributions of mudstone and sandstone in the lower Sego Sandstone. Both the Sego Sandstone and the modern analogues are dominated by a grain size mix of mud and fine-grained sand (Orton and Reading, 1993), so similar hydrodynamic conditions and geomorphologies are expected. The modern analogues are all tide-dominated deltas developed under a tropical climate, with high suspended sediment loads and containing mangrove-dominated flora, which facilitates trapping of fine-grained sediment on the lower delta plain. In contrast, the lower Sego Sandstone was deposited under a temperate climate with hot summers and cool winters (Sellwood and Valdes, 2006), is dominated by bedload-produced structures, and lacked salt-tolerant mangroves, which did not evolve until the latest Cretaceous (Greb et al., 2006).

Another shortcoming of the Fly, Mahakam and Yangtze river delta analogues is the absence of significant wave-influence, which contrasts with the abundant storm-induced, HCS-bearing sandstone
beds in the delta front deposits of the lower Sego Sandstone (Van Wagoner, 1991; Willis and Gabel, 2001; Legler et al., 2014). Consequently, the latter aspect of the “lower Sego Delta” has been used to infer mixed tide- and wave-influence (e.g. Legler et al., 2014), rather than being tide-dominated (cf. Willis and Gabel, 2001). Tide-dominated and wave-dominated areas can also coexist laterally in modern deltas, such as in the modern Orinoco delta (Warne et al., 2002; Aslan et al., 2003). Alternatively, deltas can also change their process character as they evolve with time. For example, tide- and/or fluvial-dominated deltas often characterise broad, shallow shelves, which can evolve into wave-influenced deltas as they prograde across narrow shelves towards deeper water (Yoshida et al., 2007; Ainsworth et al., 2008). Examples of the latter are the modern Mekong delta (Ta et al., 2002) and the Pliocene Orinoco delta (Chen et al., 2014; Bowman and Johnson, 2014). The “lower Sego Delta” can be interpreted in the light of this latter analogue. The higher abundance of wave-reworked facies in the more distal parts of the lower Sego Sandstone (Legler et al., 2014) suggests that storm-wave processes became gradually more dominant over time during progradation. In contrast, it can be argued that storm-wave produced facies (FA2) of the lower Sego Sandstone are preferentially preserved in more distal locations, below fairweather-wave base, and tide-influenced facies (FA3-5) in more proximal locations, above fairweather-wave base (Legler et al., 2014). The Niger delta has a similar distribution of depositional processes with a wave-dominated shoreline (Allen, 1965b; Allen, 1964; Allen, 1970) and a tide-dominated lower delta plain where channel-fills and bars amalgamate laterally to form laterally continuous tide-dominated sand sheets (Oomkens, 1974). However, the distributary channels of the Niger are narrow, and hence, distinct mouth bars form at their mouths (Allen, 1964; Allen, 1965b) while no mouth bar deposits are recognized in the lower Sego Sandstone. However, with the evidence currently available, it cannot be resolved whether the higher abundance of wave-dominated structures (FA2) in distal parts of the lower Sego Sandstone is the result of a temporal increase of wave-dominance or the preferential preservation of wave-dominated structures at the palaeo-coastline.
Figure 3.12B shows the envisioned depositional model for the lower Sego Sandstone, taking elements of the modern analogues which best fit the facies and architectural characteristics of the lower Sego Sandstone. The modern Niger and Mekong deltas form a good analogue for the distribution of a wave-dominated delta front and a tide-dominated lower delta plain. The Mahakam delta is taken as an analogue for the lateral distribution of active distributary channels and abandoned parts of the delta plain, where former distributary channels have become estuaries. The morphology of the river mouths of the Mekong and Fly rivers can be taken as an analogue for the wide river mouths without fluvial mouth bars which are envisioned for the depositional model of the lower Sego Sandstone.
Figure 3.11. Two depositional models for the Sego Sandstone. (A) Mudstone content (FA3) varies down-dip forming a down-dip sand (FA4, high relief channels)- mud (FA3)- sand (FA4 low relief and FA2)- mud (FA1) successions following the model of (Dalrymple and Choi, 2007). (B) Mudstone content varies laterally with sandstones (FA4) deposited in active distributary channels and mudstones (FA3 and FA5) deposited in abandoned distributary channels. The latter model (B) is preferred because in model (A) it is expected that the mudstone interval has the same flow conditions as the sandstone intervals up-dip and down-dip. However, bedform size in FA3 does not confirm this.
Figure 3.12. (A) The morphologies of selected modern analogues for the lower Sego Sandstone discussed in the text: the Niger (Allen, 1965b; Allen, 1970), Mekong (Nguyen et al., 2000), Mahakam (Allen and Chambers, 1998; Salahuddin and Lambiase, 2013) and Fly River (Dalrymple et al., 2003; Harris et al., 2004) deltas. The outline of the Sego Sandstone outcrop belt is shown at the same scale as detailed maps of the Niger and Mekong deltas, while the Fly and Mahakam deltas are shown in outline at the same scale (black and white) and are enlarged in detailed maps. (B) Proposed model for the lower Sego Sandstone. Active distributaries show a downstream increase in (flood) tidal influence. Abandoned distributaries become estuaries and show a landward fining trend. See text for further discussion. (C) Conceptual model for the distribution of facies in tide-dominated deltas based in part on the Fly River delta (modified after Dalrymple and Choi, 2007).
3.8 Conclusions

The lower Sego Sandstone is a net-regressive deltaic unit that exhibits mixed tide and wave influence. Deposits in the western (proximal) part of its outcrop belt, where fluvial and tidal processes interacted, have not been previously studied in detail, and consist of five facies associations. A complete idealized regressive succession consists of bioturbated mudstones containing planar, lenticular and wavy beds of siltstone and very fine-grained sandstone (FA1), which pass gradationally upwards into hummocky cross-stratified, very fine-grained sandstones (FA2), which in turn are overlain across a sharp-to-erosional surface by ripple cross-laminated and trough cross-bedded, sparsely bioturbated, fine-grained sandstones containing mudstone drapes (FA3, FA4). This succession is interpreted to record offshore deposits (FA1) overlain by storm-dominated, lower shoreface deposits on the distal delta front (FA2) and a tide-dominated facies-association complex on the proximal delta front and lower delta plain (FA3, FA4). Channelised mudstones and heterolithic mudstones (FA5) occurring towards the top of the lower Sego Sandstone are interpreted as abandoned distributary channel fills.

Up to four regressive successions are interpreted in the proximal, lower Sego sandstone (S₁, S₀, S₁ and S₂-S₃, following and extending the nomenclature of earlier workers). The first and second successions (S₁, S₀) each contain only offshore and distal delta front deposits (FA1, FA2), and are capped by offshore deposits (FA1) across an abandonment surface. These successions represent only the palaeoseaward parts of regressive tongues. The third succession (S₁) consists of offshore and lower shoreface deposits (FA1, FA2) and proximal delta front to lower delta plain deposits (FA4), in which palaeocurrents are directed predominantly onshore. Tide-dominated proximal delta front to lower delta plain deposits (FA4) are interpreted as the deposits of laterally migrating tide-influenced mid-channel bars that infilled shallow, low-relief channels in which flood tidal currents were dominant. The fourth succession (S₂-S₃) contains only proximal delta front to lower delta plain deposits (FA3 and FA4), including a small proportion of channelised mudstones and muddy heterolithics that are interpreted to infill abandoned distributaries (FA5). Ripple cross-laminated, heterolithic sandstones (FA3) with a relatively high mudstone content and containing lenticular, wavy and flaser bedding with bidirectional
palaeocurrent directions, are common in this succession. The upper part of this succession (S$_3$) comprises cross-beded sandstones (FA4) with predominantly offshore-directed palaeocurrents that overlie high-relief channelised erosion surfaces, and which are interpreted as active distributary channels in which fluvial and/or ebb-tidal currents were dominant. Laterally extensive (>10 km) abandonment surfaces are absent in the tide-dominated deposits of the S$_2$-S$_3$ succession, and its subdivision is not straightforward.

The distribution of facies associations and mudstones in the proximal delta front to lower delta plain deposits (FA3, FA4, FA5) in the S$_1$, S$_2$ and S$_3$ intervals are explained by the avulsion and abandonment of active distributary channels. The relatively muddy strata of S$_2$ interval represent an abundance of abandoned distributary channels, which effectively became tide-dominated estuaries during abandonment, while the relatively sandy strata of the S$_3$ interval mainly contain the deposits of active fluvio-tidal distributaries and the S$_1$ interval of marine reworked sand. This model emphasises localised, autocyclic abandonment of parts of the delta, and is comparable to aspects of the modern Ganges-Brahmaputra and Mahakam river deltas.
Chapter 4. Depositional evolution of a progradational to aggradational, mixed-influenced deltaic succession: Jurassic Tofte and Ile formations, southern Halten Terrace, offshore Norway

4.1 Introduction

Coastal-deltaic depositional systems are commonly classified in terms of the relative interaction of wave, tide and fluvial processes (Galloway, 1975; Boyd et al., 1992). The facies characteristics of the three end members of this classification scheme can be defined with a reasonable level of confidence, supported by well described modern depositional systems (e.g. Penland et al., 1988; Dalrymple et al., 2003; Fanget et al., 2014). However, many ancient shallow-marine deposits display evidence of mixed process regimes in which two or more processes were significant (i.e. they preserve mixed-energy depositional systems), which are more problematic to classify. Defining the relative interaction of processes in a precise way is challenging but also central to developing predictive process-oriented interpretations of coastal-deltaic reservoirs (Ainsworth et al., 2011).

The physical depositional processes controlling the character of clastic coastal deposits vary both spatially and temporally (e.g. Boyd et al., 1992; Ainsworth et al., 2011; Olariu, 2014), and these variations are reflected in vertical and lateral facies changes in paralic depositional systems. The spatial and temporal evolution of coastal systems in especially complex in tectonically active basins such as rift basins, because creation of accommodation space, creation of sediment source areas, sediment routing and basin physiology are affected by tectonic activity (Ravnås and Steel, 1998). Understanding of the evolution of coastal systems is important in hydrocarbon exploration and production in order to predict the presence and quality of hydrocarbon reservoirs. A facies-based approach is applied herein to an evaluation of the Lower to Middle Jurassic Ror, Tofte and Ile formations of the southern Halten Terrace, offshore mid-Norway, which together constitute a net progradational to aggradational, mixed-influenced, coastal-deltaic succession, based on a particularly rich and high-quality subsurface
database (Ravnås et al., 2014). The succession hosts significant oil and gas resources on the Halten
Terrace (e.g. Alve, Åsgard, Heidrun, Kristin, Mikkel, Morvin, Njord, Norne, Skalv, Skuld and Urd fields;
Harris, 1989; McIlroy, 2004; Morton et al., 2009; NPD, 2016), but reservoir distribution and character
are difficult to predict away from limited well penetrations in the southern Halten Terrace.

The overall distribution and structural setting of these formations are well known based on over 30
years of exploration evaluation and field development studies (Gjelberg et al., 1987; Blystad et al.,
1995; Ravnås et al., 2014). However, detailed sedimentological analysis has been restricted to
relatively small areas in the northern part of the Halten Terrace (Harris, 1989; McIlroy, 2004). The gross
depositional setting of the Tofte and Ile formations is one of a progradational-to-aggradational clastic
wedge that passes basinwards into marine mudstones (Gjelberg et al., 1987; Ravnås et al., 2014). Such
clastic wedges may be either (1) externally forced by temporal variations in the balance between
relative sea-level and sediment supply (the “A/S ratio” sensu Muto and Steel, 1997) or, (2) record the
internal response of the depositional system to steady external forcing (autoretreat sensu Muto and
Steel, 1992; Muto 2001; Muto et al., 2007). The aims of this paper are: (1) to evaluate the spatial and
temporal variations in the process regime (i.e. the relative degree of wave, tide and fluvial influence)
of the Tofte and Ile formations; and (2) to develop a sedimentological and stratigraphic model that
accounts for stratal architecture, grain-size partitioning (mass balance) (Strong et al., 2005; Bijkerk et
al., 2016) and variations in depositional facies and process regime (Olariu, 2014), and (3) provide a
framework for future prediction of sandstone reservoir distribution and character in the southern
Halten Terrace.

4.2 Geological setting

The Halten Terrace is a N-S trending, 200 km long and 100 km wide, rhomboid shaped fault-bounded
terrace offshore mid-Norway (Figure 4.1A,B). It is bounded by prominent, westward dipping Mesozoic
extensional faults: (1) the Bremstein Fault to the east, (2) the Vingleia Fault to the southeast, (3) the
Klakk Fault to the west, (4) a series of en echelon faults that make up the Revfallet Fault Complex to
the north, and (5) the Kya Fault within the Halten Terrace (Figure 4.1B) (Blystad et al., 1995). These faults and the various bounding structural elements are the result of Mesozoic rifting between NW Europe and Greenland (Doré, 1991; Doré, 1992; Brekke et al., 2001).

The normal faults on the Halten Terrace ranges were intermittently active between the Permian and the early Cretaceous, with temporal and spatial variations in the rate and amount of extensional displacement (Figure 4.2). The Bremstein Fault, which separates the Halten Terrace from the Trøndelag Platform (Figure 4.1B), and the Revfallet Fault Complex (outside of the study area in the north) were both active during middle Jurassic (Garn Formation) to earliest Cretaceous (Spekk Formation) times (Corfield and Sharp, 2000; Richardson et al., 2005; Marsh et al., 2010; Wilson et al., 2013). In the northwestern part of the Halten Terrace (outside the study area), the Smørbukk and Trestakk faults were active between the late Triassic (Åre Formation) and earliest Cretaceous (Spekk Formation).

Upper Triassic to Lower Jurassic strata (Grey Beds, Åre, Tilje, Ror, Tofte and Ile formations; Figure 4.2), display subtle, fault-related thickness changes, whereas younger strata (Garn, Melke and Spekk formations; Figure 4.2) exhibit more substantial thickness changes (Corfield and Sharp, 2000; Marsh et al., 2010).

The standard lithostratigraphic nomenclature of Dalland et al. (1988) has been adopted in this study (Figure 4.2). Triassic strata consist of continental red beds and evaporites (Grey Beds and Red Beds in Figure 4.2). Two distinct halite-bearing intervals have been recognized, which form a decollement interval for younger Mesozoic extensional faults (Jacobsen and Van Veen, 1984). During the Rhaetian to Hettangian, the Åre Formation was deposited as a series of coal-bearing alluvial and coastal plain deposits (Figure 4.2). Fluvial sandstone bodies in the upper part of the Åre Formation show evidence of tidal current action, indicating the onset of marine transgression, and are overlain by a series of retrogradationally stacked shoreface deposits (Thrana et al., 2014).

Three regressive-transgressive siliciclastic wedges define the Lower-Middle Jurassic strata of the Halten Terrace (Gjelberg et al., 1987). Ravnås et al. (2014) describe the regional distribution of the
clastic wedges and the overall depositional process regimes, which were active during their deposition.

The first (oldest) wedge (Pliensbachian to Toarcian) comprises the Tilje Formation (Figure 4.2) (Martinius et al., 2001; Martinius et al., 2005; Ichaso and Dalrymple, 2009; Ichaso and Dalrymple, 2014; Ichaso et al., 2016). The second wedge (Toarcian-Aalenian) consists of the Ror, Tofte and Ile formations (Harris, 1989; McIlroy, 2004; Ravnås et al., 2014). The third, and youngest (Bajocian-Bathonian) wedge comprises the Not and Garn formations (Figure 4.2) (Corfield et al., 2001; Messina et al., 2014).

Increased rates of rifting commenced during deposition of the third wedge (Corfield and Sharp, 2000; Richardson et al., 2005; Marsh et al., 2010; Wilson et al., 2013). This resulted in widespread marine flooding followed by prolonged deposition of offshore shales of the Melke and Spekk formations (Dalland et al., 1988; Wilson et al., 2013). This study focuses on the middle clastic wedge (Tofte and Ile formations). These deposits are relatively poorly studied, but there is clear evidence for a mix of depositional processes (Harris, 1989; McIlroy, 2004; Ravnås et al., 2014), and during an episode of the rift basin with relatively little tectonic activity (Corfield and Sharp, 2000; Richardson et al., 2005; Marsh et al., 2010). This gives a good opportunity to study the spatial and temporal evolution of depositional processes of coastal systems in rift basins.
Figure 4.1. (A) Palaeogeographic map of northern Europe and surrounding areas during the Middle Jurassic showing the context of the Halten Terrace (modified after Ziegler, 1988; Doré, 1992; Engkilde and Surlyk, 2003; Ravnås et al., 2014; Eide et al., 2016). (B) Map showing the main structural elements of the Halten Terrace (modified after Blystad et al., 1995). (C) Map of the study area showing the locations of the wells and well correlations panels used in this study (Norwegian Petroleum Directorate (NPD), 2016).
Figure 4.2. Lithostratigraphy of the Jurassic of the Halten Terrace (modified after Dalland et al., 1988; Gradstein et al., 2012; Ravnäs et al., 2014).
4.3 Dataset and Methods

This study is based on data from 34 wells through the Ile, Tofte and Ror formations in the southern Halten Terrace (Figure 4.1C; Table 4.1). Each well contains an extensive and consistent suite of wireline logs, including gamma ray (GR), neutron porosity (NPHI) and bulk density (RHOB) logs, and there is high-quality core coverage from 20 wells (1323 m of total core length; Fout! Verwijzingsbron niet gevonden.). The cores were described on a 1:50 vertical scale (5 cm resolution). Facies analysis has been carried out based on lithology, grain-size and sorting, primary sedimentary structures, bioturbation index (BI, Taylor and Goldring, 1993) and trace fossil assemblages (MacEachern and Bann, 2008). Facies have been grouped into recurring facies associations. The facies associations were calibrated with their wireline log expression, in order to determine their diagnostic wireline log expression and their reservoir properties. The studied succession was subdivided into genetically related stratigraphic units in order to facilitate correlation and reservoir mapping (e.g. thickness, net sand and facies association proportions). The deep burial (3.5-4.5 km) results in well compacted rocks with extremely well-preserved cores, which aids sedimentological observation. However, both seismic imaging and palynomorph preservation are poor, which limits their value in guiding high-resolution correlations within formations.

4.4 Facies analysis

A wide range of facies types were identified and grouped into 8 facies associations. Figure 4.3 provides an overview of the facies and 8 facies associations in graphical table form, Figure 4.4-6 show core photos that are representative of particular facies associations, and Figure 4.7 shows representative core logs for each facies association. Table 4.3 is an overview of the wireline log expression of the facies associations.
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Table 4.2. Database of cores used in this study.

Figure 4.3 (next page). Summary graphical table of facies and facies associations. Lithology abbreviations: vf.=very fine; f.=fine; m.=medium; c.=coarse; vc.=very coarse; m.=muddy; mst.=mudstone; hl.=heterolithic; s.=sandy; sst.=sandstone; p.=pebbly; cgl.=conglomerate. Trace fossil abbreviations: as.=Asterosoma; Carb. Debris=carbonaceous debris, co.=Cosmorhaphe; cr.=Cruziana; cy.=Cylindrichnus; e.t.=escape trace, gy.=Gyrochorte; he.=Helminthopsis; lo.=Lockeia; op.=Ophiomorpha; pa.=Palaeophyces; pl.=Planolites; ro.=Rosellia; sc.=Schaubcylindrichnus; sk.=Skolithos; sl.=Scolicia; te.=Teichichnus; th.=Thalassinoides; zo.=Zoophyces; WF.=wireline log facies.
### 4.4.1 FA1 Prodelta and offshore

#### 4.4.1.1 Description

FA1 is the mudstone-dominated part of coarsening upward facies successions (Figure 4.4, 4.7C, 4.8). It consists of dark, organic-rich and relatively homogeneous mudstones (facies F1a, Figure 4.4K), which pass gradationally upwards into mudstones containing centimetre-scale siltstone and very fine-grained sandstone layers and elongate, slightly asymmetrical ripples (facies F1b, Figure 4.4L). Rare asymmetrical ripples occur in the upper part of successions (facies F1d). Carbonaceous debris commonly lines individual laminae.

Bioturbation increases upwards in both intensity (from BI=0-3) and diversity. Trace fossils include *Planolites, Palaeophycus, Schaubcylindrichnus, Helminthopsis, Teichichnus, Thalassinoides, Ophiomorpha, Scolicia* and *Zoophycos* (Figure 4.4H-J). Locally the BI is higher (BI=3-4) and there is an increase in trace fossil diversity, including the trace fossils *Scolicia, Palaeophycus* and *Cosmorhaphe* (facies F1c, Figure 4.4H-J). The sand content, bioturbation index (BI=2-5) and trace fossil diversity increase upwards, with the latter including *Planolites, Palaeophycus, Schaubcylindrichnus, Helminthopsis, Thalassinoides, Ophiomorpha, Scolicia* and *Zoophycos*. FA1 is gradationally overlain by FA2.

FA1 has a distinctive wireline log expression, most notably its high and upward decreasing GR profile (70-120 API). At the base of the Ror Formation, which is not cored, very high GR values of 100-150 API
are present locally. FA1 is further characterized by high NPHI (10-30 %) and high RHOB (2.5-2.7 g/cm$^3$) values, which both decrease upward (e.g. in the Ror Formation in Figure 4.9).

4.4.1.2 Interpretation

The low-diversity trace fossil assemblage of facies F1a and F1b constitutes an impoverished distal *Cruziana* ichnofacies (MacEachern and Bann, 2008), which indicates deposition in a chemically stressed marine setting, possibly in a brackish prodelta environment where fresh and marine waters mixed (MacEachern and Bann, 2008). The mudstone-dominated lithology, low degree of bioturbation, high organic carbon content and abundant carbonaceous debris suggests sedimentation in a low-energy, stressed, dysoxic environment. The abundance of both combined flow and current ripples may be indicative of offshore sediment transport by currents associated with fluvial discharge and/or storm-wave action (Arnott and Southard, 1990; Bohacs et al., 2014; Li et al., 2015). Those intervals with higher bioturbation indices and trace fossil diversities (facies F1c and F1d) reflect deposition in a less stressed and more open marine environment with lower deposition rates (MacEachern and Bann, 2008), probably in offshore areas lateral from the main sediment entry points along a deltaic coastline (Bohacs et al., 2014; Li et al., 2015).

Figure 4.4 (next page). Panel showing photographs of facies associations FA1-FA4, which typically occur together in upward-coarsening facies successions, colour coded by facies association (Figure 4.3). For abbreviations see Figure 4.3. (A) Poorly sorted cross-bedded fluvial sandstone (facies F3b; well 6406/9-3 4730 m). (B) Very poorly sorted cross-bedded pebbly fluvial sandstone (facies F3a; well 6406/9-3 4727 m). (C) Very poorly sorted cross-bedded pebbly sandstone with *Skolithos* (facies F4a; well 6406/9-1 4617 m). (D) Two cross-bedded coarse-grained sandstones beds (indicating high fluvial discharge, facies F4b) separated by a heterolithic fine-grained sandstone bed with bioturbation (indicating tidal and marine reworking, facies F4c; well 6406/9-1 4608 m). (E) Hummocky cross-stratified very fine-grained sandstone with interbedded siltstone and sandstone lenses (facies F2a; well 6406/9-1 4784 m). (F) Current ripple cross-laminated sandstone (facies F2b; well 6406/5-1 4506 m). (G) Current ripple cross-laminated (facies F2b) and cross-bedded sandstone (facies F2d; well 6406/5-1 4507 m). (H-J) Heterolithic bioturbated offshore mudstones (facies F1d; H, well 6406/5-1 4522 m; I, well 6406/9-1 4855 m; J, well 6406/9-1 4856 m). (K) Homogenous prodelta mudstone (facies F1a; well 6406/9-1 4811 m). (L) Combined flow ripples forming very fine-grained sandstones lenses in prodelta mudstone (facies F1b; well 6406/9-1 4803 m).
4.4.2 FA2 Delta front and mouth bar

4.4.2.1 Description

FA2 gradually overlies FA1 and consists of coarsening upward facies successions of very fine- to very coarse-grained sandstones (Figure 4.4, 4.7C and 4.8). The basal part of the facies succession consists of 10-20 cm thick, very fine to fine-grained sandstone beds containing hummocky cross-stratification (HCS) (facies F2a, Figure 4.4E). This grades upward into very fine to medium-grained sandstones with current-ripple cross-lamination, with carbonaceous debris lining foreset laminae (facies F2b, Figure 4.4F). This coarsens upwards into medium- to coarse-grained, cross-bedded sandstones, often with mudstone clasts lining the toesets and foresets of cross-beds, and occasionally with mud drapes along their foresets (facies F2d, Figure 4.4G). The cross-bedded sandstones pass gradationally upwards into more poorly sorted, very coarse-grained sandstones with scattered granules (facies F2e). Individual beds fine upward, but the succession coarsens upwards overall.

The BI is low (0-2) and trace fossils include Planolites, Palaeophycus, Skolithos, Rosselia and escape traces. Locally the bioturbation is moderate in intensity (BI of 2-5) and more diverse, including Planolites, Palaeophycus, Ophiomorpha, Rosselia, Helminthopsis, Teichichnus, Schaub cylindrichnus and Thalassinoides (facies F2c).

FA2 is characterized by high to upward-decreasing GR (20-70 API) and RHOB (2.1-2.5 g/cm³) and low to upward-increasing NPHI (15-30 %) responses (e.g. in the Tofte Formation in Figure 4.9). The GR, RHOB and NPHI responses reflect the upward decrease in mudstone content and relative increase of porosity.
4.4.2.2 Interpretation

The coarsening-upward facies succession combined with abundant current ripple cross-lamination and dune-scale cross-bedding suggests upward shallowing in a subaqueous current-dominated setting (De Raaf et al., 1965; Wright, 1977; Elliott, 1978; Fielding et al., 2005a; Hampson et al., 2011). The coarse grain size and poor sorting within the highest energy facies is typical of deposition close to river mouths (Orton and Reading, 1993). The low intensity of bioturbation is consistent with high sedimentation rates in a delta front setting (MacEachern and Bann, 2008). The low diversity of trace fossil assemblages of the Cruziana and Skolithos ichnofacies also implies a stressed environment (MacEachern and Bann, 2008). This is consistent with a river mouth bar setting, where high sedimentation rates and fresh water inflow results in a stressed environment. Mud draped cross-stratification indicates that tidal currents affected depositions (Boersma, 1969).

The presence of HCS in the lowest part of the succession indicates the influence of storm waves (Dott and Bourgeois, 1982; Duke, 1985; Keen et al., 2012). The absence of HCS in shallower and coarser grained parts of the succession may reflect the relatively coarse grain size, rather than a change in depositional processes (Dott and Bourgeois, 1982; Yoshida et al., 2007). The high degree of sorting in the coarser grained facies F2b-d is probably indicative of wave reworking (Yoshida et al., 2007; Hampson et al., 2011). This analysis suggests a mixed energy depositional setting, comprising a fluvial- and wave-influenced and tide-affected delta front environment (Fwt or Wft sensu Ainsworth et al. 2011).

4.4.3 FA3 Fluvial channel

4.4.3.1 Description

An erosional contact always separates FA3 from the underlying FA, most commonly FA1 or FA2. FA3 consists of sharp-based, fining upward, medium- to very coarse-grained, poorly sorted sandstones, which locally contain granules and pebbles. A quartz pebble lag directly overlies the erosional basal contact. Subsequently, poorly sorted, cross-bedded pebbly sandstones (facies F3a, Figure 4.4B) pass
upwards into moderately sorted cross-bedded sandstones (facies F3b, Figure 4.4A). Elsewhere, there are interbedded alternations of facies F3a and F3b separated by sharp boundaries. Homogenous, structureless mudstones (c. 0.5-1 m thick), with sharp tops and bases (facies F3c), are occasionally present and occur towards the top of the fining upward successions. Bioturbation is absent.

The wireline log values of FA3 are similar to those of FA2, but the logs show low and blocky to upward-increasing GR (20-50 API) and RHOB (2.1-2.5 g/cm3) responses and high and upward-decreasing NPHI (15-30 %) values. These responses correspond to the upward-fining trend of the facies succession (Figure 4.9). The thicker homogenous mudstones (facies F3c) are visible in the wireline logs (e.g. 50-150 API GR readings) but thinner layers are below vertical resolution.

4.4.3.2 Interpretation

The erosional base, poor sorting, upward decreasing grain size trend, primary sedimentary structures and absence of bioturbation are all consistent with fluvial channel deposition (Allen, 1965a; Leeder, 1973). In places, the repeated interbedding of facies F3a and F3b may record multi-storey stacking of bars, consistent with deposition on rapidly migrating bars (Miall, 1996; Hartley et al., 2016). The intervals of homogenous mudstones with sharp tops and bases which are interbedded with sandstones of this FA are interpreted to have been deposited in low-energy abandoned channels, because facies F3c occurs exclusively in the upper part of this upward fining facies succession (i.e. they overlie intervals of facies F1a and F1b). Evidence for tidal or wave influence is absent, suggesting that FA3 represents the deposits of purely fluvial channels (F sensu Ainsworth et al. 2011).

4.4.4 FA4 Marine-influenced fluvial channel

4.4.4.1 Description

FA4 comprises fining upward successions of cross-stratified, moderate to very well sorted, fine- to very coarse-grained sandstones. These overlie a sharp basal contact, which is often lined by a lag of quartz granules, pebbles and mud clasts. The overlying cross-bedded sandstones contain mudstone laminae and mud clasts in their toesets and along their foresets, which highlight cross-bed definition (facies
F4a, Figure 4.4C). Farther upward, the deposits contain a similar suite of sedimentary facies but are moderately sorted (facies F4b, Figure 4.4D). These deposits are interbedded with mud-draped, wavy-laminated, fine-grained sandstones (facies F4c) and heterolithic mudstones containing current-ripped sandstone lenses (facies F4e). Locally, homogenous, very well sorted, medium- to coarse-grained sandstones are present (facies F4d). These sandstones have a sharp base and show a slight upward decrease in sorting, which enables cross-bedding to be observed. Homogenous mudstones with a thickness of 1-5 cm are occasionally present in this facies.

Bioturbation in the sandstones (facies F4a-d) is generally low in intensity (Bl=0-3) and trace fossils include Ophiomorpha, Skolithos, Palaeophycus and Cylindrichnus. The heterolithic deposits of F4e contain Planolites, Palaeophycus and Teichichnus. The wireline log expression is indistinguishable from that of FA3, so the identification of FA4 depends on the availability of core (Figure 4.9).

4.4.4.2 Interpretation

The sharp, erosional base and fining upward grain size trend of successions of FA4 suggests deposition in a channel. The poor sorting and coarse grain size of sandstones in the lower part of the succession indicates that texturally immature sediment was transported and deposited by strong unidirectional currents during times of high fluvial discharge (Dalrymple et al., 2015b; Gugliotta et al., 2015; Gugliotta et al., 2016; Jablonski and Dalrymple, 2016). However, there are also indicators of tidal currents, wave reworking and marine salinities: The frequency of mud drapes and the heterolithic toesets of cross-beds is indicative of bedforms formed by tidal currents (Boersma, 1969; Terwindt, 1971; Van den Berg et al., 2007). The heterolithic mudstones and sandstones (facies F4c,e), which are interbedded with coarser grained sandstones, reflect tidal current deposition during periods of low fluvial discharge (Dalrymple et al., 2015b; Gugliotta et al., 2016; Jablonski and Dalrymple, 2016). The homogenous mudstones (facies F4b and F4d) resemble fluid mudstones, indicating that wave action or tidal currents kept mud formed by salinity-induced mud flocculation in suspension (Ichaso and Dalrymple, 2009). The mature texture of well-sorted, medium- to coarse grained sandstones in facies F4d is attributed to
sorting by waves, while the absence of HCS is attributed to the coarse grain size of available sediment (Yoshida et al., 2007). Cross-bedding in facies F4d may be due to dune migration in response to a combination of unidirectional fluvial, tidal and/or wave-generated longshore currents at the river mouth. The presence of the fluid mudstone deposits in this facies is not conclusive evidence for distinguishing between tide- and wave-generated currents (Ichaso and Dalrymple, 2009), but is consistent with deposition of facies F4d in an outer estuary mouth or in a river mouth subject to wave action.

The presence of *Ophiomorpha, Skolithos, Palaeophycus* and *Cylindrichnus*, which constitute the *Skolithos* ichnofacies, in the sandstone facies (F4a-F4d) signifies some marine influence, possibly brackish salinities, within a relatively unstable and rapidly shifting sandy substrate (MacEachern and Bann, 2008). The low diversity suites of *Planolites, Palaeophycus* and *Teichichnus* burrows (*Teichichnus* ichnofacies) in heterolithic deposits (facies F4e) is consistent with deposition on bars with brackish salinities and intermittent sedimentation (e.g. as represented by the spreiten of *Teichichnus*) (Pemberton et al., 2009). In summary, FA4 represents deposition in tide influenced, wave affected fluvial channels (*Ftw* succession *sensu* Ainsworth et al. 2011), and represents the transition between fluvial (FA3) and tidal (FA5) channel end-members.

4.4.5 FAS Tidal channel

4.4.5.1 Description

FA5 occurs laterally to FA3 and FA4, and is interbedded with FA6 (Figure 4.9, 4.9). It consists of sharp-based, fining upward successions of heterolithic, very fine- to medium-grained sandstones. The base of successions of FA5 commonly comprises either a 1-5 cm-thick homogenous mudstone layer, or several thinner mudstone layers (facies F5a, Figure 4.5B,F). Bioturbation in these homogenous mudstones is low in intensity (BI=0-1) and diversity (*Palaeophycus* and *Planolites*). Overlying cross-bedded fine- to medium grained sandstones frequently contain foresets that are lined with either mud clasts or mud drapes (facies F5b, Figure 4.5B). These sandstones fine upward into very fine- to fine-
grained sandstones with apparent bidirectional cross-lamination and current ripples, which also contain mud clasts and mud drapes (facies F5c, Figure 4.5C). Bioturbation in the latter two facies is variable (BI 0-4), with trace fossils ranging from low diversity suites (*Palaeophycus* and *Planolites*) to more diverse suites that include *Lockeia*, *Gyrochorte*, *Ophiomorpha*, *Palaeophycus*, *Planolites*, *Cylindrichnus*, *Asterosoma* and escape traces. FA5 displays a variable wireline log expression. Where sandstone rich, FA5 has a similar response to FA4, and where sandstone poor it appears similar to FA6 (e.g. in the Ile Formation in Figure 4.9).

### 4.4.5.2 Interpretation

The sharp, erosional base and fining upward heterolithic facies succession indicates deposition within a tidal channel (T or Tf *sensu* Ainsworth et al. 2011). This interpretation is supported by abundant tidal indicators, most notably the fluid mudstones (Ichaso and Dalrymple, 2009), mud-draped cross-stratification (Boersma, 1969; Terwindt, 1971; Van den Berg et al., 2007) and apparent bidirectional current ripple cross-lamination (Reineck and Wunderlich, 1968). The fining upward heterolithic facies succession resembles those developed in tide-influenced point bars, which form inclined heterolithic strata (Reineck, 1958; Thomas et al., 1987; Choi et al., 2004). The large ranges in bioturbation intensity and trace fossil diversity indicate a wide range in environmental conditions. For example, the bispecific trace fossil suite of *Palaeophycus* and *Planolites* indicates chemical stress, most likely due to brackish salinities due to mixing of fresh river water and marine water. In contrast, the more diverse suites, which constitute the *Skolithos* and *Cruziana* ichnofacies indicate that these channels were connected to the sea, without the input of fresh water from fluvial sources (MacEachern and Bann, 2008).
Figure 4.5. Panel showing photographs of facies associations FAS and FA7, and related erosional surfaces, colour coded by facies association (Figure 4.3). For abbreviations see Figure 4.3. (A) Erosional surface overlain by a lag of quartz pebbles and 1-10 cm wide mud clasts (facies F4a; well 6406/8-1 4464 m). (B) Sharp based sandstones with mud draped cross-bedding (facies F5c) and mud-draped cross-lamination (facies F5b) and homogenous mudstones (facies F5a; well 6407/6-5 2512 m). (C) Apparent bidirectional, mud-draped, cross-laminated, heterolithic sandstones with bioturbation (facies F5b; well 6407/6-5 2510 m). (D) Muddy sandstone with rootlets (facies F7b) overlain by a coal bed (F7c; 6407/10-2 3447 m). (E) *Thalassinoides* burrows constituting a *Glossifungitis* ichnofacies, that marks a firmground surface (facies F6d; well 6407/6-3 2589 m). (F) Homogenous mudstone with synaeresis cracks sharply overlain by cross-laminated sandstone (facies F5a; well 6407/6-5 2553 m). (G) Thin coal horizon with rootlets penetrating the underlying sandstone (facies F7c; 6407/10-2 3454 m).
4.4.6 FA6 Tidal coastline

4.4.6.1 Description

In general, FA6 gradationally overlies FA5 and has an erosional upper contact with FA5, but it is also less commonly interbedded with FA3 and FA4. At the top of the Ile Formation, FA6 is gradationally overlain by FA7. FA6 consists of heterolithic, very fine- to fine-grained sandstones containing two types of sedimentary structures. The first type (facies F6a) consists of 10-15 cm thick planar laminated, low angle cross-bedded and undulatory bedded, very fine- to fine grained sandstone beds (facies F6a, Figure 4.6A,B). The laminae within these beds are defined by sorting of grains of different mineralogy, and laminae are commonly paired (Figure 4.6B). The second type (facies F6b) consists of heterolithic, very fine- to fine grained, cross-laminated sandstones, which appear planar over the width (c. 10 cm) of the core. The current ripple cross-lamination shows abundant apparent bidirectional palaeocurrent directions, which often have the same attitude and approach the angle of repose. The latter suggests opposite palaeocurrent directions, although a component of different sections through 3D ripples cannot be completely excluded. Mud drapes are most common in the toesets and foresets of the cross-laminations (Figure 4.6A,C). The two dominant bedding styles (facies F6a and F6b) are interbedded on a decimetre-to-metre scale and may be overprinted by bioturbation of variable intensity (BI=0-5). In successions where facies F6a and F6b alternate, the beds of facies F6b are more intensely bioturbated than the adjacent beds of facies F6a. Beds that are destratified by the high intensity of bioturbation (BI=4-5) are classified separately as facies F6c. In all three facies, trace fossil diversity varies from sporadic occurrences of *Palaeophycus*, *Planolites* or *Scolithos* to more variable suites, including *Thalassinoides*, *Ophiomorpha*, *Cylindrichnus*, *Asterosoma*, *Rosselia*, *Teichichnus*, *Lockeia* and *Schaubcylindrichnus*. Facies F6a also contains escape traces. Generally, those beds with a lower bioturbation index also have a lower trace fossil diversity.

Locally, thin intervals (5-20 cm) in this facies association contain quartz granules and small pebbles floating in a matrix of mudstone and/or sandstone. In some cases, the quartz granules and pebbles
infill *Thalassinoides* burrows, forming *Glossifungites* ichnofacies surfaces (MacEachern et al., 1992) (facies F6d, Figure 4.5A, E).

FA6 is characterized by a serrate GR log response (20-80 API), which reflects the centimetre-to-metrescale, heterolithic bedding style. NPHI values are generally low (5-15 %) and RHOB values are intermediate to high (2.3-2.7 g/cm³).

### 4.4.6.2 Interpretation

The mud-draped current ripple cross-lamination (facies F6b), with variable and possibly bidirectional palaeocurrent trends, resembles deposition from tidal currents (Reineck and Wunderlich, 1968). In contrast, the low-angle cross-bedded, very fine- to fine grained sandstone beds of facies F6a are interpreted as the expression in core of HCS (Scott, 1992), which resulted from storm wave action (Dott and Bourgeois, 1982; Duke, 1985; Keen et al., 2012). The paired laminae in these beds can be interpreted to reflect the daily inequality of semi-diurnal tides (Vakarelov et al., 2012; Wei et al., 2016), but this implies that a single bed containing 30 paired laminae represents a storm with a duration of at least one month. More plausibly, the paired laminae can be attributed to sorting of different mineralogies during changing oscillation power of waves with alternating wave heights within a wave group (Hamm et al., 1993). Some of the more planar laminated beds could also represent deposition by river floods caused by seasonal storms (Gugliotta et al., 2016), which simultaneously increase both the strength of unidirectional flows and the height, frequency and intensity of oscillatory waves (‘storm flood’). Rapid deposition of these event beds is supported by the relatively low bioturbation intensity and the presence of escape traces. *Asterosoma* and *Schaubcylindrichnus* are indicative of relatively open marine conditions (MacEachern and Bann, 2008). This, and the close affinity with the channelised facies FA3, FA4 and especially FA5, which is usually gradationally overlain by FA6, suggests that deposition took place in subtidal to intertidal palaeo-water depths. This implies deposition in upper delta front to lower delta plain environments, within a tide-dominated, but wave- and fluvially-influenced, open coastline (Twf or Tfw sensu Ainsworth et al. 2011). Neogene-to-modern tidal flat
deposits are often rich in carbonaceous material (Van Straaten and Kuenen, 1957; Meor et al., 2013). However, large salt tolerant flora (e.g. mangroves) only evolved since the latest Cretaceous and flourished during the Neogene (Greb et al., 2006). Consequently, no abundant carbonaceous debris is expected in these Jurassic deposits, and neither is it observed. While such a mixed-energy coastline is common in the present-day, it is less commonly interpreted in the geological record (Johnson, 1975; Fan, 2012; Eide et al., 2016). The combination of HCS and heterolithic facies reworked by tidal currents has been observed in both modern linear coastlines with tidal flats (Yang et al., 2005; Fan, 2012) and their ancient equivalents (Basilici et al., 2012).

The trace fossil assemblages constitute the Skolithos and Cruziana ichnofacies (MacEachern and Bann, 2008), but their large variability in trace fossil intensity and diversity probably reflects variations in position with respect to the input of fresh water and sediment along the coastline (Li et al., 2011; Korus and Fielding, 2015; Ayranci and Dashtgard, 2015). In locations close to river mouths, brackish water and high sedimentation rates can result in impoverished trace fossil suites and low bioturbation intensity. In locations more distant from or updrift of river mouths, more fully marine salinities and lower sedimentation rates resulted in more diverse and more intense bioturbation (MacEachern and Bann, 2008). The thin intervals containing quartz granules, pebbles and Glossifungitis are interpreted to represent time of erosion and winnowing of FA3, 4, and/or 5 (all contain quartz granules and pebbles) by tides and/or waves.
Figure 4.6. Core photographs of facies association FA6. (A) Overview 3 m interval containing alternations of facies F6a (dark blue) and F6b (light blue; well 6406/5-1). (B) Undulatory and planar laminated sandstones of (facies F6a, well 6406/8-1 4409 m). (C) Apparent bidirectional cross-lamination with mud drapes (facies F6b; well 6406/5-1 4357 m).
4.4.7 FA7 Coal mire

4.4.7.1 Description

FA7 gradationally overlies alternations of FA5 and FA6 and is only present in the top 20 m of the Ile Formation. FA7 consists of 20-50 cm thick cross-bedded (facies F7a) and current ripple cross-laminated (facies F7b), heterolithic (0-25 % mudstone) very fine- to fine-grained sandstones with rootlets and abundant disseminated carbonaceous debris. Capping these two facies, are 5-50 cm thick carbonaceous mudstone and coal beds (facies F7c, Figure 4.5D,G). Bioturbation displays low intensity (BI=0-3) and diversity, and mainly consists of Planolites, Cylindrichnus and Teichichnus.

Wireline logs of facies F7a and F7b are similar to those of FA6. In contrast, the coals (facies F7c) are easily distinguished due to their low RHOB (<2 g/cm³) and high NPHI (>45%) values.

4.4.7.2 Interpretation

The coals and underlying rootlet-penetrated sandstones indicate subaerial accumulation of peat in wetland conditions as histosols (Birkeland, 1999; Soil Survey Staff, 1999). The low-diversity trace fossil assemblage, which constitute impoverished Cruziana and Skolithos ichnofacies, and low bioturbation intensity in facies F7a and F7b supports a brackish, marginal marine setting (MacEachern and Bann, 2008). It is interpreted that the coals formed in a peat mire environment on the lower delta plain (Van Straaten and Kuenen, 1957).

4.4.8 FA8 Lower shoreface

4.4.8.1 Description

FA8 is only present in one well in the south of the study area, where it gradationally overlies mudstones of FA1. Towards the top of the Ile formation, FA8 is interbedded with coals of FA7. FA8 consists of successions of very fine- to fine-grained sandstone beds (10-30 cm), which are internally dominated by HCS, commonly overprinted by bioturbation of low intensity (BI=0-3) (facies F8a). The sandstone beds become increasingly amalgamated upwards, and intervening shales become correspondingly
thinner and less abundant. Where bioturbation is more intense (BI=4-5), it obscures the HCS (facies F8b). Trace fossils include Planolites, Palaeophyces, Schaubcylindrichnus, Thalassinooides, Teichichnus, Skolithos, Cylindrichnus, Lockeia, Ophiomorpha and escape traces.

The wireline log expression of FA8 consists of multiple, stacked successions of high and upward-decreasing GR (30-70 API), NPHI (10-20 %) and RHOB (2.2-2.5 g/cm³) values.

### 4.4.8.2 Interpretation

HCS indicates deposition under the influence of oscillatory waves between storm- and fair-weather wave-bases (Dott and Bourgeois, 1982; Duke, 1985; Keen et al., 2012). The high intensity of bioturbation and diverse trace fossil assemblage, which constitutes a mixed Cruziana and Skolithos ichnofacies, indicate deposition under fully marine conditions away from river-derived fresh water and sediment supply (MacEachern and Bann, 2008). FA8 therefore records wave-dominated deposition in the absence of tides or river influence (W sensu Ainsworth et al., 2011).
Figure 4.7. (A,C) Example core logs through upward-shallowing, progradational successions of facies associations FA1, FA2, FA3 and FA4 in wells 6407/6-3 and 6406/9-2. (B,D) Example core logs through aggradational successions of facies associations FA5, FA6 and FA7 in wells 6407/10-2 and 6407/6-5.
4.5 Vertical facies successions and lateral facies distributions

The facies analysis presented above provides the framework for interpreting vertical and lateral relationships between facies associations (Figure 4.9, 4.9). This allows interpretation of key stratal surfaces, notably marine flooding surfaces, and their lateral extent as determined from well log correlations (Figure 4.10). The key stratal surfaces forming the base and top of the studied interval are constrained by biostratigraphy. However, due to deep burial (3.5-4.5 km), both seismic imaging and palynomorph preservation are poor, which limits their value in guiding high-resolution correlations within the studied interval. The succession comprises three genetically-related intervals, defined here in terms of depositional phases (from oldest to youngest): Phases I and II are progradational, and Phase III is aggradational (Figure 4.9). The successions are correlated and mapped throughout the study area (Figure 4.10, 4.10), and a depositional model for each succession is presented (Figure 4.12).
Figure 4.8. Cross plots for neutron porosity (NEU) vs. bulk density (DEN) and gamma-ray (GR) vs. sonic (DT) for wells 6409/9-3, 6409/9-1, 6406/9-2 and 6406/8-1.
Figure 4.9. Example well (6407/6-5) showing the three vertical facies succession that are used to subdivide the studied strata into stratigraphic intervals. For legend see Figure 4.10.
Figure 4.10. Correlation panel with north-south (A-C) and west-east (D-F) orientations. For location see Figure 1C.
Figure 4.11 (previous page). Maps for stratigraphic intervals progradational phase I, progradational phase II and aggradational phase III (Fig. 8): (A) isopach of total thickness for progradational phase I; (B) extent of condensed section marked by high GR values in basal part of progradational phase I; (C) sandstone thickness (FA2, FA3, FA4) for progradational phase I, with inferred sediment supply routes (black arrows); (D) channel-fill deposit (FA3, FA4) thickness for progradational phase I, with inferred sediment supply routes (black arrows); (E) isopach of total thickness for progradational phase II; (F) proportions of wireline log facies in each well (pie charts) for progradational phase II, and highlighting areas of gradational and “sharp-based” delta-front successions; (G) isopach map of aggradational phase III; (H) proportions of wireline log facies in each well (pie charts) for aggradational phase III.

Figure 4.12. Depositional models (A,C) and interpreted process regime (B,D, modified after Ainsworth et al., 2011) for progradational phases I and II (A,B) and aggradational phase III (C,D).
4.5.1 Progradation Phase I

4.5.1.1 Description

This succession consists of mudstones of the Ror Formation and sandstones of the overlying Tofte Formation, and is bounded by two prominent marine flooding surfaces (Figure 4.9). The gross thickness of this mudstone-dominated succession varies between 200 m in the northwest and 100 m in the north, east and south (Figure 4.10, 4.10A). The log expression is characterized by a gradual upward decrease in GR, RHOB and NPHI values, which is bounded by two sharp upward increases in GR, RHOB and NPHI (Figure 4.9). The base of this succession is characterized by a thin (0-10 m) uncored interval with very high GR readings (>100 API; Figure 4.10), which is only present in the centre of the basin (Figure 4.11B). This is succeeded by a series of coarsening upward successions that comprise the following (from base to top): prodelta and offshore mudstones (FA1), mixed fluvial- and wave-dominated, tide-influenced delta front and mouth bar sandstones (FA2), and fluvial- (FA3) and marine-influenced fluvial channels (FA4). The coarsening upward successions average c. 20 m of thickness, with a range of 5 to 50 m (Figure 4.7A,C). Internally they are laterally variable, with individual coarsening upward successions only correlated over limited distances (c. 25-50 km), and not over the entire basin (c. 150 km) (Figure 4.10). The facies proportions are also variable, although FA1 and thin beds of FA2 are prominent in most wells. However, locally FA1 may occur in thinner intervals when the succession is dominated by sand-rich intervals of FA2, FA3 and FA4 (Figure 4.10). The net sandstone thickness, which reflects the thickness of FA2, FA3 and FA4, is greatest in the northwest (70 m) and decreases south-eastwards (10 m) before increasing in the east and southeast (50 m) (Figure 4.11C). The net channel-fill thickness (i.e. the combined thickness of FA3 and FA4) is also greatest in the northwest (40m), decreases south-eastwards to zero in the centre of the study area and increases again towards the east and southeast (10-40 m) (Figure 4.11D).
4.5.1.2 Interpretation

The bounding surfaces of this succession represent abrupt increases in water depth from channelised sandstones (FA3, FA4) to marine mudstones (FA1), and are interpreted as maximum flooding surfaces based on their regional extent (Gjelberg et al., 1987; Ravnås et al., 2014). The thin interval of high GR values at the base of the succession is interpreted as a condensed section, reflecting very low deposition rates. The correlation of flooding surface and condensed section the base of this succession is constrained by biostratigraphy. The extent of the condensed section defines the basin-centre at this time (Figure 4.11B). However, the correlation of the flooding surface at the top of phase I is an interpretation due to poor biostratigraphic resolution, and could be diachronous.

The coarse grain size (up to very coarse-grained and pebbly sandstone) of the coarsening upward successions (FA1, FA2, FA3 and FA4) suggests a high-energy and high-gradient fluvio-deltaic system. Similar sandstone textures in modern systems are often associated with braid deltas, which are dominated by high-discharge fluvial input (McPherson et al., 1987; Nemec and Postma, 1993; Orton and Reading, 1993). A braid delta interpretation is supported by sandstone distribution patterns (FA2, FA3 and FA5), which indicate a series of lobate protrusions from the opposite margins of the basin (Figure 4.11A, C, D). Evidence of storm wave reworking is provided by HCS (facies F2a) and by the well-sorted nature of the delta front and mouth bar sandstones (facies F2b, F2c and F2d). This can be compared to wave reworking in modern coarse-grained braid deltas, such as the Skeiðarársandur delta (Hine and Boothroyd, 1978), the Godavari delta (Nageswara Rao et al., 2005; Nageswara Rao et al., 2015) and, to a lesser extent, the more protected Burdekin delta (Fielding et al., 2005b; Fielding et al., 2005a; Fielding et al., 2006). Although tidal indicators are difficult to recognize in coarse-grained deposits (Dashtgard and Gingras, 2007), the presence of marine-reworked channels (FA4) implies that tidal influence affected some channels. Other potential modern analogues occur in bayhead deltas in fjords such as the Klimaklini and Honuthko (Syvitski and Farrow, 1983) and Bella Coola river deltas in Canada (Kostaschuk and McCann, 1983; Kostaschuk, 1985). The envisaged depositional environment
is a fluvial- and wave- dominated, tide-affected braid delta (Wft or Fwt sensu Ainsworth et al. 2011) (Figure 4.12A, B).

The successions are vertically stacked to define an overall progradational trend (Figure 4.9, 4.9). The northwest to southeast reduction in the total thickness, net sandstone thickness (FA2, FA3 and FA4) and proportion of channelised facies (FA3 and FA4) supports south-eastward delta progradation (Figure 4.11A, C, D), which downlapped onto the basin-centred condensed section (Gjelberg et al., 1987). The eastern margin of the basin displays an opposing trend, with a north-westwards reduction in net sandstone thickness (FA2, FA3 and FA4) and channelised facies proportions (FA3 and FA4) (Figure 10C, D). This indicates the north-westward progradation of either smaller delta lobes or the distal termination of a larger system to the southeast of the study area. Because deltas prograded into the basin centre from multiple directions, abandonment surfaces cannot be correlated across the basin centre. This relationship is shown in Figure 4.10 by interfingering of abandonment surfaces from either sides of the basin in the basin centre. Due to poor biostratigraphy, the exact mode of interfingering is uncertain.

4.5.2 Progradational Phase II

4.5.2.1 Description

This succession is 30-100 m thick and consists of mudstones of the Ror Formation and overlying sandstones of the lower Ile Formation (Figure 4.9, 4.9, 4.10E). The base of the succession is marked by a sharp upward increase in GR, RHOB and NPHI values. The top is picked immediately above the lowermost channel-fill sandstone of the Ile Formation (FA3, FA4 or FA5) (Figure 4.9 and 4.9). This succession is laterally extensive across most of the study area, but is either absent or too thin to distinguish on certain horst blocks, particularly in the Njord Field (Figure 4.10B) and on the footwall of the Bremstein Fault Complex (Figure 4.10D, E, F). The significance of these thickness patterns is discussed later, in the context of younger strata (aggradational phase III; Figure 4.9).
This interval mainly comprises a single coarsening upward succession of prodelta and offshore mudstones (FA1), delta front and mouth bar sandstones (FA2) and channelised sandstones (FA3, FA4 and FA5) (Figure 4.9 and 4.9). In many locations there is an abrupt vertical change where erosionally-based channelised facies of FA3, FA4 and FA5 directly overlie marine mudstones of FA1 (e.g. well 6407/6-5; Figure 4.10). In this latter example, channel sandstones overlie an erosional surface lined by quartz pebbles, while the preceding marine mudstones (FA1) also contain dispersed coarse sand grains and, occasionally, granules. The locations where the vertical succession is gradual (comprising FA2) form two broad north-south striking belts parallel to the basin margins (Figure 4.11F).

4.5.2.2 Interpretation

The base of this succession is marked by an abrupt increase in water depth, which defines a regional flooding surface (Gjelberg et al., 1987; Ravnås et al., 2014). The overlying coarsening upward succession indicates the widespread, laterally-coalesced succession of fluvial- and wave- dominated, tide-influenced deltas (Wft, Fwt or Ftw sensu Ainsworth et al. 2011) (Figure 10F). Hence, similar depositional environments characterised both Phases I and II (Figure 4.12A,B). However, the vertical and lateral stacking patterns of facies associations are different, most notably where channelised sandstones (FA3, FA4, FA5) directly overlie marine mudstones (FA1). These sharp-based, tide- and marine-influenced and partly channelised sandstones (FA3, FA4 and FA5) could represent the transgressive backfilling of an incised valley above a sequence boundary (Catuneanu, 2006; Dalrymple et al., 2006). However, the sharp-based sandstones form a laterally extensive sheet, which is inconsistent with elongate sand body geometries typical of valley-fills (Figure 4.10, 4.10F). In addition, the occurrence of coarse sand grains and granules beneath the erosion surface (in well 6407/6-5) suggests proximity to a contemporaneous coastline with coarse-grained fluvial sediment input. Hence, a forced regressive shoreline interpretation is proposed, in which prograding deltas followed a descending regressive trajectory (Posamentier and Morris, 2000; Helland-Hansen and Hampson, 2009; Prince and Burgess, 2013). In contrast, those locations comprising gradual coarsening upward facies successions (Figure 4.11H) record ascending regressive shoreline trajectories. Such successions occur
in two north-south striking belts that lie east and west of the basin centre. It is interpreted that these two belts formed simultaneously as deltas from either side of the rift basin and prograded with ascending regressive shoreline trajectories during a time of increased A/S ratios. An explanation for increased creation of accommodation space could be increased subsidence due to some minor rift activity (Ravnås and Steel, 1998). Modern analogues to this system, where fluvial-dominated deltas prograde into shallow water basins with a near-horizontal shoreline trajectory, include the modern Burdekin delta (Fielding et al., 2005a) and the Neogene and modern Volga delta (Kroonenberg et al., 1997; Overeem et al., 2003; Hinds et al., 2004; Kroonenberg et al., 2005). These examples contain sharp-based mouth bars and distributary channels that cut into offshore mudstones, similar to some of the Phase II successions.

4.5.3 Aggradational Phase III

4.5.3.1 Description

This succession corresponds to the middle and upper Ile Formation and has a thickness of c. 100 m in the west and thins gradually to c. 40 m in the east of the study area (Figure 4.9, 4.9, 4.10G). The base of this succession is picked at the top of the lowermost fluvial channel-fill sandstone (FA3, FA4 and FA5) in the Ile Formation. The transition from progradational phase II to aggradational phase III is a transition from deposition on the delta front to deposition on the delta top. Therefore this transition is diachronous and occurs when the shoreline progrades across the area. The top of the succession is defined by a sharp increase in GR, RHOB and NPHI values at the base of the Not Formation (Figure 4.2, 4.8). Internally, the GR, RHOB and NPHI values are low and show a slight upward increase (Figure 4.9, 4.9). The eastward thinning is gradual, except for locations on horst blocks, such as in the Njord Field (Figure 4.10B) and on the footwall of the Bremstein Fault Complex (Figure 4.10D, E, F). Where the Ile Formation is much thinner, progradational Phase II and aggradational Phase III cannot be distinguished. Cores from these locations contain up to six firmground surfaces marked by the Glossifungites ichnofacies (MacEachern et al., 1992), which are not recognized elsewhere in thicker,
more complete sections of the Ile Formation. These surfaces are also overlain by winnowed lags of granules and small pebbles (facies F6d, Figure 4.5A, E). These firmground surfaces are present locally, and cannot be correlated between wells.

Phase III is dominated by vertical alternations of tidal channels (FA5) and tidal coastal deposits (FA6), with localised occurrences of cross-cutting channels (FA3 and FA4). Delta front successions (FA1, FA2) are absent, and the proportion of channel-fills (FA3, FA4 and FA5) decreases upwards. Channel-fill sandstones are more abundant in the northeast and west (Figure 4.11G). Coal layers (FA7) occur towards the top of the succession, but only in the south (Figure 4.7B, D, 4.9B, E, F, 4.10F, H). Biostratigraphic data suggest that Phase III correlates with interbedded offshore mudstones (FA1), lower shoreface (FA8) and coals deposits (FA7) in the southwest (well 6406/11-1S in Figure 4.10A, F).

4.5.3.2 Interpretation

The vertical stacking of abundant tide-influenced channel-fill sandstones (FA5) and tidal coastal plain deposits (FA6) throughout the succession implies aggradation of a lower delta plain with shallow subtidal to intertidal water depths. In this context, fluvial channel-fill sandstones lacking tidal influence (FA3) are interpreted as upper delta plain deposits that developed landward of the fluvial to tidal transition (Dalrymple et al., 1992; Dalrymple and Choi, 2007). The concentration of coals (FA7) in the upper part of the succession in the southern part of the study area reflects development of subaerial peat mire environments. Hence, water depth during Phase III deposition oscillated between shallow-subtidal to subaerial exposure, thereby recording a period of lower delta plain aggradation. The top of the succession is marked by an abrupt increase in water depth (Figure 4.2 and 8), which corresponds to a major, basin-wide flooding surface (Gjelberg et al., 1987; Ravnås et al., 2014).

During Phase III, two main types of shoreline were preserved: (1) non-deltaic wave-influenced tidal shorelines, and (2) mixed fluvial- and tide-dominated deltaic coastlines. Non-deltaic successions comprise alternating tidal channels (FA5) and tidal shorelines (FA6), which display high bioturbation intensity. In combination, these facies associations represent tide-dominated lower delta plains and
associated coastlines, which were distant from active fluvio-deltaic sediment input points (Figure 4.12C,D) and subject to periodic storms and wave reworking (Yang et al., 2005; Basilici et al., 2012; Fan, 2012). In ancient and modern analogues of this type of non-deltaic coastline episodic HCS-bearing storm deposits are interbedded with tidal current ripples (Yang et al., 2005; Basilici et al., 2012). Only locally was there sufficient elevation and fresh ground-water to enable peat mires to develop (FA7, Figure 4.11F). The envisaged depositional model was a tide-dominated and wave-influenced coastline (Tw sensu Ainsworth et al. 2011).

Deltaic successions are recognised where FA5 and FA6 display lower bioturbation intensity and diversity and where fluvial channels (FA3) and marine-influenced channels (FA4) are in close proximity. This would be consistent with relatively high sedimentation rates close to fresh water input points at river mouths (MacEachern and Bann, 2008; Li et al., 2011; Korus and Fielding, 2015; Ayranci and Dashtgard, 2015). The close association of fluvial- (FA3) and tide-influenced channels (FA4 and FA5) supports deposition in a mixed tide- and fluvial-dominated delta. Evidence for minor wave reworking in mouth bars (facies F4d) and tidal flats (facies F6a) indicates subordinate wave influence (Ftw or Tfwsensu Ainsworth et al. 2011) (Figure 4.12C,D). The Mahakam delta provides an analogue for a mixed energy, fluvial- and tide-influenced delta. The Mahakam delta plain comprises laterally adjacent active distributary fluvial channels and abandoned distributaries, which are effectively tide-dominated estuaries (Allen and Chambers, 1998). Active distributaries are filled with fine- to medium grained sand on much of the delta plain, and coarse-grained sand at the apex of the delta (Allen and Chambers, 1998; Salahuddin and Lambiase, 2013), which could be analogous to the grain size difference between FA3 and FA4. The abandoned distributary channels of the Mahakam contain a mix of mud and fine grained sand (Allen and Chambers, 1998; Salahuddin and Lambiase, 2013), which compares favourably to the heterolithic very fine- to fine grained tidal channels of FA5. The higher proportion of finer grained tide-dominated channels in the modern Mahakam delta plain is also comparable to that seen in the Phase III succession (Figure 4.10). Ancient analogues can be found in the Cretaceous Sego Sandstone, USA (Willis and Gabel, 2001; Willis and Gabel, 2003; Legler et al., 2014; Van Cappelle et al.,
2016), the Jurassic Neill Klinter Group, Greenland (Dam and Surlyk, 1998; Ahokas et al., 2014b; Eide et al., 2016) and the Jurassic Lajas Formation, Argentina (McIlroy et al., 2005; Gugliotta et al., 2015; Rossi and Steel, 2016). Each of these three examples contains similar aggradational successions to those documented during Phase III deposition, containing (1) relatively coarse-grained fluvial channel-fill sandstones (cf. FA3, FA4), (2) relatively fine-grained and heterolithic tidal channel-fill sandstones (cf. FA5), and (3) wave-reworked mouth bar and delta front sandstones. The Neill Klinter Group also contains heterolithic tidal flat deposits (cf. FA6). Fluvial channel-fill sandstones in the Lajas Formation and Neill Klinter Group are pebbly and coarse-grained, as are those in the Ile Formation.

In aggradational Phase III, a higher proportion of coarse-grained channel-fill sandstones (FA3, FA4 and FA5) are present in the east, north and west, which correlate to offshore mudstones (FA1), shoreface sandstones (FA8) and coal mire deposits (FA7) to the southwest in well 6406/11-1S (Figure 4.11F, H). In the context of the regional setting (Figure 4.1A), it is interpreted that sediment input was transverse to the rift axis, both from the west and east, and along the rift axis from the north. The occurrence of offshore (FA1) and shoreface (FA8) deposits to the south can be explained by three possible scenarios: (1) a change in depositional process from tide-dominated to wave-dominated, due to progradation along the rift-axis, which widened to the south (Fig. 1A) (Legler et al., 2014; Olariu, 2014); (2) wave-dominated deposition was contemporaneous with the tide-dominated deposition along the depositional strike of the shoreline outside of the embayment; and (3) sedimentation could have been locally sourced from a neighbouring horst block or uplifted footwall (Ravnås and Steel, 1998).

The restricted occurrence of coals (FA8) in the south can be explained in two ways: (1) a coal-bearing lower delta plain environment was located in the south, with a coeval, non-coal-bearing upper delta plain in the north; and (2) coals only developed locally on slowly subsiding or uplifted horst blocks and footwall blocks associated with faults that were active during Ile deposition.

The reduced thickness of Phase II and Phase III successions (Figure 4.9) on the Njord Field horst block (Figure 4.10B) and in the footwall of the Bremstein Fault Complex (Figure 4.10D, E, F) confirms active
faulting during deposition of Phases II and III. This is supported by evidence of erosion and winnowing by wave and/or tidal action in these areas, which resulted in the localised occurrence of firmgrounds and lags within the Ile Formation. Subsequently, uplift of the footwall of the Bremstein Fault Complex resulted in the erosion and thinning of Ile Formation in the Draugen Field (Figure 4.10E, F) (Goesten and Nelson, 1992; Provan, 1992).

4.6 Stratigraphic models for temporal evolution of the “Tofte and Ile deltas”

A stratigraphic model is outlined below to account for the temporal and spatial changes in the grain size characteristics and depositional process regime during the deposition of Phases I to III. The model considers the observed stratal architectures, grain size characteristics and facies patterns within the context of the sediment routing systems, from source area to depositional sink (Figure 4.13A).

Figure 4.13. A) Map synthesising the interpreted location and size of the “Tofte and Ile sediment routing systems”, from their erosional source areas to their depositional sinks (Gjelberg et al., 1987; Ziegler, 1988; Surlyk, 1990; Doré, 1992; Alsgaard et al., 2003; Vosgerau et al., 2004; Nøttvedt et al., 2008; Elliott et al., 2012; Sømme et al., 2013; Sømme and Jackson, 2013; Ravnås et al., 2014; Elliot et al., 2015; Eide et al., 2016). (B) Sketch cross-section illustrating the potential thickness of the principal “Tofte and Ile sediment routing system” along the major axial depositional system, as required
to explain the overall progradational-aggradational-retrogradational architecture of the associated strata via autoretreat.

Grain size distributions expected from mass-balance consideration of this cross-section are also shown. (C, D) Expanded view of the distal part of Figure 12B showing grain size distributions (C) and stratal architecture and facies-association distributions (D) in the study area and adjacent areas.

4.6.1 Catchment and hinterland

Transverse catchments feeding rift basins are usually numerous, but relatively small, because they are developed in the footwall of faults bounding rotated intra-basinal or basin-marginal blocks (Leeder and Gawthorpe, 1987; Ravnås and Steel, 1998; Martinsen et al., 2005; Ravnås et al., 2000; Elliott et al., 2012; Elliott et al., 2015; Sømme et al., 2013). However, the abundance of coarse grained, extra-basinal sediment, which cannot be derived from reworking the finer grained, underlying successions, precludes local, intra-basinal source areas. It has been proposed that some of the sediment at the Halten Terrace was derived from Greenland (Morton et al., 2009). However, the presence of a palaeo-high between the Baltic Shield and Greenland (Doré, 1992; Nøttvedt et al., 2008; Eide et al., 2016) precludes a source of sediment as far Greenland (Figure 4.13A). Hence, it appears likely that extra-basinal sediment source areas were located close to the Halten Terrace (Ravnås et al., 2014). The Halten Terrace may have formed part of a fault-bounded, south-facing embayment with major sediment sources to the west, east and along the axis of the rift to the north, and a sink in the rift-axis to the south (Ziegler, 1988; Doré, 1992). This is consistent with regional reconstructions (Gjelberg et al., 1987; Ravnås et al., 2014), with the Halten Terrace forming the conjugate margin to age equivalent strata on East Greenland (Surlyk, 1990; Alsgaard et al., 2003; Vosgerau et al., 2004; Eide et al., 2016).

4.6.2 Change from progradation (Phases I and II) to aggradation (Phase III) to transgression (Not Formation)

The change from net progradation (Phases I and II) to net aggradation (Phase III), followed by regional transgression (basal Not Formation), can be explained by a combination of two possible mechanisms. In the first model, the stratal architecture may record a cycle of relative sea-level and/or sediment supply change (i.e. decreasing and then increasing A/S ratio). In this interpretation, (1) net
progradation is due to sediment supply being greater than the rate of accommodation creation (A/S < 1), (2) aggradation reflects sediment supply being in balance with the rate of accommodation creation (A/S = 1), and (3) transgression reflects accommodation creation outpacing sediment supply (A/S > 1).

In rift basins, such conditions occur during inter-rift periods of tectonic quiescence when subsidence rates are reduced while sediment supply remains constant (Ravnås and Steel, 1998; Ravnås et al., 2000; Ravnås et al., 2014).

In the second model, the stratal architecture may represent autoretreat (Muto and Steel, 1992; Muto, 2001; Muto et al., 2007) of the deltas under conditions of constant sediment supply and a constant rate of accommodation creation (i.e. constant A/S ratio). Such boundary conditions may occur in rift settings, such as the Halten Terrace, during times of tectonic quiescence when subsidence rates and sediment supply are constant (Muto and Steel, 1992). Under such conditions, deltas may prograde initially across the antecedent basin-floor topography. However, the total surface area of deposition on the delta plain increases as progradation continues, and progressively more of the supplied sediment is trapped within the expanding delta plain. Therefore the sediment supply to the shoreline progressively decreases, causing a reduced shoreline progradation rate and eventual retrogradation without any external forcing.

The models are not mutually exclusive, so a combination of both models may have influenced the study area. Initially simultaneous delta progradation from all the sediment delivery pathways suggests that an allogenic forcing mechanism drove the progradation. Delta progradation and progressively higher rates of sediment accumulation on the delta plain may have driven autogenic aggradation and autoretreat (Muto and Steel, 1992; Muto and Steel, 2001). The architecture and dimensions of the major axial routing system during progradation (Phases II) and aggradational (Phase III) had the following cross-sectional geometry (Figure 12B): (1) length scale (250-500 km), (2) thickness (100-150 m), and (3) assumed slope of the coastal plain (0.01°). Based on this depositional framework, the time scale for autoretreat of the axial system is estimated at 1.5-9 Myr using the methodology of Muto et
al. (2007). This is similar to the observed duration for progradation-to-aggradation of Phases II and III of 5-10 Myr (Ror and Ile formations, Figure 4.2). The abrupt flooding at the top of aggradational Phase III is explained by renewed rift activity (Ravnås and Steel, 1998; Ravnås et al., 2000; Ravnås et al., 2014) causing increased subsidence, which rapidly terminated autoretreat (cf. Fig. 6A of Muto et al., 2007).

4.6.3 Sediment dispersal and mass balance considerations

For sediment routing systems with a closed sediment budget, and a uniform proportion of input grain sizes, the theory of mass balance predicts that downsystem transitions in grain size and associated facies belts occur at fixed locations along a length scale normalised to deposited sediment mass (Strong et al., 2005). In either of the allogenic and autogenic forced models outlined above, progradation followed by aggradation implies that the grain size transitions reflect upstream retreat of the sediment routing system over time, as a progressively greater proportion of sediment mass is stored on the delta plain (Muto and Steel, 1992). By implication, mass balance consideration implies that the coarsest grain sizes in lower delta plain and delta front deposits will be encountered during initial progradation, when pebbly, coarse sand was transported to the coastline where it was deposited in mouths bars (FA2) and distributary channels (FA3, FA4). During this period, fine-grained sand may have been bypassed to the basin floor, as is common in coarse-grained deltas (Walker, 1966; Hampson, 1997; Rohais et al., 2008; Backert et al., 2010), transported alongshore out of the study area (Hampson et al., 2014b) or the input sediment supply may have lacked this grain size fraction (Michael et al., 2013). During subsequent delta plain aggradation, pebbly, coarse sand could have been stored in fluvial channels (FA3) on the upper delta plain, with only very fine-to-medium sand reaching the lower delta plain, where it was deposited in tidal flats and channels (FA5, FA6; Fig. 12B, C).

4.6.4 Link to depositional process regime

The depositional process regime of siliciclastic coastlines is influenced by several, partly linked factors such as grain size, river gradient and length (McPherson et al., 1987; Orton and Reading, 1993; Dashtgard and Gingras, 2007), shelf width and changes in shoreline morphology during regressions
and transgressions (Boyd et al., 1992; Ainsworth et al., 2011; Olariu, 2014). Therefore, the change from progradation in Phase II to aggradation in Phase III and the related decrease in sandstone grain size affects the preserved record of depositional process regimes. It is widely accepted that depositional systems become more tide-dominated during transgressions as the coastline rugosity increases and tidal currents are amplified in newly formed coastal embayments and drowned fluvial valleys (Boyd et al., 1992; Dalrymple, 2006b; Yoshida et al., 2007). Also, coarse grained coastal deposits are more likely to be dominated by fluvial processes due to the proximity of these coastlines to the hinterland and the related more rapid response to changes in sediment supply from the hinterland (Hine and Boothroyd, 1978; Kostaschuk and McCann, 1983; Syvitski and Farrow, 1983; Kostaschuk, 1985; McPherson et al., 1987; Orton and Reading, 1993; Fielding et al., 2005b; Fielding et al., 2006). In addition, recognition of tidal process in coarse grained deposits is problematic (Dashtgard and Gingras, 2007). Consequently, the change in depositional process from fluvial and wave dominated in progradational Phase I and Phase II to more tide dominated processes in aggradational Phase III may reflect a combination of (1) change in coastal behaviour (Olariu, 2014), and (2) the available sediment texture.

4.6.5 Reservoir distribution and quality prediction

The distribution of porous lithologies forming hydrocarbon reservoirs (sandstones in this study area) is predicted using the mass balance of sediment dispersal. The reservoir quality depends on the amount of heterogeneities and connectivity between sandstone bodies. The mass balance predicts, and the data shows that coarse grained sandstones are deposited towards the top of the coarsening upward successions during progradational phases I and II (FA2, FA3 and FA4). The reservoir quality is high due to the coarse sand grain size and lack of interbedded mudstones. These deposits are mixed fluvial- and/or wave-dominated deposits, so it is inferred from both depositional models of such systems and the data presented herein that these sandstones form well sorted, relatively coarse grained, laterally extensive sandstone bodies of high reservoir quality.
During Aggradational Phase III, more sediment is deposited on the delta top set, causing the mass balance to shift palaeo-landward. Therefore, relatively fine grained sand and mud was deposited during aggradation (FA5 and FA6). This means that the presence of reservoir rocks (channel-fills of FA3 and FA4) decreases upward through Aggradational Phase III. Also the reservoir connectivity decreases upward. Where channelised deposits in progradation phases I and II are inferred to be lateral extensive, correlative sandstones bodies, the channelised deposits in Aggradational Phase III cannot be correlated between wells. Also the reservoir quality decreases upward, due to the increase in mud-drapes and fluid mudstones deposited in tide-dominated environments which are more abundant upward in the succession.

4.7 Conclusions

- The Early to Middle Jurassic Ror, Tofte and Ile formations of the Halten Terrace, offshore mid-Norway preserves three genetically-related phases of clastic coastal-deltaic deposition, with distinct mixed-influenced process regimes. The older Phases I and II are progradational, and the younger Phase III is aggradational.

- The deposits of Phase I represent a series of progradationally-stacked, mixed fluvial- and wave-dominated, tide-affected deltas. The main delta system prograded southeastward, transverse to the rift axis. A series of minor deltas prograded westward from the opposite eastern margin of the basin.

- The deposits of Phase II represent a similar depositional environment to those of Phase I, but record only a single period of progradation. Commonly the channel fill deposits erosionally overlie marine mudstones, indicating that part of the progradation occurred with a descending regressive shoreline trajectory. In contrast, gradational coarsening upward successions were deposited during a period of ascending regressive shoreline trajectories.

- The deposits of Phase III are the result of aggradational stacking of (1) mixed tide- and fluvial-dominated (wave-affected) delta deposits, and (2) non-deltaic tide-dominated, wave-influenced
coastline deposits. These two environments were contemporaneous and record a spatial change in depositional process regime.

- The decrease in grain size, decrease in fluvial influence and increase in tide influence from progradational Phase II to aggradational Phase III is interpreted to be the result of a combination of quiescence in rift activity and autogenic retreat of the delta front. Initially, reduction in rift activity caused a decreasing A/S ratio. This resulted in delta progradation from multiple source areas. During progradation, an increasing proportion of the available sediment was stored on the lower delta plain, causing a decrease in progradation rate followed by delta aggradation (i.e. autoretreat). Eventually renewed tectonic activity resulted in abrupt flooding at the top of aggradational Phase III. Assuming that the proportion of different grain-sizes in the sediment supply were time invariant, and a that an increasing proportion of the sediment budget was deposited on the delta topset, then mass balance considerations predict a decrease in grain size during delta topset aggradation, as observed in the change from Phase II to Phase III. The change from progradation to aggradation, and the related change in available grain size, may have enhanced the preservation signature of tidal processes.

- Reservoir quality is highest in the well sorted, relatively coarse grained, mixed fluvial- and/or wave-influenced, lateral continuous sandstone bodies of progradational phases I and II. When more sediment was deposited on the delta top during Aggradational Phase III, mass balance shifted, and more fine grained sand and mud was deposited resulting in fewer and poorer quality reservoirs higher up in the succession.
Chapter 5. Spatio-temporal evolution of coastal process regimes: southwestern margin of the Campanian Western Interior Seaway, USA

5.1 Introduction

Shallow-marine depositional systems are commonly interpreted in terms of the relative interaction of waves, tides and fluvial influence. A classification into these three end-members was first proposed by Galloway (1975) for regressive river deltas. Later, the classification was expanded to non-deltaic and transgressive coastlines, categorizing a range of modern depositional environments (Boyd et al., 1992). This classification scheme was extended further by Ainsworth et al. (2011) in an attempt to approximately quantify the relative importance of waves, tides and fluvial depositional processes for the full spectrum of clastic 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) of the Western Interior Seaway of North America. This is done in order to understand how coastal deposition system evolve over time. This can be applied to predict reservoir/aquifer distributions and quality in hydrocarbon exploration and production respectively. The study area is regionally-extensive (Figure 5.1), covering 600,000 km² and comprising parts of New Mexico, Colorado, Utah and Wyoming (USA) 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). This has 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 (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; e.g. Van Wagoner et al., 1990). Where the strata occur in the subsurface, mapping can be extended 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 (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). Several previous studies have reconstructed palaeoshoreline 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 been compiled at low temporal resolution (McGookey et al., 1972; Kauffman and Caldwell, 1993). In any case, all these studies reconstructed the palaeo-coastline positions, but not the depositional process regime along these palaeo-coastlines, which is the aim of this work.

The main aim of this paper is to synthesise the spatial and temporal variations in coastal morphology and depositional processes in the studied strata in order to better understand the development of such systems and make predictions in other, more data-poor cases. Of especial interest is the distribution and potential origin of regressive coastlines with a regionally pronounced tidal signature (Aschoff and Steel, 2011b; Steel et al., 2012), because tide-dominated depositional systems are often heterogeneous and forms reservoir rocks from which it is challenging to recover hydrocarbons. The presence of a “Utah Bight” embayment, in which tides can be amplified, has long been recognized (McGookey et al., 1972). However, the relationship between the temporal and spatial evolution of this embayment and the depositional processes of the coastal and shallow marine environments has yet not been fully determined. Part of this assessment into the spatial and temporal variations in coastal morphology and depositional processes utilises the classification method of Ainsworth et al. (2011).

The secondary aim of this study is the application of this classification scheme on some of the world’s best-exposed and most widely studied strata, including continuous outcrops and densely spaced wells, which will enable the assessment of suitability and degree of uncertainty of the methodology of Ainsworth et al., (2011). The main unknown of this method (Ainsworth et al., 2011) is in how much detail and accuracy deposits can be classified without over-interpreting detailed observations from data-poor areas elsewhere (e.g. core data).
Figure 5.1. A) Regional tectonic framework. 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 (various sources, see sub-heading ‘Active Structures’).
5.2 Geological setting

From the Late Jurassic to the Paleogene, the Farallon plate was subducted below the western margin of the North American plate (DeCelles, 2004). 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). Due to the subduction of the cold Farallon slab, dynamic subsidence took place across much of the width of the North American plate, causing the epicontinental Western Interior Seaway to flood the continent (Figure 5.1A,B) (Liu and Nummedal, 2004; Liu et al., 2011; Liu et al., 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). 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 (Burtner and Nigrini, 1994; DeCelles and Mitra, 1995; DeCelles et al., 1995; DeCelles, 2004; Yonkee and Weil, 2010; Yonkee and Weil, 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; Yonkee and Weil, 2015). There is subtle evidence for the activity of some Laramide-style structures during the Campanian and Maastrichtian, from changes in stratal thickness, drainage directions and mineralogical composition (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). 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 fluvio-deltaic clastic wedges with an overall north-south striking 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, 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; Lawton and Bradford, 2011; Lawton et al., 2014; Szwarc et al., 2014).

Several previous studies have reconstructed palaeoshoreline 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 been compiled at low temporal resolution (McGookey et al., 1972; Kauffman and Caldwell, 1993). In any case, all these studies reconstructed the palaeo-coastline positions, but not the depositional process regime along these palaeo-coastlines, which is the aim of this work.

During the Campanian, the study area was located in the northern hemisphere between 40-55° 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 harbouring savannah vegetation, and just north of the study area there was a transition to cool temperate climates harbouring 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, 1990; Slingerland et al., 1996; Slingerland and Keen, 1999).

5.3 Dataset

The synthesis presented here is based on published outcrop studies in the study area and original work on the tide-influenced progradational lower Sego Sandstone (Van Cappelle et al., 2016). The mapped timeslices have been chosen to coincide with periods of maximum transgression and regression of the clastic wedges. These coincide with pronounced differences in process regime in the Book Cliffs outcrop belt, along the southern margins of the Uinta Basin, east-central Utah, and the Piceance Basin, west-central Colorado (Aschoff and Steel, 2011b). In ascending stratigraphic order (oldest to
youngest), the five timeslices are: (1) maximum transgression of the wave-dominated Star Point Sandstone and lower Blackhawk Formation, (2) maximum regression of the wave-dominated upper Blackhawk Formation and Castlegate Sandstone, (3) maximum regression of the tide-influenced Sego Sandstone, (4) maximum regression of the wave-dominated Corcoran Member, and (5) maximum regression of the wave-dominated Rollins Member (Aschoff and Steel, 2011a).

5.3.1 Lithostratigraphy and biostratigraphy

The five timeslices have been defined based on ammonite biostratigraphy. A detailed ammonite biostratigraphy is available for the fully marine (offshore marine mudstones) parts of the Campanian of the Western Interior Seaway, with an average age resolution of 0.5 Myr (24 zones in 12 Myr). Ten of these ammonite zones are dated (Cobban et al., 2006). The gross environments of deposition (EOD) for each timeslice have been mapped based on regional cross-sections. These tie the ammonite zones to the lithostratigraphy of Campanian strata preserved within the following groups of Tertiary Laramide basins (Figure 5.2): (1) San Juan Basin (Molenaar et al., 2002), (2) Uinta and Piceance basins (Young, 1955; Cobban, 1969; Gill and Hail, 1975; Fouch et al., 1983), (3) Denver, Middle Park and Sand Wash basins (Izett et al., 1971), (4) Rock Spring Uplift (Roehler, 1978), (5) Washakie and Wind River basins (Merewether et al., 1997), (6) Hanna and Laramie basins (Gill et al., 1970), and (7) Bighorn and Powder River basins (Gill and Cobban, 1966). At the resolution of the ammonite biostratigraphic zones, the large-scale transgressions and regressions that define clastic wedges are diachronous (Figure 5.2) (Krystinik and DeJarnett, 1995). Each regressive-transgressive clastic wedge contains multiple, shorter-duration shallow-marine tongues, or parasequences, which are below the resolution of the biostratigraphic framework (10’s of meters thick), and which cannot be traced from outcrop belt to outcrop belt or within a long (10’s kilometres) outcrop belts. Because these parasequences are not regionally correlative, it is inferred that they are the result of autogenic or local allogetic forcing. Therefore, regressive and transgressive shorelines at the scale of these higher-resolution tongues are shown on the same timeslice, but are clearly labelled as belonging to a smaller-scale, either regressive or a transgressive, phase.
Figure 5.2 (next page): Chrono-, bio- and litho-stratigraphic framework of Campanian strata in the Tertiary (Laramide) basins in the study area (Figure 5.1C). The dashed white lines are the mapped time-slices. Chronostratigraphy after Gradstein et al. (2012).
5.3.2 Active structures

Thrust faults and folds that were active during Cretaceous deposition have been mapped (Figure 5.1C) (DeCelles, 2004). In the Sevier fold-and-thrust belt, the following three main groups of faults were active during the Campanian: (1) the Absaroka Thrust in the Wyoming Salient (Burtnet 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; Lawton, 1986; Miall and Arush, 2001).

5.4 Gross environments of deposition

Six gross EODs are identified and mapped within the studied strata (from landward to seaward): (1) alluvial-to-coastal-plain sandstones, (2) coastal-plain coals, mudstones and sandstones, (3) shoreline and inshore sandstones, (4) gravity-flow siltstones and sandstones, (5) marine mudstones, and (6) marine marls and chalk (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). Each EOD is described and interpreted briefly below. Detailed process classification of the shoreline and inshore sandstones is given in the next section.

5.4.1 Alluvial-to-coastal-plain sandstones

These deposits consist of amalgamated, erosively-based fine-to coarse grained, fining upward fluvial channel-fill sandstones. The sandstones contain cross-beds organised on inclined surfaces which dip in the same direction as the cross-bedding. This supports deposition on downstream-accreting bars in a
braided river setting. These channel-fill deposits form sandstone-dominated sheets (80-100% sand), which are 20-100 m thick and extend laterally for 10-100’s of kilometres (see orange-coloured units in Figure 5.2) (Gill and Cobban, 1966; Van de Graaff, 1972; Van Wagoner, 1995; Martinsen et al., 1999; Yoshida, 2000; Hampson, 2010; Aschoff and Steel, 2011a; Leva López and Steel, 2015).

5.4.2 Coastal-plain coals, mudstones and sandstones

These deposits are mudstone-dominated (50-100% mudstone) and consist of coal seams, rooted (carbonaceous) mudstones, thin sheet sandstones, and generally subordinate coarse-grained, sharp-based, fluvial channel fill sandstones (c. 20-80% sandstone). The coals occur within laterally-extensive mudstones, which were deposited in flood plain environments. Interbedded tabular to sheet-like sandstones were deposited as crevasse spays. Fluvial channel-fills mostly occur as isolated sand bodies, with subordinate occurrences of multilateral and multistory sand bodies (Hampson et al., 2012; Flood and Hampson, 2014; Flood and Hampson, 2015).

5.4.3 Shoreline and inshore sandstones

Shoreline and inshore sandstones display the greatest variability. Five main types are identified (discussed in more detail in Section 5.5), as follows: (1) wave-dominated shorefaces, delta fronts and estuaries; (2) fluvial-dominated deltas; (3) mixed tide- and wave-influenced deltas; (4) tide-dominated deltas; and (5) barrier islands and back-barrier lagoons. The majority of sand bodies deposited in these environments are coarsening and sandier upward, very fine- to medium grained sandstone successions. These sand bodies are correlative over tens of km’s along depositional strike and 10-50 km along depositional dip. They gradationally overlie marine mudstones and pass updip (palaeo-landward) into coastal-plain coals, mudstones and sandstones and downdip (palaeo-seaward) into marine mudstones. This EOD includes “isolated” shallow-marine sandstones, which are distinctive by being entirely encased in marine mudstones in the west-east orientation of most outcrop belts (Bergman and Snedden, 1999). They display similar facies characteristics to other shoreline and
coastal-sandstones. Moreover, when mapped-out to the north and south, they correlate landward to shoreline sandstones of high rugosity (Hampson et al., 2008; Mellere and Steel, 2000).

5.4.4 Gravity-flow siltstones and sandstones

Within the marine mudstones, decimetre to metre thick planar laminated and current ripple cross laminated siltstones and very fine- to medium grained sandstones are present. These sediments have been interpreted to be deposited from shallow water turbidity currents which have been linked to river mouths in the correlative shoreline and coastal-sandstones (Johnson, 2003; Pattison et al., 2007; Pattison, 2005).

5.4.5 Marine mudstones

The deposits of this environment (facies OS) contain claystones, mudstones and siltstones bioturbated (bioturbation index sensu Taylor & Goldring, 1999, BI=2-6) by Planolites, Palaeophycus, Thalassinoides, Scolicia, Teichichnus, Chondrites, Schauberclindrichnus, Helminthopsis and Zoophycos of the Cruziana and Zoophycos ichnofacies. The fine grain size and trace fossil assemblages of the Zoophycos and Cruziana ichnofacies indicate deposition in a low energy marine environment. These deposits dominate the basinward part of the study area.

5.4.6 Marine marl and chalk

In the most basinward part of the study area, marl and chalk beds are rhythmically interbedded on a decimetre scale. Decimetre scale alternations are astronomically forced, whereas variations in clay content over tens of metres are due to external influxes of clay (Locklair and Sageman, 2008). The deposits are bioturbated by Chondrites, Zoophycos, Teichichnus, Planolites and Thalassinoides of the Zoophycos ichnofacies (Savrda and Bottjer, 1989).
5.5 Shoreline and inshore facies associations and depositional process regimes

The five shoreline and inshore depositional environments are considered in more detail in relation to (1) facies types, (2) facies successions, and (3) depositional processes (Figure 5.3). Bioturbation intensity is described using the Bioturbation Index (BI) scheme of Taylor & Goldring (1993).

5.5.1 Wave-dominated shorefaces, delta fronts and estuaries

5.5.1.1 Description

Shoreface deposits consist of coarsening-upward successions of mudstones, siltstones and very fine- to medium-grained sandstones, which are locally capped by coal seams (Figure 5.3-5.7). 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 also contain swaley cross-stratification (SCS) (facies pLSF). The upper part of the succession consists of fine- to medium-grained cross-bedded sandstones with palaeocurrent directions approximately parallel to the palaeoshoreline 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 a variable and upward-decreasing diversity and intensity of bioturbation (BI=1-5) (Kiteley and Field, 1984; Van Wagoner et al., 1990; Kamola and Van Wagoner, 1995; Hampson and Howell, 2005; Hampson et al., 2011).

5.5.1.2 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 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 palaeo-water depths. Therefore this facies succession represents a prograding wave-dominated shoreface (W sensu Ainsworth et al., 2011).

5.5.1.3 Variation in facies motif

Three groups of observations provide evidence to diverge 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 chemico-physical stresses, such as sedimentation rate and sea-water salinity. 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, due to variations in river discharge. In areas away from river mouths, salinities were uniformly fully marine and deposition rates were uniformly low, resulting in a diverse suite of trace fossils and high bioturbation intensities (MacEachern and Bann, 2008). 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 palaeogeographical reconstructions it is interpreted that the coastline had a locally cuspate 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. Classically, these erosion surfaces have been interpreted in a sequence stratigraphic framework as sequence boundaries (Van Wagoner et al., 1990; Van Wagoner, 1991). However, these channels which cut down from the top of shoreface successions can also be interpreted to have been feeding the genetically related adjacent 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 provide evidence to
interpret the local influence of fluvial processes in wave-dominated deltas (Wf *sensu* Ainsworth et al., 2011).

In some locations, coastal plain deposits that correlate to wave-dominated shoreface successions contain tide-dominated channel-fills (Cole and Cumella, 2003; Kirschbaum and Hettinger, 2004; Gomez-Veroiza and Steel, 2010; Aschoff and Steel, 2011a), which have been interpreted to represent transgressed fluvial channels that effectively became estuaries when they were abandoned (facies TC, as described in the sub-heading ‘barrier island and back-barrier lagoons’) (Gomez-Veroiza and Steel, 2010). These transgressive, tide-dominated channel-fills can be attributed to a temporal and/or spatial change in depositional process in the context of an overall wave-dominated succession (Wft or Wtf *sensu* Ainsworth et al., 2011).

Figure 5.3. Wave-dominated shoreface, delta front and estuary deposits summarised in terms of: A) sedimentary logs for a prograding shoreface succession (capped by a flooding surface – blue line), B) inferred depositional processes, and C) plan view of the depositional environment. The Roman numerals in B and C indicate the following variations in depositional process and sub-environments: I) estuary, II) shoreface, and III) wave-dominated delta. This syntheis is compiled from this study and incorporates results from other workers (Kiteley and Field, 1984; Van Wagoner et al., 1990; Kamola and Van Wagoner, 1995; Hampson and Howell, 2005; Hampson et al., 2011). OS= offshore marine mudstone, dLSF= distal lower shoreface, pLSF= proximal lower shoreface, USF= upper shoreface, FS= foreshore, C= coal mire.
5.5.2 Fluvial-dominated deltas

5.5.2.1 Description

Fluvial-dominated delta successions consist of bioturbated mudstones, which coarsen upwards into medium- to coarse-grained sandstones (Figure 5.4) (Olariu et al., 2010; Hampson et al., 2011). Marine 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 (BI=0-4) and trace fossil diversity decreases to the presence of just *Ophiomorpha*, *Skolithos*, *Palaeophycus* 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 (Howell et al., 2008; Enge et al., 2010; Olariu et al., 2010). Locally, sharp-based, fine- to coarse-grained, structureless or cross-bedded sandstones are present in channelised and lenticular bodies at the top of these successions (facies DC) (Hwang and Heller, 2002; Olariu et al., 2010).

5.5.2.2 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 from turbidity currents. Beds with gradational bed-boundaries
record sustained waxing and waning currents, probably river-derived hyperpical flows (Olariu et al., 2010; Hampson et al., 2011). The occurrence of clinoforms indicates that deposition took place on relatively steeply inclined delta fronts (Howell et al., 2008; 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).

5.5.2.3 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). Mapping strata over large distances in the outcrop belts shows that wave-dominated shoreface and fluvial-dominated delta front deposits were coeval along the same palaeo-shoreline (Hampson and Howell, 2005; Hampson et al., 2011).
Figure 5.4. Fluvial-dominated delta front deposits summarised in terms of: A) sedimentary logs for a prograding delta front-mouth bar succession (capped by a flooding surface – blue line; erosive base of a distributary channels highlighted by the red line), B) inferred depositional processes, and C) plan view of the depositional environment. Schematic time lines (clinoform surfaces) are shown by the light grey lines. This synthesis is compiled from this study and incorporates results from other workers (Hwang and Heller, 2002; Howell et al., 2008; Enge et al., 2010; Olariu et al., 2010; Hampson et al., 2011). OS = offshore marine mudstone, dDF = distal delta front, pDF = proximal delta front, MB = mouth bar, DC = distributary channel.

5.5.3 Mixed tide- and wave-influenced deltas

5.5.3.1 Description

Tide-dominated or tide-influenced delta successions coarsen upward and consist of marine mudstones (facies OS) which become 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; Figure 5.5). These deposits are overlain across a low-relief erosional surface by cross-bedded fine-grained heterolithic sandstones showing bi-directional palaeo-current directions (facies LRTC). 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 of the succession, fine- to medium-grained sandstones with
high-relief (up to 20-30 m) channelised erosional bases of concave-upward geometry are present (facies HRCT). The cross-bedding with bidirectional palaeo-current directions of these sandstones contain mud drapes on their foresets and current-ripple cross-lamination in their heterolithic toesets. The latter three facies (LRTC, TB, HRTC) have a relatively low intensity (BI=0-3) and diversity in 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; Willis and Gabel, 2003; Legler et al., 2014; Van Cappelle et al., 2016; Burton et al., 2016).

### 5.5.3.2 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 (Figure 5.3). In contrast, the upper part of the succession (facies LRTC, TB, HRTC) is dominated by cross-strata that contain evidence of tides in the form of bidirectional palaeocurrents, pervasive mud drapes, and restricted trace fossil assemblages consistent with fluctuating salinity. 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 simultaneously in different parts of a mixed tide- and wave- influenced delta (*Ftw* or *Fwt* sensu Ainsworth et al., 2011; Legler et al., 2014). In this context, channelised cross-bedded sandstones with high basal erosional relief (facies HRCT) are interpreted as distributary channel-fills. These may have been overdeepened by fluvial erosion during falling relative sea-level or by tidal scour during abandonment and transgression (Willis and Gabel, 2003; Van Cappelle et al., 2016).
Figure 5.5. Mixed tide- and wave-influenced deltaic deposits summarised in terms of: A) sedimentary logs for prograding delta-front and lower delta-plain successions (capped by flooding surfaces – blue line; erosive bases of distributary channels, tidal channels and/or incised valleys highlighted by red lines), B) inferred depositional processes, and C) plan view of the mixed tide-depositional environment. The Roman numerals in A, B and C indicate different interpretation: Ia) shoreface and Ib) incised valley fill and II) mixed tide-and wave-influenced delta (Van Wagoner, 1991; Willis and Gabel, 2001; Willis and Gabel, 2003; Legler et al., 2014; Van Cappelle et al., 2016). OS= offshore marine mudstone, LSF= lower shoreface, LRTC= low relief tidal channel, TB= tidal bar, HRTC= high relief tidal channel.

5.5.4 Tide-dominated deltas

5.5.4.1 Description

In the northeast of the study area, some coarsening-upward sandstones are laterally (to the west and east) and vertically encased in mudstones, and thus appear to be detached from coeval shorelines. These so-called “isolated” sandstones (Bergman and Snedden, 1999) occur as elongate, linear bodies, usually oriented north-south or southwest-northeast, or as a series of such bodies that are laterally amalgamated to form more sheet-like geometries. Detailed mapping reveals that many of these “isolated” sandstones are connected at their northern extremity to time-equivalent shoreline deposits (Mellere and Steel, 1995; Mellere and Steel, 2000; Hampson et al., 2008). Bioturbated mudstones (facies OS) coarsen gradually upward into bioturbated, muddy, very fine- to fine-grained sandstones (faces DB/BT). 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.
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). 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 palaeocurrents (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 HSS (Boyles and Scott, 1982; Tillman and Martinsen, 1984; Gaynor and Swift, 1988; Walker and Bergman, 1993; Hampson et al., 2008). Locally, the erosional base of the uppermost facies has higher relief, forming channelised geometries (Mellere and Steel, 2000). Palaeocurrents 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 “isolated” sandstone bodies, and (3) oblique to the strike of inclined surfaces within the “isolated” sandstone bodies.

5.5.4.2 Interpretation

The upward decrease in mudstones content, 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 energy on the delta front and related (tide-influenced) distributary channels mouth bars. Deposition of cross-beds on inclined surfaces that are oriented oblique to cross-bed palaeocurrent directions indicates deposition on laterally accreting barforms (Boyles and Scott, 1982; Gaynor and Swift, 1988), which were elongated north-south, sub-parallel to the regional palaeoshoreline trend (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 toeset of these barforms (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 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. 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 al., 2008).

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). The first interpretation of the “isolated” sandstones was the shelf-ridge on a wave-dominated shelf model (Tillman and Martinsen, 1987; Tillman and Martinsen, 1984; Gaynor and Swift, 1988). It was interpreted that sand supply to these ridges came from the north. Modifications to this model included sand supply from reworked transgressed shoreline deposits (Boyles and Scott, 1982), which was supported by analogy with modern transgressive shelves, most notably the Atlantic shelf of North America (Swift and Field, 1981) (W sensu Ainsworth et al., 2011). However, there is no evidence for the presence of shoreline deposits in the direct vicinity of the “isolated” sandstones which could have been reworked, and there is no proof of high frequency, high amplitude sea-level variation during the Campanian such as there were during deposition of the Quaternary analogue. Modern shelf ridges are sand-dominated and contain shell-hash, quartz sand and mud clasts in their troughs, which suggests that they migrate over an erosional surface with a lag (Swift and Field, 1981). In contrast, the “isolated” sandstones in the WIS are heterolithic and gradationally coarsen upward from marine mudstones, and, although there is some evidence for wave-processes, their dominant sedimentary structures are tide-generated.
Alternatively, the “isolated” shallow-marine sandstones have been interpreted in a sequence stratigraphic framework (Van Wagoner et al., 1990) as tide-dominated, estuarine incised valley fills (Bergman, 1999) (Figure 5.6; Tfw sensu Ainsworth et al., 2011) that were eroded directly into marine mudstones during relative sea-level falls and lowstands. The north-south orientation of the interpreted valleys is inferred to follow subtle tectonically induced relief, possibly related to long-lived basement structures (Martinsen, 2003). However, incised valleys have erosional bases, which are not observed in any examples of “isolated” sandstones in the study area.

A third interpretation envisages the “isolated” sandstones as the preserved remnants of rugose, forced-regressive and lowstand shorelines from which thin coastal-plain deposits were removed by later transgressive erosion. Indeed, transgressive erosion is inferred to have detached the “isolated” sandstone bodies from other coeval shoreline deposits in some interpretations, although they remain attached in other interpretations. Shoreline types encompassed by this third interpretation include wave-dominated shorefaces (Walker and Bergman, 1993; Bergman, 1994; Bergman and Walker, 1995) and spits (Nielsen and Johannessen, 2008) (Figure 5.6; W sensu Ainsworth et al., 2011). The main drawback of this interpretation is the lack of evidence for wave-action as a dominant process and the inconsistency in facies character with wave-dominated shoreface deposits elsewhere in the WIS.

The most recent, and our preferred, interpretation is of tide-dominated deltas which built out southward (Mellere and Steel, 1995; Mellere and Steel, 2000; Hampson et al., 2008; Suter and Clifton, 1999) (Ftw sensu Ainsworth et al., 2011). This interpretation of a prograding delta is in agreement with 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). By analogy with the Yangtze Delta (Yang, 1989; Berné et al., 2002), it has been interpreted that the elongated coarsening upward successions in the WIS were deposited as ebb-tide-dominated mouth bars (‘tidal bars’) which prograded southward.
Figure 5.6. “Isolated” shallow-marine sandstones, which are interpreted as tide-dominated deltas, summarised in terms of: A) sedimentary logs for a prograding delta-front bar (capped by flooding surfaces – blue line; localised erosion near bar top highlighted by red lines), B) inferred depositional processes, and C) plan view of the depositional environment for. OS= offshore marine mudstone, DB/BT= distal bar/ bar toe, MB/BC= medial bar/ bar centre, PB/BT= proximal bar/ bar top.

5.5.5 Barrier islands and back-barrier lagoons

5.5.5.1 Description

Fluvial-dominated delta front and, more commonly, wave-dominated shoreface successions are overlain by thin successions of laterally continuous, heterolithic, carbonaceous mudstones, coals and lenticular, heterolithic sandstones which extend palaeo-landward of their associated, underlying coarsening-upward successions (Figure 5.7). 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). A common associated facies is sharp-based, lenticular and channelised, 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 palaeocurrent directions, which are locally organised on inclined surfaces. Bioturbation intensity is variable (BI=0-4) and trace fossils include Planolites, Palaeophycus, Ophiomorpha and
Thalassinoïdes (Donselaar, 1989; Devine, 1991; Olsen et al., 1999; Kamola and Van Wagoner, 1995; Hampson et al., 2011). Less commonly observed are sharp based, lenticular and channelised, 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 palaeo-seaward locations (facies TIC) (Donselaar, 1989; Olsen et al., 1999). Also found in palaeo-seaward locations are lenticular and convex-upward, coarsening-upward, heterolithic, very fine- to medium-grained sandstones. Sedimentary structures include bidirectional current-ripple cross-lamination and cross-beds that were deposited on clinoforms which dip palaeo-landward (facies FTD) (Kamola and Van Wagoner, 1995). In more palaeo-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 palaeo-seaward dipping clinoforms (facies BHD) (Donselaar, 1989; Olsen et al., 1999; Kamola and Van Wagoner, 1995).

5.5.5.2 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. 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 channelised sandstones at the top of wave-dominated shoreface successions (facies TI) are interpreted to be the result of reworking of shoreface sands in tidal inlets at the palaeo-seaward margin of the lagoon (Donselaar, 1989; Olsen et al., 1999). Coarsening-upward, lenticular and convex-upward sandstones containing palaeo-landward-dipping clinoforms (facies FTD) are interpreted as flood tidal deltas that were
genetically-related to tidal inlets (Kamola and Van Wagoner, 1995). Coarsening-upward successions that are richer in carbonaceous material and contain palaeo-seaward-dipping clinoforms, and which are found in more palaeo-landward locations (facies BHD), have been interpreted as prograding bayhead deltas (Donselaar, 1989; Olsen et al., 1999; Kamola and Van Wagoner, 1995). Fining-upward, channelised, inclined heterolithic sandstones with bidirectional palaeocurrent directions (facies TC) are interpreted as tidal channels (Donselaar, 1989; Devine, 1991; Olsen et al., 1999; Kamola and Van Wagoner, 1995). These channels could have been part of bayhead deltas (Donselaar, 1989) or the marine shoreline (Devine, 1991). Overall, these various facies formed in a more palaeo-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). Additionally there is evidence for fluvial processes in the form of bayhead deltas (Wtf or Wft sensu Ainsworth et al., 2011). 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 palaeo-seaward-lying barrier island successions.
Figure 5.7. Barrier island and back-barrier lagoonal deposits summarised in terms of: A) sedimentary logs (erosion surfaces at base of tidal inlets and tidal channels highlighted by red lines; transgressive surface as dashed blue line) B) inferred depositional processes, and C) plan view of the depositional environments (Donselaar, 1989; Devine, 1991; Kamola and Van Wagoner, 1995; Olsen et al., 1999; Hampson et al., 2011). TI= tidal inlet, FTD= flood tidal delta, LG= lagoon, BHD= bayhead delta, TC= tidal channel.

5.6 Timeslice reconstructions

Spatial and temporal variations in depositional processes are synthesised in five palaeogeographic timeslices (Figure 5.8-5.12). These build on previous palaeogeographic reconstructions (e.g. McGookey et al., 1972; Kauffman and Caldwell, 1993). Two isopach maps are also presented (Figure 5.13) based on published cross sections (Merewether et al., 1997; Molenaar et al., 2002; Anna, 2012). The coastal morphology and depositional processes along the palaeoshoreline are reviewed separately for each timeslice, with the palaeoshorelines described from south to north.

5.6.1 Timeslice 1 (83 Ma)

Timeslice 1 corresponds to the *Scaphites Hippocrepis* I biozone (83 Ma) (Figure 5.8). Shoreline deposits of the Point Lookout Sandstone in the San Juan Basin record overall northeastward progradation during the early Campanian (Molenaar et al., 2002). The Point Lookout Sandstone comprises both progradational wave-dominated shoreface and transgressive barrier-island and back-barrier lagoon systems (Devine, 1991)
Biostratigraphic dating in the Kaiparowits Plateau of south-central Utah is poor (Figure 5.2). 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 *Desmociaphites* 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).

In the Wasatch Plateau and Book Cliffs area of east-central Utah, the Star Point Sandstone and the lower part of the Blackhawk Formation are mapped in this timeslice. These strata contain progradational wave-dominated shoreface and fluvial-dominated (wave-influenced) delta front deposits, and accompanying transgressive barrier-island and back-barrier lagoon deposits (Van Wagoner et al., 1990; Kamola and Van Wagoner, 1995; Hampson and Howell, 2005; Hampson et al., 2011; Gani et al., 2015). Fluvial-dominated (wave-influenced) deltas co-existed lateral to wave-dominated strandplains, with fluvial-dominated (wave-influenced) deltas showing a southward-deflected, asymmetrical planform (Hampson et al., 2011; Forzoni et al., 2015). The latter is attributed to wave-generated longshore currents as evidenced by southward-directed palaeocurrents in adjacent wave-dominated shoreface successions (Slingerland and Keen, 1999; Hampson and Howell, 2005; 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 palaeocurrents measured at outcrop and subsurface mapping using well-log data (Johnson, 2003; Hampson, 2010). Gravity-flow siltstones and sandstones fed from the west have been correlated and linked to the position of river mouths in fluvial-dominated delta front successions or to 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), and we speculate that
they were supplied via rivers that fed now eroded shorelines between the Kaiparowits Plateau and the San Juan Basin.

In Wyoming, this timeslice correlates to progradational wave-dominated shoreface and delta front and transgressive barrier-island and back-barrier lagoon successions of the lower part of the Rock Springs Formation in the Rock Springs Uplift area (Roehler, 1978; Roehler, 1990) and the Eagle Formation in the Bighorn Basin (Fitzsimmons and Johnson, 2000).

5.6.2 Timeslice 2 (79 Ma)

Timeslice 2 corresponds to the Baculites sp. (smooth) biozone (79 Ma) (Figure 5.9). In the San Juan Basin, this timeslice is represented by wave-dominated shoreface and barrier-island and back-barrier lagoon successions of the Cliff House Sandstone (Figure 5.9) (Donselaar, 1989; Olsen et al., 1999; Molenaar et al., 2002). Subsurface mapping using well logs indicates that asymmetrical 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).

In east-central Utah and western Colorado, wave-dominated shoreface successions of the upper part of the Blackhawk Formation and lower Castlegate Sandstone were deposited (Van Wagoner, 1995; Yoshida, 2000; Hampson et al., 2008). These successions were fed by rivers of the Castlegate sandstone, and by wave-driven longshore currents that were predominantly southward-directed (Hampson et al., 2008; Hampson, 2010).

In northern Wyoming, the wave-dominated shoreface and barrier-island and back-barrier lagoon deposits of the Judith River Formation were deposited in the Bighorn Basin (Fitzsimmons and Johnson, 2000). In southern Wyoming, the Haystack Mountain Formation was deposited. The western exposures of this formation comprise wave-dominated shoreface successions (Mellere and Steel, 1995; Mellere and Steel, 2000), but further east the formation consists of tide-dominated deltas which are attached to wave-dominated shoreface sandstones (Mellere and Steel, 1995; Mellere and Steel, 2000). These shoreline-connected tide-dominated deltas correlate further south, in the Sand Wash Basin, to the seemingly “isolated” tide-dominated delta deposits of the Wise Gulch Sandstone, Berry
Gulch Sandstone and Morapos Sandstone (Boyles and Scott, 1982; Kiteley and Field, 1984; Hampson et al., 2008). Palaeocurrents in these sandstones are southwest-directed, sub-parallel to the regional palaeoshoreline trend. It has been interpreted that this palaeocurrent direction is the result of ebb-tide dominance in a southward prograding tide-dominated delta (Hampson et al., 2008; Suter and Clifton, 1999).

5.6.3 Timeslice 3 (77 Ma)

Timeslice 3 corresponds to the *Baculites Reduncus* biozone (77 Ma) (Figure 5.10). In the San Juan Basin, New Mexico, this timeslice encompasses the related regressive wave-dominated shoreface and transgressive barrier-island and back-barrier lagoon deposits of the Cliff House Sandstone and Pictured Cliffs Sandstone, which contain evidence for southeast-directed longshore currents (Figure 5.10) (Palmer and Scott, 1984; Donselaar, 1989; Olsen et al., 1999; Molenaar et al., 2002). In the Uinta Basin, east-central Utah, and Piceance Basin, west-central Colorado, the regressive, mixed tide- and wave-influenced deltaic deposits of the Sego Sandstone occur in the timeslice (Willis and Gabel, 2001; Willis and Gabel, 2003; Legler et al., 2014; Van Cappelle et al., 2016; Burton et al., 2016). In the Sand Wash Basin, northwestern Colorado, this timeslice is represented by wave-dominated shoreface successions of the Iles Formation (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 the Powder River Basin, Wyoming, the Parkman Sandstone has been interpreted to comprise wave-dominated shoreface deposits (Hubert et al., 1972). In northern Colorado, interpreted tide-dominated deltas of the apparently “isolated” Hygiene Sandstone are present (Griffitts, 1949; Kiteley and Field, 1984).

5.6.4 Timeslice 4 (75 Ma)

Timeslice 4 corresponds to the *Exiteloceras Jenneyi* biozone (75 Ma) (Figure 5.11). In the San Juan Basin, New Mexico, wave-dominated shoreface deposits of the Pictured Cliffs Formation were
deposited during this timeslice (Figure 5.11) (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. These deposits consist of wave-dominated shoreface successions (Kiteley and Field, 1984; Hettinger and Kirschbaum, 2002; Cole and Cumella, 2003; Kirschbaum and Hettinger, 2004; Gomez-Veroiza and Steel, 2010; Aschoff and Steel, 2011a; Madof et al., 2015). Barrier-island and back-barrier lagoon deposits have also been reported in the Sand Wash Basin (Masters, 1967). The time-equivalent coastal plain deposits to these various wave-dominated shoreface sandstones (Neslen, Mount Garfield and Iles formations) contain fluvial and tidally influenced fluvial channel-fill deposits (Hettinger and Kirschbaum, 2002; Cole and Cumella, 2003; 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 this timeslice have been eroded at the basal unconformity of the Teapot Sandstone (Gill and Cobban, 1966).

5.6.5 Timeslice 5 (73 Ma)

Timeslice 5 corresponds to the Baculites Compressus biozone (73 Ma) (Figure 5.12). In the San Juan Basin, New Mexico, wave-dominated shoreface successions of the Pictured Cliffs Sandstone were deposited (Tokar and Evans, 1993). In the Piceance Basin, west-central Colorado, wave-dominated shoreface successions of the Rollins Member of the Mount Garfield Formation were deposited (Hettinger and Kirschbaum, 2002; Cole and Cumella, 2003; Kirschbaum and Hettinger, 2004; Aschoff and Steel, 2011a; Madof et al., 2015). The Twentymile Sandstone in the Sand Wash Basin, northwestern Colorado, also contains wave-dominated shoreface and barrier-island and back-barrier lagoon successions (Masters, 1967). In western, central and northern Wyoming, the alluvial-and-coastal-plain sheet sandstones of the Teapot Sandstone were deposited above a tectonically forced unconformity (Gill and Cobban, 1966), which has been linked recently to the activity of Laramide-type
thrust faults (Leva López and Steel, 2015). The time-equivalent shoreline deposits are present in the subsurface of the western part of the Powder River Basin (Fox, 1993).

Figure 5.8. Palaeogeographic map for Timeslice 1: ammonite biozone Scaphites Hippocrepis I (~83 Ma) (Figure 5.2). The palaeoshoreline occupied its most western position and consisted mainly of wave-dominated coastlines and some fluvial-dominated deltas (Star Point Formation) 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 centre of the Western Interior Seaway.
Figure 5.9. Palaeogeographic map for Timeslice 2: ammonite biozone Baculites sp. (smooth) (~79 Ma) (Figure 5.2). The palaeoshoreline was predominantly wave-dominated, with a large tide-dominated delta interpreted in southern Wyoming and northwestern Colorado.
Figure 5.10. Palaeogeographic map for Timeslice 3: ammonite biozone Baculites Reduncus (~77 Ma) (Figure 5.2). During this timeslice, the palaeoshoreline was wave-dominated in the north and south, with mixed tide- and wave-influenced deltas of the Sego Sandstone interpreted in east-central Utah.
Figure 5.11. Palaeogeographic map for Timeslice 4: ammonite biozone Exiteloceras Jenneyi (~75 Ma) (Figure 5.2). During this timeslice, the palaeoshoreline was wave-dominated throughout.
Figure 5.12. Palaeogeographic map for Timeslice 5: ammonite biozone Baculites Compressus (~73 Ma) (Figure 5.2). During this timeslice, the palaeoshoreline was wave-dominated throughout.
Figure 5.13. Isopach maps for: A) lower Campanian strata (Timeslices 1 and 2; ammonite zones Scaphites Leei III to Baculites Asperiformis, 84-78 Ma; Figures 5.2, 5.8 and 5.9); and B) upper Campanian strata (Timeslices 3, 4 and 5; ammonite zones Baculites Perplexus to Baculites Eliasi, 78-72 Ma; Figures 5.2, 5.10, 5.11 and 5.12). The coloured lines with ages show the interpreted shoreline positions at maximum regression of each respective timeslice.

5.7 Discussion

5.7.1 Application of depositional process classification scheme

The classification of depositional processes and type of preserved coastal deposits is 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, in this case, 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), is problematic because a single sedimentary structure can be formed by several (co-occurring) depositional processes. 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 (cf. the process variability in barrier-island and back-barrier lagoons with shoreface successions).
Varying interpretations of the same deposits by different workers is another source of uncertainty in applying the depositional processes classification. 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 and storm-generated sand ridges (W *sensu* Ainsworth et al., 2011) and tide-dominated estuaries (Tf *sensu* Ainsworth et al., 2011). This wide variety of interpretations makes it difficult to define consistently and with confidence the corresponding ‘process space’ in the classification scheme. This problem is exacerbated where mixed depositional processes are recognised. The identification, correlation and interpretation of time-lines is another source for uncertainty. This can be illustrated in those cases where channels cut through different coastal systems, such as wave-dominated shorefaces, delta fronts and estuaries and mixed tide- and wave- influenced deltas. If the base of the channel system is a sequence boundary then there is a temporal separation between the non-channelised deposits below and the incised valley-fill above the boundary (Van Wagoner, 1991; Van Wagoner, 1995). However, if the channels are contemporaneous with the coastal/delta front successions, such as in deltaic systems, then the environment reflects mixed depositional processes (Van Cappelle et al., 2016; Hampson et al., 2008; Hampson, 2016). Hence, although we have applied the Ainsworth et al. (2011) classification scheme we also acknowledge its inherent limitations.

### 5.7.2 Spatial and temporal variation in depositional process regime

In all five timeslices (Figure 5.8-5.12), the northwest-southeast-striking palaeoshoreline in the south of the study area, in the San Juan Basin of New Mexico, was wave-dominated. The latter is evident in both regressive shoreface and transgressive barrier-island successions (Donselaar, 1989; Devine, 1991; Tokar and Evans, 1993; Olsen et al., 1999). Subordinate fluvial influence is recorded in two settings: (1) in the shoreface successions where localised active river mouths were associated with inferred wave-dominated deltas (Palmer and Scott, 1984; Devine, 1991), and (2) in bayhead deltas that were associated with back-barrier lagoon successions (Donselaar, 1989). Tidal influence is only interpreted in the transgressive barrier-island and back-barrier lagoon. In the north-south-striking
palaeoshorelines in the northern part of the study area, comprising the Great Divide, Wind River, Bighorn and Powder River basins of central and northern Wyoming, wave-dominated shoreface successions also dominate all five timeslices (Figure 5.8-5.12) (Hubert et al., 1972; Roehler, 1990; Fitzsimmons and Johnson, 2000). Hence, both the northern and southern parts of the study area were wave-dominated throughout the Campanian.

In contrast, a broader range of depositional processes is evident within the broad embayment (“Utah Bight”) that lay in the central part of the study area, including the Uinta, Piceance, Sand Wash, Washakie, Hanna and Laramie basins of Utah, Colorado and southern Wyoming (Aschoff and Steel, 2011b; Steel et al., 2012). In the oldest timeslice (Figure 5.8), when the palaeoshoreline lay in its most palaeolandward position, wave-dominated shoreface, fluvial-dominated delta front, and barrier-island and back-barrier lagoon successions were deposited (Kamola and Van Wagoner, 1995; Hampson and Howell, 2005; Olariu et al., 2010; Hampson et al., 2011). Subsequently, in the second timeslice (Figure 5.9), the palaeoshoreline was more embayed and contained two distinctive settings: (1) regressive wave-dominated shoreface and transgressive barrier-island and back-barrier lagoon successions (Van Wagoner, 1995; Yoshida, 2000; Hampson, 2010), and (2) tide-dominated deltas (Mellere and Steel, 1995; Mellere and Steel, 2000; Hampson et al., 2008). This timeslice records the first notable increase in tidal influence along the northern margin of the embayment. In the third timeslice (Figure 5.10), there are three distinct settings: (1) mixed-tide- and wave-influenced deltas (Sego Sandstone) in the Uinta and Piceance Basins (Willis and Gabel, 2001; Willis and Gabel, 2003; Legler et al., 2014; Van Cappelle et al., 2016), (2) tide-dominated deltas in the Middle Park Basin and Colorado Front Range (Griffitts, 1949; Gaynor and Swift, 1988) , and (3) an intervening area comprising wave-dominated shoreface successions in the Sand Wash Basin (Kiteley and Field, 1984; Gomez-Veroiza and Steel, 2010). The “Utah Bight” was most strongly embayed during this time (Figure 5.10). In the fourth and fifth timeslices (Figure 5.11, 5.12), the palaeoshoreline prograded eastwards, resulting in progressive infilling of the “Utah Bight” embayment. This resulted in a more linear shoreline characterised by wave-dominated shoreface successions along the entire length of the study area. However, coeval coastal
plain deposits continue to contain tidally influenced fluvial channel-fills (Masters, 1967; Kiteley and Field, 1984; Cole and Cumella, 2003; Kirschbaum and Hettinger, 2004; Gomez-Veroiza and Steel, 2010; Aschoff and Steel, 2011a).

5.7.2.1 Temporal and spatial variation in tidal influence

For the majority of the palaeoshorelines described above, most previous workers agree on the presence of evidence for tidal processes. In general, regressive and transgressive depositional environments with tidal influence are considered separately (Boyd et al., 1992). Increased tidal influence is observed along the more rugose shorelines of transgressive coastlines, most likely due to funnelling and amplification of tides within bays and estuaries (e.g. in abandoned, partially filled incised valleys). Such increases in tidal influence are localised, and do not reflect a basinwide change in depositional process (Steel et al., 2012). In the study area, such transgressive coastlines include barrier-island and back-barrier lagoon successions (Donselaar, 1989; Devine, 1991; Kamola and Van Wagoner, 1995; Olsen et al., 1999) and tidally-influenced channel-fills in coastal plain strata correlative to wave-dominated shoreface successions (Cole and Cumella, 2003; Kirschbaum and Hettinger, 2004; Gomez-Veroiza and Steel, 2010; Aschoff and Steel, 2011a). In both instances, the regional palaeoshoreline is interpreted to have been wave-dominated, and tidal influence is expressed locally in the form of tidal inlets, flood tidal deltas, lagoons and tidally-influenced fluvial channels on the coastal plain.

There are two types of regressive shoreline and inshore sandstones that display tidal influence: (1) mixed-tide- and wave-influenced deltas, and (2) tide-dominated deltas. Both latter depositional environments were only present when the coastlines were most embayed (timeslices 2 and 3) and suggest that a regional amplification of the tides occurred during these timeslices.

5.7.2.2 Potential tidal resonance in the “Utah Bight” embayment

When the “Utah Bight” had a pronounced embayed geometry (Figure 5.9, 5.11), tidal processes were prominent along its northern palaeoshorelines even during regression, implying that tides where
amplified on a regional scale. Tidal amplification takes place when the dimensions of the embayment are in resonance with the quarter wave-length of the tidal wave according to the following relationship (Allen, 1997):

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

(eq. 1)

In which \( l \) is the length of the gulf, \( T \) is the period of the tidal wave (12.42 hour for the principal semi-diurnal lunar tide \( M_2 \)), \( g \) is the gravitational constant (9.81 ms\(^{-2}\)), and \( h \) is the water depth. Estimations for the palaeo-waterdepth of the Cretaceous Western Interior Seaway are uncertain and range from 50 to 600 m (Erickson and Slingerland, 1990; Asquith, 1970). The higher-end estimates are from tectonically actively subsiding areas in the Maastrichtian (Asquith, 1970). A palaeo-waterdepth estimate of 50-300 m is more appropriate for the tectonically less active Campanian (Erickson and Slingerland, 1990). With these parameters, tidal resonance of the \( M_2 \) tide would have taken place in embayments with a length of 250-600 m. The length of the embayments in the wave-dominated timeslices 1, 2, 4 and 5 is ~200-275 km (Figure 5.8, 5.9, 5.11, 5.12). For the tide-influenced timeslice 3 (Figure 5.10), the embayment is 400-450 km long, which falls in the range for tidal resonance of the \( M_2 \) tide for palaeo-waterdepth ranges of 50-300 m. This potential for tidal resonance is in contrast with previous interpretations in which tides were locally amplified due to funnelling along coastlines with increased rugosity due to incipient Laramide restructuration, such that subsidence rates locally decreased, progradation rates increased and coastal rugosity increased (Aschoff and Steel, 2011b).

5.7.2.3 Structural constraints on the “Utah Bight” embayment

The overall position and regional orientation of the palaeoshoreline of the coastal margin of the Western Interior Seaway 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). During the early Campanian (Figure 5.14A), the northeastern limit of the “Utah Bight” palaeoshoreline protruded eastwards of the Wyoming Salient of the Sevier fold-and-thrust belt (Figure 5.8, 5.14A) (DeCelles, 1994; Yonkee and Weil, 2010). During the middle Campanian, Laramide structures within 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 shifted the depocentre in Wyoming further east (Figure 5.13B). However, Laramide structures were largely inactive in east-central Utah, where the depocentre remained in a western location, forming the deeply embayed “Utah Bight” (Figure 5.10, 5.14B). Later eastward progradation of the palaeoshoreline in Utah infilled the “Uah Bight” embayment, and coincided with a return to wave dominance (Figure 5.11, 5.13, 5.15C). The structural evolution described above can explain why there was an embayment with increased size and tidal resonance during the middle Campanian, but it does not explain why the presence of tide-dominated and tide-influenced regressive shorelines are limited to the northern margin of this embayment. The latter can be attributed to strong, wave-generated, southward-directed longshore drift along the western margin of the Western Interior Seaway (Palmer and Scott, 1984; Slingerland and Keen, 1999; Hampson and Howell, 2005; Hampson et al., 2011). As a result, leeward palaeoshorelines, which faced north, were subject to a larger wave-fetch and greater wave reworking than windward palaeoshorelines that faced south.

5.7.2.4 Modern analogues

The timeslice of the middle Campanian (Figure 5.14B) can be compared with the two modern coastal embayments: the German Bight and the Gulf of Guinea (Figure 5.15). The German Bight is a 250 km long embayment located at the southeastern margin of the North Sea. It is an epicontinental sea with a maximum water-depth of 150 m, which is distant from the shelf edge (c. 1000 km away). Although the coastlines along the German Bight are presently mainly transgressive as a result of the post-glacial Holocene transgression, there is a marked spatial change in depositional process (Nyberg and Howell, 2016; Van Straaten and Kuenen, 1957; Reineck and Singh, 1980). At the margins of the embayment, the western coast of the Netherlands and the northwestern coast of Denmark are wave-dominated shorefaces. At the northern coast of the Netherlands and the southwestern coast of Denmark, in positions towards the embayment centre, barrier-islands with tidal inlets and back-barrier lagoons are present, marking an increase in tidal influence. In the German Bight, at the centre of the embayment,
the tide-dominated estuaries of the Ems, Weser and Elbe rivers are present (Reineck and Singh, 1980). This area represents the drowned remnants of a Late Pleistocene/Early Holocene delta, which preserves the characteristic elongate sand body morphology of tide-dominated deltas (Reineck and Singh, 1980). The dimensions, facies and thickness patterns preserved in the German Bight are reminiscent of the Utah Bight (timeslice 2).

Unlike the study area, the Gulf of New Guinea is not entirely situated on the continental shelf, but it has a similar spatial scale to the “Utah Bight” (Figure 5.15). At the margins of the gulf, wave-dominated carbonate barrier reefs with tidal inlets are present in both Papua New Guinea and Australia. In locations towards the centre of the embayment along the coast of Papua New Guinea, the coastline is a wave-dominated siliciclastic shoreline on the shelf. At the centre of the gulf tide-dominated estuaries and the tide-dominated delta of the Fly River are present (Harris et al., 1993; Harris et al., 1996; Dalrymple et al., 2003). The Gulf of New Guinea differs from the study area in the sense that carbonate systems are present in part of the gulf, and that the gulf is partially at the shelf edge.

The spatial variation in depositional process of all three examples is similar. At the margins of the embayments, wave-dominated shorefaces are present. Towards most landward indented position of the embayments, tidal influence of the depositional environments increases, such that tide-dominated estuaries and tide-dominated deltas are developed.
Figure 5.14. A) Early Campanian: wave-dominated palaeoshorelines were widespread (cf. Figs. 9, 14A), with the larger-scale geometry controlled by the position of the Sevier fold-and-thrust belt and the Mogollon Highlands. The large “Utah Bight” embayment is bordered to the north by the protrusion of the palaeoshoreline in front of the Wyoming Salient. B) Middle Campanian: active Laramide structures in Wyoming forced the palaeoshoreline further eastward and caused the “Utah Bight” to become longer, which may have increased tidal amplification (cf. Figs. 5.10, 5.13B). C) Late Campanian: continued eastward palaeoshoreline progradation, especially in the central area, reduced the significance of the “Utah Bight”, which may have decreased tidal amplification and resulted in a return to more uniformly wave-dominated shorelines (cf. Fig. 13).
Figure 5.15. Same scale comparison of the spatial variation in depositional process of A) the deposits of the middle Campanian of the Western Interior Seaway, B) the Gulf of New Guinea, and C) the German Bight (modified after Nyberg and Howell, 2016; Bing Maps, 2016).
5.8 Conclusions

Published sedimentological 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 synthesised to define five different palaeoshoreline types, each with a distinctive depositional process regime: (1) wave-dominated shorefaces, delta fronts and estuaries; (2) fluvial-dominated delta fronts; (3) mixed-tide- and wave-influenced deltas; (4) tide-dominated deltas; and (5) barrier island and back-barrier lagoons. The distribution of these shoreline types has been compiled for five timeslices, which illustrate the spatial distribution of depositional processes along the southwestern margin of the seaway, as preserved in the stratigraphic record, together with syn-depositionally active structural elements.

The position of the Western Interior basin margin and associated palaeoshoreline in the study area 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 embayment of the shoreline (“Utah Bight”) was variably developed as a result.

- Early Campanian palaeoshorelines along the “Utah Bight” embayment were wave-dominated, except for fluvial-dominated deltas developed at local points of sediment input.
- The onset of Laramide thrusting in the foreland of Wyoming during the middle Campanian caused eastward progradation of palaeoshorelines north of the “Utah Bight”. However, the position of the palaeoshorelines 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 semidiurnal tide. Palaeoshorelines 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 palaeoshoreline in the study area, which was dominated by wave action along its entire length.

The synthesis and analysis of sedimentological data presented here illustrates an approach that could be widely applied to other basins, although the resulting interpretations of controls on depositional process regime will likely vary between different 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 (cf. hydrocarbon play scale), which is probably inherent to the application of any shoreline classification scheme (i.e. irreducible uncertainty). When this sedimentological 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.
Chapter 6.  Synthesis

In this chapter the facies and the facies architecture of the lower Sego Sandstone (chapter 3) and the Ile Formation (chapter 4) are compared and contrasted. The common properties of these two deposits have been distilled out and compared and contrasted with other case studies and models of tide-influenced and tide-dominated models. The features of all these examples have been compiled in order to create a more universal applicable facies model for tide-influenced deltas. It is also shortly discussed in what kind of tectonic settings tide-influenced deltas can be expected to form (chapter 4 and 5).

6.1 Comparison between the lower Sego Sandstone and the Ile Formation

Here the facies of the tide-influenced parts of the Sego Sandstone and the Ile Formation are compared. The Sego Sandstone can be sub-divided in a wave-dominated lower part and a tide-dominated upper part. The lower part consists of coarsening upward successions (S₁, S₀ and S₁ in Chapter 3) of offshore mudstones (FA1 in chapter 3), lower shoreface (FA2) and inter-channel tidal bars and tidal channel fills (FA4). The upper part consists of alternations (S₂ and S₃) of coarsening upward tidal bars (FA3), inter-channel tidal bars and channel fills (FA4) and abandoned distributary channels (FA5). It has been interpreted that the wave-dominated lower part and the tide-dominated upper part are constituents of one single mixed-tide- and wave-influenced deltaic system.

However, in the Ror, Tofte and Ile formations, the coarsening upward successions (progradational phases 1 & 2 in chapter 4) have been interpreted to form part of a different depositional system than aggradational phase 3 due to a temporal change in depositional process. Where progradational phases 1 & 2 exhibit mixed wave- and fluvial influence and only little evidence for tidal reworking is present, the facies in aggradational phase 3 show abundant evidence for tidal reworking. Therefore, only the following facies association from aggradational phase 3 will be compared with the Sego Sandstone: prodelta and offshore mudstones (FA1); fluvial channel-fill sandstones (FA3); tide-influenced fluvial
channel-fill sandstones (FA4); tidal channel-fill heterolithic sandstone (FA5); tidal coastline heterolithic sandstones (FA6); coal mire deposits (FA7); and lower shoreface sandstones (FA8) (Figure 6.1).

6.1.1 Grain size comparison

On a first order comparison, both the Sego Sandstone and Ile Formation are dominated by fine grained sandstone and can therefore be classified as fine grained sandstone deltaic deposits (Orton and Reading, 1993). Clay and silt are relatively rare and are only found in offshore marine environments and as mud drapes, mud clasts and fluid mudstones, forming generally a much smaller volume fraction than the sandstone. However the maximum grain size of the Sego Sandstone is lower medium grained sand. In contrast, fluvial channel-fills in the Ile Formation contain pebbly coarse to very coarse grained sandstones. This difference in grain size has an effect on the formation of sedimentary structures (Orton and Reading, 1993). For example, HCS only forms in silt, very fine and lower fine grained sand (Yoshida et al., 2007), current ripples do not form in coarse sand, dunes do not form in silt and very fine sand (Van den Berg and Van Gelder, 1993), and evidence for tidal currents is poorly preserved in coarse-grained deposits (Dashtgard and Gingras, 2007).

6.1.2 Vertical facies stacking

The most striking similarity between the Sego Sandstone and the Ile Formation is the alternation of sharp based, fining upward channel-fills with a relatively low mudstone content and relatively coarse sand grain size (Figure 6.1; FA4 and FA5 in the Sego Sandstone; FA3, FA4 and FA5 in the Ile Formation), and relatively mudstone-rich heterolithic sandstones with a relatively fine grain size (Figure 6.1; FA3 in the Sego Sandstone; FA6 in the Ile Formation).

6.1.2.1 Sharp based, fining upward channel-fills with a relatively low mudstone content

In the following paragraph, the sharp based, fining upward channel-fills with a relatively low mudstone content facies associations of the Sego Sandstone are summarized. In the paragraph thereafter, the sharp based, fining upward channel-fills with a relatively low mudstone content facies associations of
the Ile Formation are re-iterated and directly compared and contrasted with facies associations of the Sego Sandstone.

The channel-fill deposits in the Sego Sandstone (FA4) are separated in two groups based on their cross-sectional geometry: 1) planar, low relief, 1-10 m thick, 1-25 km wide sandstones bodies with bidirectional to flood-tide-dominated palaeocurrent directions, and 2) lenticular, high relief, 1-15 m thick, narrow (~100-300 m), elongated (2-4 km) sandstone bodies with fluvial down-stream to ebb-tide directed palaeocurrent direction. The first group is interpreted as tide-dominated channel-fills which preserve on lateral accreting mid-channel bars. The second group is interpreted as tide-influenced distributary fluvial channel-fills deposited on downstream and laterally accreting bars in a relatively stable distributary channel (Figure 6.1). Fluvial-dominated channel-fills are absent in the Sego Sandstone because even the distributary channels contain evidence of tide-influence.

In the following paragraphs, the sharp based, fining upward channel-fills with a relatively low mudstone content facies associations of the Ile Formation are summarized and directly compared and contrasted with facies associations of the Sego Sandstone. The Ile Formation contains three types of channel-fill successions. FA5 of the Ile Formation contains mud-draped cross bedding with heterolithic, current-ripple cross-laminated toesets and bidirectional current-ripple cross-laminated sandstones higher in the succession which were probably deposited on shallowly inclined surfaces. Their tide-dominated facies character leads to their interpretation as tide-dominated channel-fill deposits and a close resemblance to facies FA4 of the Sego Sandstone (Figure 6.1).

Channel-fills of FA4 of the Ile Formation contain some evidence for tidal currents in the form of mud-drapes, and evidence for higher salinity marine waters in the form of bioturbation. Therefore FA4 is interpreted to be deposited in a tide-influenced fluvial channel-fill. These sand bodies may be similar to the lenticular, high relief elements of FA4 in the Sego Sandstone. However the different facies character may be largely due to the different texture of the available sediment (Dashtgard and Gingras, 2007).
FA3 of the Ile Formation does not contain evidence for tidal currents, and has therefore been interpreted as fluvial distributary channel-fill sandstones. The Sego Sandstone does not have a facies analogue to this, but deposits of the middle Castlegate Formation which were deposited on the coeval lower delta plain do contain fluvial channel-fills.

6.1.2.2 Relatively mudstone-rich heterolithic sandstones

The relatively mud-rich heterolithic deposits, with their finer sand grain size in both units (Figure 6.1; FA3 in the Sego Sandstone; FA6 in the Ile Formation) are interpreted to have been deposited on the lower delta plain due to their close association, and alternated stacking, with channelised facies. However, no evidence for subaerial exposure is present, except for coal mire deposits in the Ile Formation (FA7) which are locally present in the uppermost Ile Formation on relatively slowly subsiding footwalls and horst blocks. In both case studies, there is abundant evidence for tidal currents by way of bidirectional current-ripple cross-lamination and the heterolithic nature of the deposits, and brackish marine salinities based on bioturbation. Additionally, FA6 of the Ile Formation contains HCS, which is evidence for storm-wave reworking. FA3 of the Sego Sandstone is interpreted to have been deposited in abandoned distributary channels/ estuaries. FA6 of the Ile Formation is interpreted to reflect deposition in sub- to inter-tidal parts of tidal flats. These environments are similar and reflect the dominance of laterally-accreting tidal channel point bars within the tidal flats (Van Straaten and Kuenen, 1957; Reineck, 1958; Choi et al., 2004).

6.1.2.3 Offshore mudstones and wave-influenced coarsening upward successions

The coarsening upward successions of offshore mudstones (FA1), shoreface deposits (FA2) and tidal channels fills (FA4) of the Sego Sandstone are not present in the Ile Formation. However, the tide-dominated aggradational phase 3 of the Ile Formation may possibly correlate to offshore (FA1) and shoreface (FA8) deposits in well 6407/11-1S. Although these deposits can be interpreted as having been locally sourced around tilted fault-blocks (Ravnås and Steel, 1998), to be alongshore equivalents of the Ile Formation to the south, or to represent a change in process during progradation (Ta et al.,
2002; Olariu, 2014), it is also possible that these offshore (FA1) and shoreface (FA8) deposits represent a spatial variation in process. In the latter case, deposition of wave-dominated facies took place in the distal part of mixed tide- and wave-influenced deltas at the point of furthest progradation. With the latter interpretation an analogy can be made with the Sego Sandstones, where tide- and wave processes interacted in the distal reaches of the delta (Legler et al., 2014), and with the Niger delta, where the delta front consists of wave-dominated shorefaces, while the lower delta plain is largely tide-dominated except for the distributary channels (Oomkens, 1974).
Figure 6.1. Table with a comparison between the facies of the Sego Sandstone and the Ile Formation (sst=sandstones, mst=mudstones, vf=very fine, f=fine, m=medium, c=coarse, vc=very coarse, fm=formation). Above the table, example logs of the Sego Sandstone (left, Figure 3.7) and Ile Formation (right, Figure 4.7) are re-used for comparison between the logs.
6.2 Facies model for mixed-tide- and wave-influenced deltas

A facies model for mixed-influence deltas, and in particular the fluvial-to-tidal transition will be presented here. It builds on the observations and conclusions taken from the Sego Sandstone and the Ile formation, but also takes in account elements of other examples from literature on ancient and modern analogues. Firstly, a description of a typical distal to proximal transect is presented, with offshore mudstone deposition in the most distal parts of the system, wave-reworking at the delta front, tidal erosion and re-deposition within the lower delta plain, and the upstream transition to purely fluvial sedimentation in the upper delta plain (Dalrymple and Choi, 2007). Later, the stacking of these facies during progradation with different shoreline trajectories will be discussed. Also, lateral variability due to avulsion will be discussed.

6.2.1 Distal to proximal facies successions in mixed-influence deltas

During initial progradation, prodelta mudstones are deposited (Figure 6.1). The main mode of deposition is settling from suspension and deposition from gravity flow deposits. Gravity flow deposits in prodelta successions are usually interpreted to be driven by fluvial input, and therefore classified as fluvial dominated (Ainsworth et al., 2011; Ainsworth et al., 2015). Wave and tidal processes are only expected to be present above storm wave base and fair-weather wave base respectively. Prodelta deposits below storm-wave base are by definition too deep for currents originating at the sea-surface to interact with the sea-bed. Nevertheless some weak wave-influence can be expected when a storm and gravity flow coincide, and the top of gravity flow deposits are affected by some wave-action during the waning phase, resulting in some combined flow-ripples (De Raaf et al., 1965; Plint, 2014).

Where wave-processes are absent, prodelta mudstones coarsen upward into tide-dominated heterolithic deposits. By definition, river deltas are influenced by rivers, so also fluvial influence is to be expected. These type of deposits with the absence of wave-processes are rare. Ancient examples clearly showing this are the Eocene Dir Abu Lifa Member in Egypt (Legler et al., 2013), the Jurassic Tilje Formation offshore Norway (Martinius et al., 2001; Martinius et al., 2005; Ichaso et al., 2016) and the
Jurassic Lajas Formation in Argentina (McIlroy et al., 2005; Rossi and Steel, 2016). A modern example with cores to constrain vertical facies successions is the Fly delta in Papua New Guinea (Dalrymple et al., 2003). All these deltas were deposited in long embayments that protected the delta front from wave-action. The embayments can be structurally controlled, representing fault-bounded areas of rapid subsidence (Wonham et al., 1997; Carr et al., 2003; Jackson et al., 2005; Legler et al., 2013).

Where wave-processes operate, prodelta mudstones are interbedded with storm event beds deposited between fair-weather and storm wave-base. Deposits of tidal currents are unlikely to be preserved here, because deposition takes place below fair-weather wave-base. This means that during normal condition, tidal currents are not strong enough to rework the sea-bed. However, during storms, (mixed tide-) wave generated storm-surges do rework the sea-bed and generate storm beds (Dashtgard et al., 2012; Vakarelov et al., 2012). It is likely that at the coastline, also shallower depositional environments such as upper shoreface and foreshore environments are present. In modern mixed-tide- and wave-influenced deltas such as the Mekong delta (Nguyen et al., 2000; Ta et al., 2002) and Niger delta (Allen, 1964; Allen, 1965b; Allen, 1970; Oomkens, 1974; Reijers, 2011), many locations at the delta front show a shallowing upward profile typical of a wave-dominated shoreface. However, especially the upper part of the shoreface successions have a low preservation potential because during delta progradation, erosion and re-deposition in tidal channels will rework the sediment above fair-weather wave-base, such as is interpreted in the Sego Sandstone (Legler et al., 2014).

At shallower water-depths, above fair-weather wave base, and in more land-locations, where deposition is protected from wave-reworking due to a limited wave-fetch, tidal currents rework the sediment continually. Erosion takes place, and the channels thus formed will be filled by tide-dominated facies. This can either occur on laterally accreting longitudinal bar-forms in the centre of the channel, or as inclined heterolithic strata (IHS) deposited on laterally accreting point bars in more landward positions (planar deposits of FA4 of the Sego Sandstone; FA5 of the Ile formation) (Reineck,
In these relatively distal locations close to the shoreline, ebb and flood tides are usually asymmetrical, with the flood tide being of shorter duration and stronger in flow velocity, therefore, flood-tidal palaeocurrent directions are more likely to be preserved (Johnson et al., 1982). Farther palaeolandward, towards the apex of the delta, the tidal energy decreases and fluvial currents become more dominant (Dalrymple and Choi, 2007). Therefore, ebb-tide directed palaeocurrents are more likely to be preserved in this location. The position of these distributary channels are more stable, so the channels are narrower and higher relief (lenticular, high relief FA4 in the Sego Sandstone; FA4 in the Ile Formation). Eventually, the tidal range diminishes upstream and purely fluvial channels are present in the most proximal parts of the delta (FA3 in the Ile Formation). The length-scale along which the fluvial-to-tidal transition takes place depends on ratio between the magnitude of the fluvial and tidal currents, and the seasonal variation thereof (Dalrymple and Choi, 2007; Dalrymple et al., 2015b; Gugliotta et al., 2016).
Figure 6.2. (A) Downstream variations in depositional process regime along a tide-influenced river delta, from fluvial-dominated upstream, to tide-dominated further sea-ward to wave-dominated at the coastlines (modified after Dalrymple and Choi, 2007). (B) Simplified map of a mixed-influence delta and the relative location of deposition of the facies in Figure 6.1 (same colour coding as the central column) with respect to the main depositional process at the point along the downstream transect of (A). (C) Down-stream grain-size distributions. Up to pebbly medium to coarse sand is deposited in fluvial channels on the upper delta plain. In the lower delta plain tidal currents keep mud in suspension and deposit a large
proportion of mud during slack water. At the coastline continuous wave-action is keeps mud in suspension, so the sand deposited here is relatively clean (modified after Dalrymple and Choi, 2007). (D) Cross-section along the same transect indicating the relative vertical position of the facies from Figure 6.1, also with respect to fair-weather and storm wave-base.

Figure 6.3. Model for facies preservation of tide-influenced deltas. For facies colour coding see Figure 6.1. A) With a flat shoreline trajectory the more proximal facies (i.e. distributary fluvial channels) cut through previously deposited distal facies (i.e. tidal channels) and tidal flat deposits deposited on the delta plain in shallow (inter-tidal) palaeowater-depths have a very small preservation potential. B) With a rising shoreline trajectory the more proximal facies (i.e. distributary fluvial channels) cut through previously deposited distal facies (i.e. tidal channels) just as with a flat shoreline trajectory. However, now there is a larger preservation potential for tidal flats due to aggradation of the lower delta plain. Note that in both models just a simple proximal-distal relation is used. With distributary fluvial channels and abandoned distributary channels co-occurring laterally, a more complex facies stacking of more tide-influenced facies is the result.

6.2.2 Vertical and lateral stacking facies in a mixed-influence delta

The position of facies belts in both wave-dominated and tide-dominated transgressive estuaries is relatively stable (Dalrymple et al., 1992; Dalrymple and Choi, 2007). A reason for this is that tide-dominated estuaries are often deposited within incised valleys which dictate the position of the estuary (Dalrymple et al., 2006). However, in progradational systems, the focus of deposition advances
to more distal locations over time, and hence, the positions of all the facies described above advance basinward. When the shoreline progrades with a descending shoreline trajectory with an angle equal to the slope of the lower delta plain, no net-aggradation of the lower delta plain takes place (Prince and Burgess, 2013). The result is that fluvial and tidal channels all cut down from the same stratigraphic level, causing a relatively thin and poorly preserved delta plain deposit. An example of this architecture is shown in the conceptual model of Figure 6.2C. Real examples of this architecture are the first progradation in the Upper Cretaceous Drumheller Member (Alberta, Canada) where tidal channels and distributary fluvial channel cut down from the same level (lowest 50m in Figure 6.4C; Ainsworth et al., 2015) and the Holocene (most recent) progradational succession of the Niger delta where tidal channels and fluvial channels also cut down from the same (modern day land-) surface (Figure 6.4B; Oomkens, 1974). However, when the shoreline trajectory is ascending or descending with an angle less than the slope of the lower delta plain, aggradation does take place on the lower delta plain (Prince and Burgess, 2013). In this case, more proximal facies will be deposited at a progressively higher stratigraphic level during progradation. This will result in a thicker interval of channelised and non-channelised delta plain facies, with more proximal facies occurring at stratigraphic higher positions. An example of this architecture is provided by the Sego Sandstone as described in chapter 3, where the proximal, high relief distributary channels are concentrated near the top of the lower Sego Sandstones (}
Deltas are laterally variable. In certain cases such as the Fly river delta (Figure 3.12A), water and sediment flux from the apex of the delta is distributed over all the channels in the delta, feeding all parts of the delta front more or less equally. In this case there will be relatively little along strike variability, but just the proximal to distal variability along depositional dip outlined above (Dalrymple et al., 2003; Harris et al., 2004). However when avulsion of the fluvial feeder channel takes place, delta lobes will be abandoned and switched to new delta lobes. Therefore vertical and lateral stacking of facies deposited in active and inactive delta lobes takes place (Coleman, 1988; Correggiari et al., 2005). Channels feeding active delta lobes will exhibit evidence for more pronounced fluvial processes. On the other hand, abandoned channels will effectively become estuaries with more pronounced tidal processes (Allen and Chambers, 1998; Salahuddin and Lambiase, 2013). Also, on the delta plain between channels, non-channelised heterolithics deposited on tidal flats and tidal creeks will be present. A good example of this architecture is provided by the Mahakam river delta (Figure 3.12A). In the Mahakam delta, four main distributary channels are present which are fluvial-dominated, with increasing tide-influence downstream (Salahuddin and Lambiase, 2013). These distributary channels feed mouth bars at the delta front and cause delta progradation (Allen and Chambers, 1998). However, lateral to the active distributary channels, inactive tide-dominated estuaries are present. Therefore, the facies deposited in these environments will be preserved with a laterally and vertically stacked architecture.

6.3 Comparison with other and case studies models

Here the model presented in the previous sub-heading and the deposits of the lower Sego Sandstones and the Ile Formation have been compared one-on-one to other: (1) case studies of mixed-tide- and wave-influenced deltaic deposits; (2) case studies of tide-dominated deltaic deposits; and (3) models for tide-dominated deltas. The latter two comparisons put mainly emphasis on the stacking of channelised and non-channelised deposits on the lower delta plain (Figure 6.4).
6.3.1 Modern and ancient mixed-tide- and wave-influenced deltas

In Figure 6.4A-D the lower Sego Sandstone and the Ile Formation are compared to deposits of the Holocene Niger Delta (Nigeria) and the Cretaceous Drumheller Member (Alberta, Canada). The delta front of the Niger delta consists mainly of wave-dominated shorefaces. Where distributary channels reach the shoreface, small mouth bars of medium to coarse sand are deposited (Allen, 1965b; Allen, 1970). Longshore currents efficiently redistribute the sediment delivered through the distributary channels along the shoreface. Therefore the shoreline is wave-dominated and fluvial influenced. Besides distributary channels, abundant tidal channels are present which either branch from fluvial-dominated distributary channels or connect directly to sea through the shoreface (Allen, 1965b; Allen, 1970). Both fluvial and tidal channels cut down from the modern land-surface and are up to 20 m deep (Allen, 1965b). The deposits of the tide-dominated channels are up to 20 m thick as well, and form single story-multilateral sand-body complexes which cut down into offshore mudstones (Figure 6.4B; Oomkens, 1974). In more landward position, coarser grained, fluvial dominated channels are present (Oomkens, 1974). The tidal channels are only present in the lower delta plain, a belt parallel to the coastline. In Figure 3.12A the lower delta plain with tidal channels is labelled ‘mangrove swamps and mudflats’. In more proximal locations labelled ‘delta plain’ and ‘flood plain’, deposition is dominated by fluvial currents (Figure 3.12A; Allen, 1965, 1970). Due to the mix of fluvial distributary channels and mouth bars, tidal channel and a wave-dominated shoreface, the Niger Delta is classified as a mixed-wave-, tide- and fluvial influenced delta.

Compared to the model presented in Figure 6.2, the main similarity with the Niger delta is the separation between the wave-dominated shoreface and the coastlines in distal parts, and the abundance of tidal channels with a single-story multilateral architecture in the proximal parts. Another similarity is the preservation potential of the wave- and tide-dominated facies. After the Holocene progradation of the Niger Delta, wave-dominated shoreface deposits are only preserved at the delta front. During progradation, erosion of the tidal channels has eroded much of the shoreface successions. This is similar to the model presented in Figure 6.2. This part of the model is mainly based
on the lower Sego Sandstones where there is a preservational bias as well with channels eroding shoreface deposits during progradation (chapter 3; Legler et al., 2014). With respect of the stacking of fluvial-dominated and tide-dominated channels, the lower delta plain of the Niger Delta is dominated by tidal channels. However, some coarser grained fluvial dominated channels extend through the lower delta plain to the shoreline (Allen, 1965b; Oomkens, 1974). This pattern is also integrated in the lateral stacking in the model of Figure 6.2. In the model, fluvial-dominated distributary channels extend to the shoreline, while tide-dominated channels are present in lateral positions. This part of the model is mainly based on the Ile Formation where relatively rare isolated coarse-grained, fluvial-dominated, tide-influenced channels (FA4) are present in a succession which is tide-dominated elsewhere (FA5 and FA6).

The model (Figure 6.2) is also compared with the Cretaceous Drumheller Member (Alberta, Canada; Figure 6.4C; Ainsworth et al., 2015). The first coarsening upward succession of the Drumheller Member consists of wave-dominated shoreface deposits which are erosionally overlain be a low relief, lateral extensive tide-dominated sandstone body. In a proximal location, a high-relief, narrow, fluvial-dominated sandstone body is present. Ainsworth et al. (2015) interpret the base two erosional surfaces as sequence boundaries according to the model for tide-dominated deltas of Willis (2005). However, these deposits can also be interpreted as the continued progradation of a mixed-tide- and wave-influenced delta according to the model presented in Figure 6.2.

6.3.2 Modern and ancient tide-dominated deltas

In Figure 6.4E-G some cross-sections through tide-dominated deltaic deposits are shown for comparison. Firstly, the Holocene deposits of the Fly River Delta (Figure 6.4F) consist of offshore mudstones which gradually coarsen upward (Harris et al., 1993; Dalrymple et al., 2003). Both the sand grain size increases and the mud-content decreases. This coarsening upward delta front is bioturbated and it is gradually overlain by well sorted, fine grained mouth-bar deposits containing cross-bedding and rare HCS is evidence for some wave-influence. Such tide-dominated coarsening upward
successions are indicated in purple in Figure 6.4. The mouth-bar successions are deposited in sea-ward positions of the coast-normal bars with an emerged bar-top. These coarsening upward successions are eroded out from the top by channels with a coarse-grained (sometimes shell hash) channel-lag which can be filled by: (1) fining upward fluvial-dominated channel-fill successions; or (2) coarsening upward ‘tidal bars’ deposited on shore-normal, elongated mid-channel or side-channel bars (Dalrymple et al., 2003). All mouth-bar and channel-fill deposits are gradually overlain by relatively thin, fining upward tidal-flat deposits which are rooted from the top.

The deposits of the Eocene Dir Abu Lifa Member (Western Desert, Egypt; Figure 6.1E; Legler et al., 2013) have similar facies and a similar facies architecture to the deposits of the Holocene Fly River Delta. These deposits also consist of tide-dominated coarsening upward successions of offshore mudstones and delta front deposits. These coarsening upward successions are cut out from the top by channels which are filled by inclined heterolithic sandstones and mudstones with an upward decrease in sand grain-size and mud-content. These successions are overlain by thin tidal flat deposits. In the Dir Abu Lifa Member, two of such successions are stacked, and the channels of the upper succession have eroded out the upper part of the lower succession.

Comparing the tide-dominated deposits of the Fly River Delta and the Dir Abu Lifa Member to the mixed-tide- and wave-influenced deposits, the main difference is that the facies of the delta front are different. Where the tide-dominated delta fronts consists of heterolithic (bioturbated) deposits which coarsen upward in cross-bedded successions with perhaps rare HCS (Dalrymple et al., 2003; Hampson et al., 2008), the coarsening upward succession of the mixed-tide- and wave-influenced deltaic deposits consists of shoreface successions dominated by HCS. The main similarity is that the dissimilar coarsening upward successions are eroded out by wide, low-relief erosional surfaces which are infilled by tide-dominated channels.

The stacking of channelised and non-channelised deposits on the delta-plain can be compared with Devonian deltaic deposits from the Baltic Basin (Pontén and Plink-Björklund, 2007; Tänavsuu-
Milkeviciene and Plink-Björklund, 2009; Plink-Björklund, 2012). The tide-dominated deltaic deposits of the Aruküla and Kukliai formations consist of tide-dominated coarsening upward successions of offshore mudstones which coarsen upward into heterolithic deposits and cross-bedded sandstones with abundant evidence for tidal currents (Figure 6.4G). Although some low-relief erosional surfaces are present locally, no clear distinction between the deposits of the coarsening upward succession and the channel-fills such as in the Fly River Delta and the Dir Abu Lifa Member are present. This absence or at least scarcity of channelised facies is similar to the tide-dominated “isolated-sandstones” in the Western Interior Seaway described in chapter 5 (Mellere and Steel, 1995; Mellere and Steel, 2000; Hampson et al., 2008). The younger Gauja Formation contains some tide-dominated delta front deposits, but mainly consists of stacked fluvial-, tide-influenced fluvial- and tidal channels deposited on the lower delta plain (Pontén and Plink-Björklund, 2007). The stacking of these channel-fills shows a proximal- to distal increase in tide-influence due to the fluvial- to tidal transition (Dalrymple and Choi, 2007), but also a vertical increase in fluvial-influence due to delta progradation. Such stacking patterns have also been incorporated in the model described in section 6.2.2.

6.3.3 Other models for tide-dominated deltas

The model for deposition in tide-dominated delta proposed by Dalrymple and Choi (2007) is based on the well-established model tide-dominated estuaries (Dalrymple et al., 1992) and work on the Holocene Fly River Delta (Dalrymple et al., 2003). This plan-form model focuses on the interaction of fluvial and tidal processes on the lower delta plain, and indeed the part of the model for the lower delta plain presented in Figure 6.4 is largely based on and in agreement with this model. However, the model of Dalrymple and Choi (2007) has little detail on processes and facies on the delta front and the facies architecture of the whole succession. The cross-section along depositional strike through the Fly River Delta (Figure 6.4F; Dalrymple et al., 2003) gives some idea on the vertical stacking of facies, but these deposits were formed during a single phase of delta progradation during the Holocene. Therefore their model does not show the facies architecture during prolonged progradation and delta top-set aggradation.
The model for tide-dominated deltas of Willis (2005) does incorporate the evolution of the delta through time in the form of a down-dip cross-section. In this model a tide-dominated, gradationally coarsening upward succession forms during early progradation (“early highstand system tract”). During prolonged progradation (“late highstand system tract” and “lowstand system tract”), the relatively planar, low relief erosion surfaces are interpreted to form due to a combination of forced regression and scouring of tidal currents in tidal channels (Willis and Gabel, 2003; Willis, 2005). During this regression, high-relief incised valleys are interpreted to form which are back-filled by tide-dominated (estuarine) deposits during the subsequent transgression (Willis and Gabel, 2003; Willis, 2005). The use of this sequence stratigraphic terminology implies allogenic forcing. However, Willis and Gabel (2003) propose that this architecture can also form during autogenic lobe-switching.

The model of Willis (2005) has a similar architecture as proposed in this thesis (sections 6.2.1 and 6.2.2; Figure 6.2). In both models a low relief erosion surface is present which cuts down from the top of the coarsening upward succession. Both models contain a narrow, high relief erosion surface which cuts down from higher up in the succession as well. Despite these similarities in geometry which is to a degree due to the fact that both models are partially based on the lower Sego Sandstones, different genetic associations have been interpreted. In the interpretation of Willis (2005), the erosional surfaces are the results of relative sea-level fall, the formation of an incised valley and the back fill of this valley during the subsequent transgression. In contrast, in the model proposed in this thesis, this facies architecture is the results of prolonged progradation during which the more proximal facies erode down in more distal facies during progradation. In this interpretation, the high-relief erosion surface is the results of erosion and infill of a distributary channel. This is in strong contract of the interpretation of a transgressive incised valley-fill.
6.4 Where to expect tide-influenced deltas

The tide-influenced deltaic deposits of the lower Sego Sandstone and the Tofte-Ile formations were both deposited in large scale structurally controlled embayments in an epicontinental seaway. Deposition of the Tofte and Ile formation was rift controlled. Syn-rift estuarine successions deposited in half-grabens with subaerially exposed crest of rotated fault-blocks have been documented from the Suez Rift (Carr et al., 2003; Jackson et al., 2005). However these deposits are transgressive, have highly variable thickness variations and were deposited locally in half grabens. In contrast, the Ile Formation does not vary locally in thickness, and was deposited in a wide basin during a quiescence in rift activity. Better analogues of this setting are the Lower Cretaceous Bentheim Sandstone (Lower Saxony, Germany) (Wonham et al., 1997), the Middle Jurassic Bearreraig Sandstone (Skye, Scotland) (Mellere and Steel, 1996), the Early Jurassic Neill Klinter Group (East Greenland) (Surlyk, 1990; Surlyk, 1991; Alsgaard et al., 2003; Engkilde and Surlyk, 2003; Eide et al., 2016) and the Eocene Dir Abu Lifa Member (Western Desert, Egypt) (Legler et al., 2013). All these examples are progradational and were deposited in wider rift-basins due to which the deposits have a more constant thickness.
In contrast, the Sego Sandstone was deposited in an epicontinental seaway in a foreland basin (Liu and Nummedal, 2004; Liu et al., 2005; Liu et al., 2011), in which the position of the embayment in which the tides were amplified was controlled by the position of thrust faults. Modern foreland basins with an interior seaway do not exist, but analogues can be found in the Cenozoic foreland basins of the Andes. Both tide-dominated estuaries (Hovikovski et al., 2008) and tide-dominated deltas (Maguregui and Tyler, 1991; McIlroy et al., 2005; Schwartz and Graham, 2015; Rossi and Steel, 2016) have been reported from the Andean foreland. However, it is unclear to what extent the embayments in which deposition took place were structurally controlled.
References


Bryant, B., and Nichols, D.J., 1988, Late Mesozoic and early Tertiary reactivation of an ancient crustal boundary along the Uinta trend and its interaction with the Sevier Orogenic belt. in Schmidt, C.J. and Perry, W.P., eds., Interaction of the Rocky Mountain Foreland and the Cordilleran Thrust Belt, Geological Society of America Memoir 171, p.411-430.


Choi, K., 2010, Rhythmic climbing-ripple cross-lamination in inclined heterolithic stratification (IHS) of a macrotidal estuarine channel, Gomso Bay, west coast of Korea: Journal of Sedimentary Research, v. 80, no. 6, p. 550-561.


Doelling, H.H., 2002: Geologic map of the Moab and eastern part of the San Rafael Desert 30' X 60' quadrangles, Grand and Emery counties, Utah, and Mesa county Colorado, 1:100,000.


Green, M.O., and MacDonald, I.T., 2001, Processes driving estuary infilling by marine sands on an embayed coast: Marine Geology, v. 178, no. 1–4, p. 11-37.


Gualtieri, J.L., 2004: Geologic map of the Westwater 30’ x 60’ quadrangle, Grand and Uintah counties, Utah and Garfield and Mesa counties, Colorado, 1:100,000.


Harris, N.B., 1989, Reservoir Geology of Fangst Group (Middle Jurassic), Heidrun Field, Offshore Mid-Norway: AAPG Bulletin, v. 73, no. 11, p. 1415-1435.


Harris, P.T., 1988, Large-scale bedforms as indicators of mutually evasive sand transport and the sequential infilling of wide-mouthed estuaries: Sedimentary Geology, v. 57, no. 3–4, p. 273-298.


Masselink, G., Cointre, L., Williams, J., Gehrels, R., and Blake, W., 2009, Tide-driven dune migration and sediment transport on an intertidal shoal in a shallow estuary in Devon, UK: Marine Geology, v. 262, no. 1–4, p. 82-95.


Muto, T., and Steel, R.J., 2001, Autostepping during the transgressive growth of deltas: Results from flume experiments: Geology, v. 29, no. 9, p. 771-774.

Muto, T., and Steel, R.J., 1997, Principles of regression and transgression; the nature of the interplay between accommodation and sediment supply: Journal of Sedimentary Research, v. 67, no. 6, p. 994-1000.

Muto, T., and Steel, R.J., 1992, Retreat of the front in a prograding delta: Geology, v. 20, no. 11, p. 967-970.


Scholle, P.A., 2003: Geoloc map of New Mexico, 1:500,000.


Stewart, R.H., 2006, Introduction to physical oceanography, Texas A & M University.

275


Tweto, O., 1979: Geologic map of Colorado, 1:500,000.


Witkind, I.J., 2004: Geologic map of the Huntington 30' x 60' quadrangle, Carbon, Emery, Grand, and Uintah counties, Utah, 1:100,000.


Environments Offshore Norway - Palaeozoic to Recent, Norwegian Petroleum Society Special Publication 10, p.233-257.


